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GEOLOGY AND TECTONICS OF PAKISTAN

By

A. H. KAZMI
and
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GRAPHIC PUBLISHERS

ISBN 969-8375-00-7

Published in 1997 by
Graphic Publishers
5C, 6/10, Nazimabad,
Karachi - Pakistan

Printed in Pakistan by
Haqqi Offset Press,
4/5, Commercial Area,
Liaquatabad, Karachi.
Pakistan

Copies of this book may be ordered from:-
GRAPHIC PUBLISHERS
5C, 6/10, Nazimabad,
Karachi - Pakistan.

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Cover – Landsat Photomosaic of southern Pakistan.
Back cover – Landsat Photomosaic of northern Pakistan (Courtesy U.S. Geol. Survey).

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Foreword

The most important development in the Earth Sciences this century has been the plate tectonic paradigm, and the most important end-result of the Wilson Cycle is the continent-continent collisional orogenic belt, the classic example of which is the Himalayas. Although there are some important Precambrian and Palaeozoic rocks in Pakistan, the bulk of the country is dominated by the effects of the India-Asia collision. Pakistan presents for us a wealth of information about the most complicated parts of the collision zone, because it concerns not only the well-known orthogonal ranges of Kohistan and the Karakoram, but also, and most importantly, the lesser known, western transform plate boundary extending from the Makran and the Indian Ocean to Waziristan, as well as the post-collisional tectonic structures in Central Pakistan that extend as far south as the Salt Ranges.

The first publications on the geology of Pakistan appeared about 150 years ago, but research papers have been increasing in abundance exponentially for the last few decades to the present. So it is extremely difficult for the specialist or student today to get a quick grasp of the myriad relationships between the many orogenic belt, sedimentary basins, magmatic and metamorphic rocks, and mineralization. This book is therefore timely, because it brings together for us the many strands of the complex geology of this key region of the world. The great advantage of a volume of this type, in contrast to a paper in a journal, is that space allows a more detailed synthesis of the existing data, and a more comprehensive review of the ideas brought forward to explain them. This book will be extremely valuable for all professional geoscientists involved and/or interested in the geology and tectonics of Pakistan, and also to the many students who need an easy access to the literature on this classic region. Moreover, the book contains an excellent and detailed synthesis of the mineral and hydrocarbon deposits, which cannot be found elsewhere. This is the first grand review of the geology of Pakistan and its value last for many years.

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Preface

The geology of the Indian subcontinent was initially compiled in three monumental volumes, "Manual of the Geology of India" by Medlicot and Blanford (1879–87), "Geology of India" by Wadia (1919), and "Manual of the Geology of India and Burma" by Pascoe (in three volumes, 1950, 1959, 1964). These works presented a summary of the voluminous literature gathered since the establishment of the Geological Survey of India by the British Raj in 1851. Since 1947 when the subcontinent was divided into smaller states, geological research has been pursued more vigorously in these newly independent countries. Pakistan, which was largely unmapped and unexplored at the time of Independence, has now been covered with extensive geological mapping and research. Numerous papers on different aspects of the geology of various regions have been published, but many of these are not readily available. More importantly, the published work is fragmentary and an integrated up-to-date account covering the entire country and the adjacent regions is lacking. To fill this void, we have endeavoured to put together in this volume a summary of the contributions on the geology and tectonics of Pakistan, alongwith a brief sketch of the geology of the surrounding regions. This book is primarily intended for the use of graduate students and researchers. But we hope that it will also provide useful and much needed information to non-specialists, as well as to planners, economists, environmentalists, and investors who may be interested in mineral exploration and mining.

Acknowledgement: Encouragement by Mr. Ghulam Ishaq Khan, former President of Pakistan, and financial support from Infaq Foundation and Shell Pakistan Ltd. made possible the publication of this book. Pakistan Petroleum Ltd., Occidental of Pakistan , Pak-Arab Refinery Ltd., Sui Southern Gas Company Ltd., Lakhra Coal Development Company, and the Pakistan Mineral Development Corporation also made valuable contributions towards printing of the book.

We would like to thank A. A. K. Ghauri, Kabita Goswami, M. Haneef, Farhat Hussain, M. Asif Khan, I. A. K. Jadoon, John F. Shroder, Peter J. Treloar and Peter Warwick for their review of some sections of the book and valuable suggestions. We are grateful to S. Jamiluddin, S. Hamidullah and Iftikhar Ahmad for reading and correcting proofs of some of the chapters. M. P. Searle, P. J. Treloar and R. H. Siddiqui kindly provided some of the photographs. We appreciate the keen interest taken by the Director General, Geological Survey of Pakistan, in the publication of this book. We are also thankful to many authors for permission to use their figures and to the U.S. Geological Survey and the Geological Survey of Pakistan for the Landsat composite imageries used in the book. During the preparation of this book valuable logistic support and technical assistance was provided by Messers Mohammad Ali Mirza, S. Ghazanfar

Abbas, S. H. Gauhar, Mushtaq Hussain, I. H. Qureshi, S. Mehmood Raza, Abbas Ali Shah, A. M. Subhani (from Geological Survey of Pakistan) and Mr. Bashir Ansari, (Lakhra Coal Development Company) during the preparation of this book. Ms. Meher Gul and Malka Saimeen Ozar most patiently typed the text of the book. Our appreciation and gratitude are due to Q. Z. Abid of Graphic Publishers for his cooperation, devotion and assistance in the preparation and printing of this volume.

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Introduction

Geoscientific research during the past two decades has established Pakistan as one of the most fascinating parts of the globe. A collision mountain belt comprising the Himalaya, Karakoram and Hindukush Ranges forms its northern part. Extensive nappes and thrust sheets with Barrovian metamorphism, large granitic batholiths, and sutures marked by melanges, ophiolite and high-P metamorphic rocks characterise these mountain ranges. Some parts of the mountains, rising at a fast rate of over 5 mm/yr, expose granitic plutons only a few million years old. In terms of plate tectonics the opening of the western part of the Indian Ocean and the 4,000 km northward drift of India and its collision with Asia are unique events. Within a small region one can study a major transform plate boundary (the Chaman Fault), subduction-related suture zones (Indus and Shyok sutures), fossil island arcs (Chagai, Kohistan, Ladakh), and a trench-arc system with active plate subduction and deformed plate margins (Makran). Seismotectonically, a large part of Pakistan is very active and several neotectonic features have been mapped.

The exposed rock sequence includes Precambrian metamorphic and plutonic rocks, Paleozoic, Mesozoic and Paleogene pericratonic shelf deposits which form the platform cover and the marginal fold belt, and an extensive and exceptionally thick pile of Neogene molasse (Siwaliks) that fills the foredeep. Famous for their rich and exotic vertebrate fauna, the Siwaliks are the product of intense denudation that accompanied the uplift of the Karakoram and the Himalayas, which may have removed half of the elevated crustal mass in these rapidly rising mountains. Part of this debris was deposited on the ancient Siwalik flood plains, but by far the greater amount was carried to the sea to form the second largest submarine fan in the world, the Indus fan.

Pakistan has been geologically well-known for several decades for its great mountains, extensive glaciers, devastating earthquakes, exotic and prolific Neogene vertebrate fauna, chromite-bearing ophiolites, Precambrian and Paleozoic succession of the Salt Range, the abundant oroclinal flexures and enigmatic syntaxes in its mountain ranges, and the deep gorges and canyons that highlight the antecedent drainage. These geologic features had been largely revealed by the reconnaissance surveys of the early pioneers who explored vast areas despite a lack of proper topographic base maps, inhospitable terrain, hostile tribal conditions and absence of roads and communication system. Geological Survey geologists, based at Calcutta, would travel to Sindh, Balochistan, Punjab, and NWFP on elephants, horses, or on foot (Heron 1953), and on approaching their destination would often find themselves in the midst of skirmishes between the British troops and the local chieftains or the Afghans.

Griesbach (1881) has given an account of one such incidence. After travelling on horse back for several weeks from Calcutta, he arrived at Quetta and was promptly enlisted by the local British garrison commander to fight the Afghans. He spent several days in the trenches near the foot-hills of the Takatu Range, crawling out of the trench every night to collect rock fragments from outside and studying them in day time in the comparative safety of his trench. Braving great risks, many geologists went with the early mountaineering expeditions to the Karakoram and the High Himalaya with relatively poor equipment, and brought back valuable data on geology and glaciology (Godwin-Austen 1864, De Filippi 1910, Dainelli 1922, 1928, 1934, 1939, Desio 1930, 1955). The foundation of the geology of Pakistan was thus laid by such pioneers.

Some of the earliest notices and accounts of local geology are by Barnes (1832) on the Salt mines in the Punjab, Fleming (1843) on the Salt Range and its mineral deposits, Carter (1844) on the hills and soils between Hyderabad and the mouth of the Indus, Carter (1861) on geology of Sind and Balochistan, Mereweather (1852) on an earthquake in upper Sind, Frere (1853) on geology of a part of Sind, Stoliczka (1865) on the Himalayan geology, Medlicott (1868) on Murree-Kotli area and the Salt Range, Wynne (1870-1891) on the geology of parts of Punjab, NWFP (including Kohat and Salt Range), and Kashmir. Most of these and subsequent investigations were of a reconnaissance nature concerning minerals, regional geology and engineering geology. More detailed reports were, however, published on mountaineering expeditions, earthquakes, meteorites and paleontology (Heron 1953).

Medlicott and Blanford (1879-87) published the first "Manual of the Geology of India," and Oldham (1888) compiled the first bibliography which was revised by La Touche (1917). Wadia published his monumental book "Geology of India" in 1919 while Pascoe (1959, 1964) published his comprehensive "Manual of the Geology of India and Burma" in three volumes. These two books sum up the Geology of Pakistan as it was known then. It was largely based on the surveys and reports by Hayden (1915) and Wadia (1932, 1937) for Chitral and Gilgit region; Middlemiss (1896), Davies (1926, 1927, 1930, 1938) and Eames (1951) for NWFP; Gee (1934, 1935, 1937,1940, 1945) for the Salt Range; Blanford (1882), Oldham (1890), Griesbach (1893) and Vredenburg (1901, 1909) for Balochistan and Blanford (1876, 1878, 1879 and 1883) for Sindh. In these early surveys a large volume of invertebrate and vertebrate fossils were also collected and described in several classic papers (see list of references).

Amongst these, Vredenburg's contributions have been the most outstanding and lasting. Without any means of communication, except the camel, and in the face of hostile tribesmen, he covered vast tracts of desert and barren mountain ranges of Chagai, Kalat, Las Bela and Makran. Not only the quality and accuracy of his observations merit appreciation but his monumental and herculean effort in surveying an area of over 155,400 sq km in less than 10 years, under most difficult conditions, are unparalleled in the annals of geological mapping and surveying. He used crude quarter-inch scale hachure maps, yet his mapping has been good and accurate as borne out by the photogeological survey of Balochistan half a century later (HSC 1960). His summarised stratigraphy of Balochistan (1909) stood the test of time surprisingly well with only minor modifications. Vredenburg was also an accomplished paleontologist and has described the invertebrate fauna from Balochistan and Sindh (1904; 1906a,b,c; 1907a,b; 1908a,b; 1923, 1925, 1928a,b).

Pakistan geology turned a new leaf with the dawn of Independence in 1947. Most of the joint assets of trained manpower, equipment, library, technical data, museum etc., were left behind in Calcutta. It began anew from scratch, with a small but active and devoted team of five geologists and two chemists. Having lost most of its traditional sources of industrial minerals located in India and the waters of its three large rivers apportioned to India, Pakistan keenly and actively embarked on a crash programme of mineral exploration including coal and oil, search for and evaluation of new damsites for water and power, and investigation and development of its groundwater resources. During the period 1947-1958, the Geological Survey was expanded from 7 to over 50 geoscientists. Geology departments for teaching and research were established in the universities at Karachi, Hyderabad, Lahore and Peshawar, providing considerable momentum to the geological surveys and research. Large areas with possible mineral prospects were mapped on 1:63,360 (or smaller) scale. Several mineral showings and some workable deposits were discovered (Heron 1954) and a large number of water resources and engineering geology projects were investigated. These led to the construction of Warsak, Mangla, Tarbela and several smaller dams and weirs, and successful implementation of many groundwater development schemes.

Aerial photography of Pakistan on 1:40,000 scale in 1952–53, through the courtesy of the Canadian Government, provided a powerful tool to geologists. This was followed by UNESCO support to the Geology Department of the Punjab University and extensive technical assistance to GSP by USAID/USGS (1956–1966). These organisations were transformed into sophisticated geological research and investigation agencies. Vast areas were mapped, several research papers and reports were published, the first geological map of Pakistan (Bakr and Jackson 1964) was printed, the stratigraphic code of Pakistan was published (Rehman 1962), and the first Bibliography of Pakistan was compiled (Offield 1964). Major advances in geology, stratigraphy and mineral exploration were made as a result of contributions from a large number of Pakistan and foreign geologists (Ahmad 1969, Khan 1985). While the concept of plate tectonics was getting roots as a unifying hypothesis for explaining a variety of geological phenomena, and remote sensing techniques and applications of satellite imagery to geological research were being introduced, Pakistan was in the process of construction of the monumental Karakoram Highway. Passing across the Himalayan and Karakoram ranges, it opened up this hitherto little known region to the outside world. These developments attracted the attention of geoscientists from several countries and paved the way for extensive international collaboration between Pakistani scientists and those from other countries, largely through technical assistance from USAID/USGS, CENTO, US National Science Foundation, UNESCO, IGCP, JICA, CIDA, Australian Research Grants Committee, institute di Geologia University di Milano, Italian National Council of Research, Natural Environment Research Council (UK), Institute of Geological Sciences (UK), National Geographic Society, British Council (London), Smithsonian Institution and various other organisations from France, Germany, Switzerland, New Zealand and other countries. It is beyond the scope of the present work to mention the names of the vast number of geologists from the world over who have in recent years flocked together for geological research in Pakistan. However, the comprehensive list of references at the end of the book will provide some idea of this wide and wonderful international collaboration.

Here we would, in particular, mention the outstanding contribution made by Prof. Ardito Desio, who led several geological and mountaineering expeditions to the Hindukush–Karakoram region between 1952 and 1987, including the main expedition for the ascent of K2 (8,611 m). Work done by him and his associates has laid the foundation of the geology of the Karakoram Mountains, which is so vividly borne out by the extensive current usage of the several geological names assigned by him to the various stratigraphic units, magmatic bodies and tectonic features.

Prior to this book some of the geoscientific work in Pakistan had been summarised in a number of valuable volumes, notably those by Gansser (1964), Shah (1977), Tahirkheli and Jan (1979), Farah and DeJong (1979), Gupta and Delany (1981), Tahirkheli (1982), Shams (1983), Haq and Milliman (1984), La Fort and others (1986), Kazmi and Snee (1988), Malinconico and Lillie (1989), Kazmi and O'Donoghue (1990), Searle (1991), Shroeder (1992), Kadri (1995), and Bender and Raza (1995). However, presently this is the most up-to-date and hopefully the most comprehensive book in which the geology of the entire country has been reviewed along with a summary of the geological setting of the surrounding regions. It contains a fairly elaborate account of the geomorphic features of Pakistan, including the oroclinal hills and majestic mountain ranges, snow capped peaks, surging glaciers, varied land forms, antecedent drainage, the Indus River system, the vast and unique Indus Plain, the Indus Delta, the Cholistan and Thar Deserts and, finally, the coastal and offshore submarine features.

The regional geological setting of Pakistan in relation to her surroundings has been discussed and the plate tectonic scenario, that eventually led to continental collision and produced the various geomorphic and geologic features of the present time, is briefly outlined. This is followed by a detailed description of the tectonic framework of Pakistan and the various tectonic zones.

The book includes an up-to-date review of the stratigraphy of Pakistan, metamorphism and magmatism, and an effort has been made to incorporate results of the latest findings and research particularly in the northern areas. This is followed by an account of the neotectonics of Pakistan. Towards the end an elaborate presentation is given on the mineral and fuel resources of Pakistan.

There are obvious gaps in our knowledge, both region-wise and subject-wise. Some areas such as the Salt Range, Kohat, Hazara, Lower Swat, NE Balochistan and Sindh have been mapped and studied in much greater detail whereas information for places like the High Himalayas, Karakoram, Makran and the tribal areas of the NWFP is yet of a reconnaissance nature. In some fields such as tectonics, petrology and geochemistry of the Himalayas and the Karakoram, there has been a steady flow of published literature in recent years. Other aspects of geology, such as sedimentology, paleontology and paleoecology of the sedimentary basins, have received little attention, except for the recent work on the Neogene in the northern part of the country.

Inevitably, therefore, in our account of the Geology and Tectonics of Pakistan, there is a certain amount of imbalance in the details and descriptions of the geologic features of different regions. It is sincerely hoped that future research would fill these voids and gaps. We conclude with remarks from Macaulay, “Ages are spent in collecting materials, ages more in separating and combining them. Even when a system has been formed, there is still something to add, to alter or to reject.”

2

Geomorphic Features

REGIONAL PHYSIOGRAPHIC SETTING

One of the most conspicuous spots on the physiographic map of Asia is the Pamir knot, also known as the roof of the world. It comprises a 3,660 m high plateau on which several formidable linear mountain ranges, over 5,200 m in altitude, converge from all directions. The more significant of these are the Tien Shan, the Kunlun, the Karakoram, the Hindu Kush and the Alai Ranges (Fig. 2.1). These ranges contain snow-capped lofty peaks such as Muztagata (7,545 m), Mt. Kongpur (7,719 m), Peak Communism (7,495 m) Peak Lenin (7,134 m), K2 (8,611 m) and Tirich Mir (7,690 m). The southern ranges– the Hindu Kush and the Karakoram form the northern part of Pakistan.

Farther southwards, the Himalayan chain forms an extensive oroclinal system of hill ranges stretching from Assam to Balochistan. They comprise a series of more or less parallel ranges, dissected by enormous gorges, and punctuated by a number of mountain basins and plateaus. In Pakistan, the Himalayan Range has a NW-SE trend up to the Jhelum gorge, beyond which its axis bends to form arcs and hairpin syntaxial bends. Starting from Kaghan region and the Jhelum gorge, the orographic features have a general southwesterly or southerly trend up to the Makran coast, where form they swing westward and join the Iranian system of ranges (Fig. 2.2, Photo. 1).

Towards the southeast the mountain ranges are followed by a vast alluvial plain– the Indus Plain, and the formidable Thar Desert which occupies the eastern margin of Pakistan. Farther southwards, there is the Arabian Sea which conceals some interesting morpho-tectonic features. Pakistan is thus comprised of four major discrete geomorphic terraines, namely the mountain ranges, the Indus Plain, the Thar Desert, and the offshore areas. They are as unlike in their physical features as in their geological characteristics and history. A more comprehensive account of these terraines is given in the following pages.

THE MOUNTAIN RANGES

Pakistan's most exotic geomorphic features are indeed the mountains (Photos. 2 to 5). They contain some of the most impressive examples of geomorphic architecture sculptured by wind, glacial, fluvial and mass movement processes. Northern Pakistan is comprised of spectacular mountains capped by some of the highest peaks on Earth (K2 8,611 m, Nanga Parbat 8,126 m, Broad Peak 8,047 m, Gasherbrum I 8,068 m etc.). This region is traversed by some of the longest glaciers (Siachen 72 km, Biofo 62 km, Baltoro 60 km, Hispar 58 km), and deepest gorges (Indus gorge 6,000 + m). Sir Francis Younghusband, one of the earliest explorers of the Karakoram (1887–1889), was so overwhelmed by the sheer size and grandeur of the mountains, the beauty and barrenness of the landscape of the country through which he had travelled that during a Royal Geographical Society meeting (March 1940) he remarked that this region resembled "hundreds of Matterhorns put together" (Miller 1982). The northern regions of Pakistan still contain many

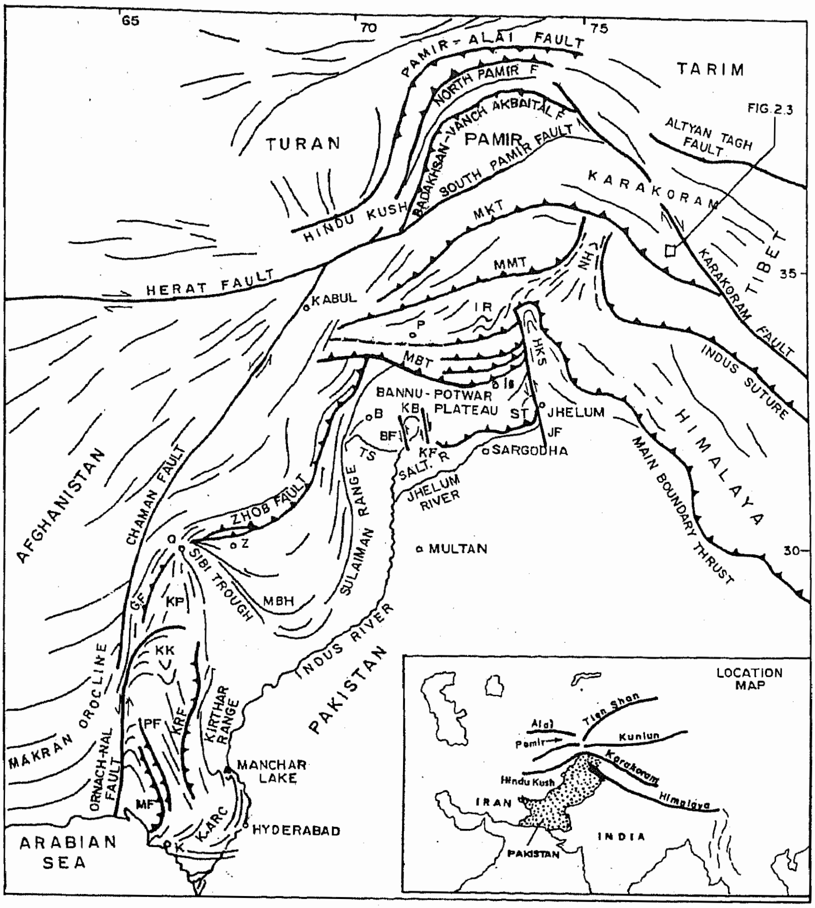


Fig. 2.1. Map showing location and major tectonic trends on and around the northwest part of the Indo-Pakistan subcontinent. B=Bannu, HKS=Hazara-Kashmir Syntaxis, Is=Islamabad, JK=Jacobabad-Khairpur high, K=Karachi, KB=Kalabagh, KK=Khuzdar Knot, KP=Kalat Plateau, MBH=Mari-Bugti Hills, NH=Nanga Parbat-Haramosh Massif, P=Peshawar, TS=Trans-Indus Salt Range, Q=Quetta, Z=Ziarat. ***Faults:*** BF=Bannu, GF=Ghazaband, JF=Jhelum, KF=Kalabagh, KRF=Kirthar, MF=Mor, PF=Pab, MBT=Main Boundary Thrust, MKT=Main Karakoram Thrust, MMT=Main Mantle Thrust, ST=Salt Range Thrust. (Modified from Sarwar and DeJong 1979).

untraversed glaciers, untrodden summits, unconquered peaks, and sizeable tracts of unmapped and unsurveyed areas.

The mountains of Pakistan consist of five discrete orographic zones, namely the Karakoram and the High Himalayas, the Lesser Himalayas and subsidiary ranges, the Sulaiman-Kirthar Mountains, the Makran Oroclinal Ranges and the Chagai – Ras Koh Ranges (Fig. 2.2). These zones are characterised by distinct geological processes and geomorphic features and conform to the broader tectonic zones discussed elsewhere in this book.

The Karakoram and the High Himalayas

The Karakoram and the High Himalayas are comprised of the highest mountain ranges of Pakistan, with average heights in the range of 5,000 m or more. They form an important climatic divide between the monsoonal Indian subcontinent and the deserts of Central Asia. Their summits are covered with perpetual snow and the upper slopes commonly contain cirques and valley glaciers. Considering parallelism and continuity of their orographic axes and geological structure, the mountain ranges in this region may be divided into the following four main groups, namely (a) the Karakoram – Hindu Kush Arc, (b) the Koh-i-Ghizar and Ladakh - Deosai Ranges, (c) the High Himalayas and (d) the Indus Kohistan, Swat and Dir Ranges (Photo. 6). The first three comprise parallel to

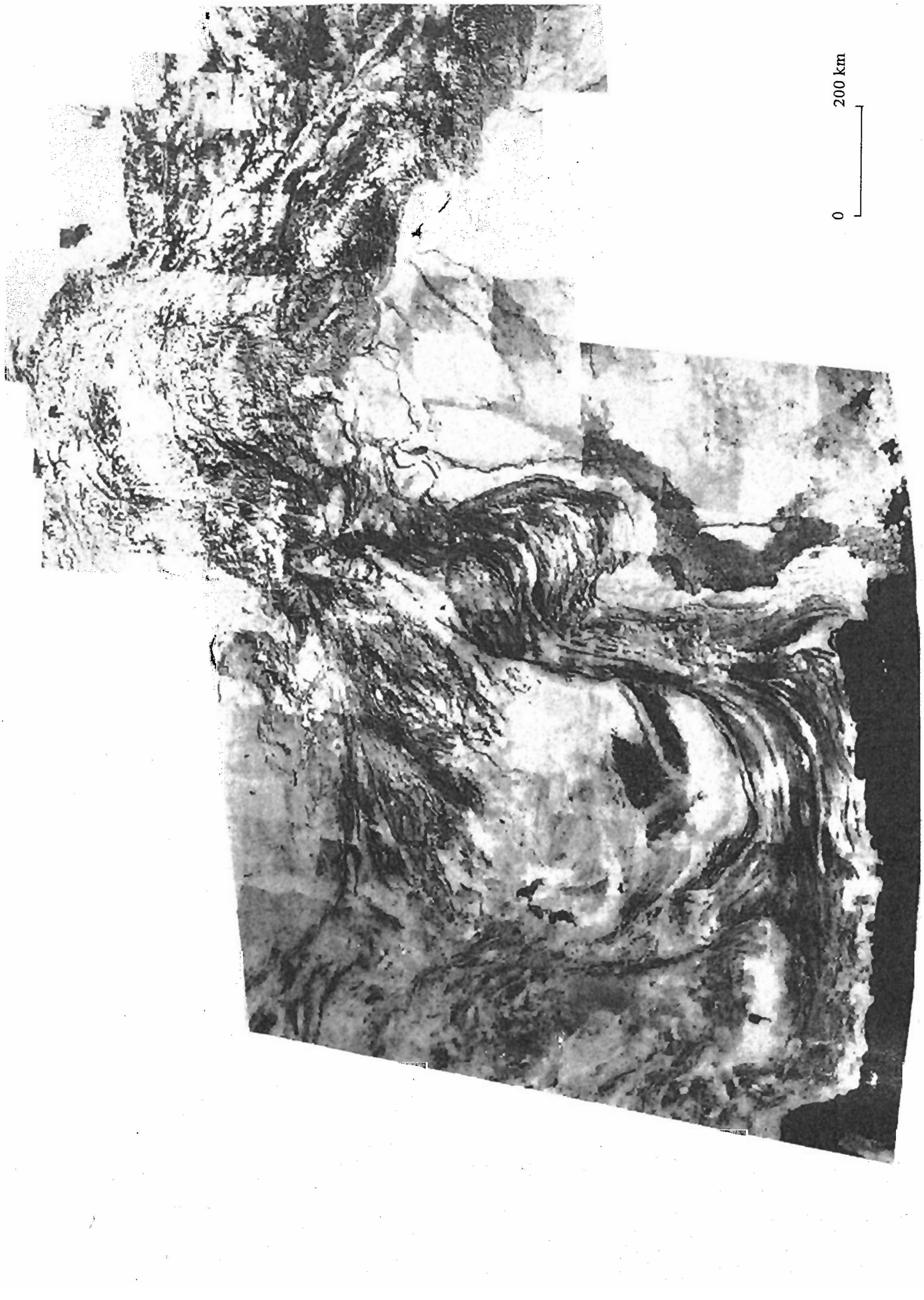


Photo. 1 - Satellite photo mosaic of Pakistan and adjacent regions (courtesy R. D. Lawrence).

Fig. 2.2. Map showing drainage and physical features of Pakistan.

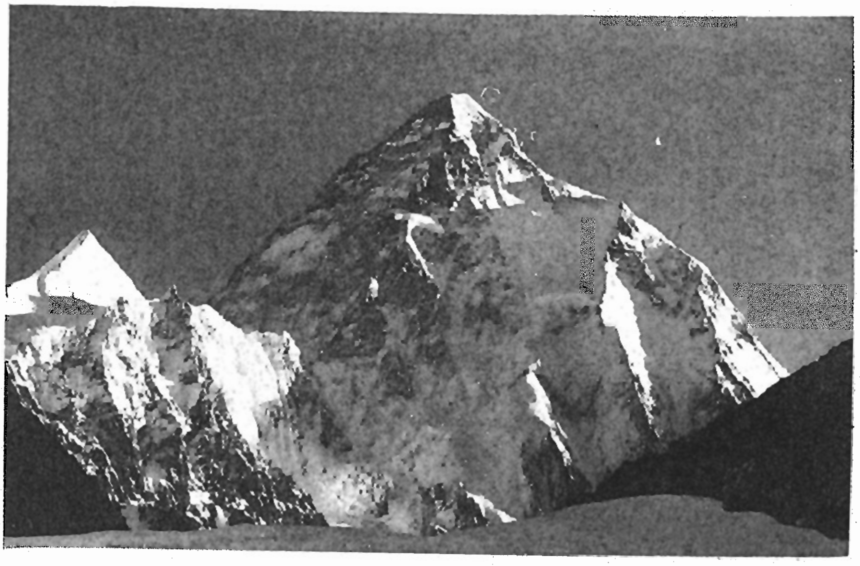


Photo. 2 - Photograph showing the south face of K2 (8610 m). Deep-crustal amphibolite gneisses crop out in this region. (Photo: M. P. Searle).

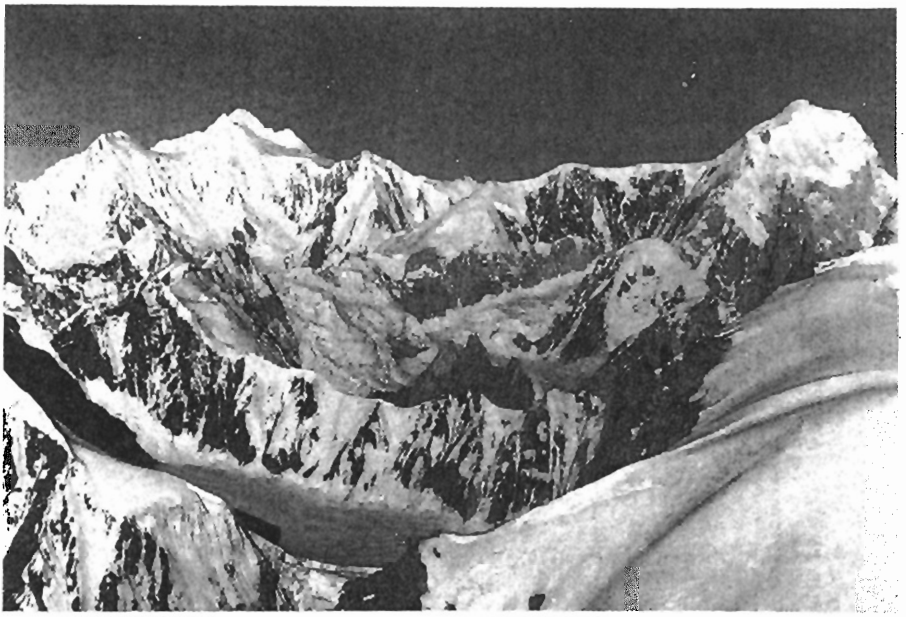


Photo. 3 -Panorama from Makrong Chish. Spantik Peak (7027 m) right background, Chongo Lungma Glacier, central background, Haramosh Peaks (7397 & 6217 m) left background. The MKT runs along the Chongo Lungma Glacier. (Photo: M. P. Searle).

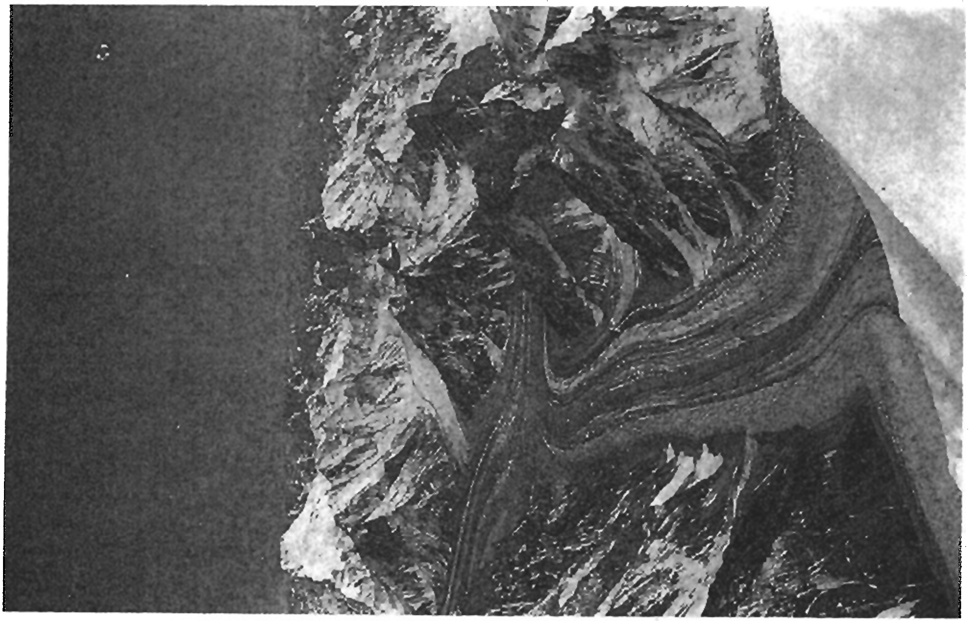


Photo. 4 - View from K2 looking south. In the centre, the Concordia Circus and convergence of the Baltoro Glaciers. (Courtesy Ardito Desio).

Photo. 5- Rakaposhi Peak (7788 m) as seen from Hini. The Minapin Glacier and its snout and moraines are seen in the foreground. (Photo: A. H. Kazmi).

subparallel east-west trending arcuate ranges whereas the last group is comprised of less sharply defined, north-south to northeast-southwest oriented, relatively smaller ranges. In the first three groups the range axes conform to the general geological structure and the strike of the rock formations. The range axes are, however, transverse or oblique to the geological structure in the last group.

The Karakoram–Hindu Kush Arc

In the northern-most region of Pakistan, the Karakoram and Hindu Kush Mountains form an east-west arcuate chain (convex to the north), extending from Ladakh to Chitral and beyond. In Pakistan this mountain belt is nearly 600 km long and it continues beyond for at least 250 km to the southeast and 500 km to the southwest. Moorcroft was the first geographer to apply the name Karakoram (meaning "black gravel", due to its black gravelly slopes) to this great mountain chain (Moorcroft and Trebeck 1841). It appears that the exact extent of the Karakoram and Hindu Kush Ranges has never been precisely demarcated by any of the previous workers and there seems to be considerable confusion. This stems from the fact that the whole tract of mountain ranges between the High Himalayas and the Pamirs, as may be seen on a good physiographic map or the Landsat imageries, appears to be an overlapping mass of parallel to subparallel mountains, which at some places, have been carved out into large, roughly rectangular masses, or the main range has been sculptured to form longitudinal or transverse branches. A number of these branches or offshoots are large enough to be identified as distinct orographic features (Photo. 6).

According to Burrard and Hyden (1908), the Karakoram Range originates from near Rudok and Aling Kangri (7,314 m) in Tibet and extends up to Afghanistan. According to them "the Karakoram and Hindu Kush ranges of mountains are different sections of the same crust fold, the eastern portion of the fold is known as the Karakoram range, the western portion as the Hindu Kush. The range does not change its name at any particular natural feature....". However, for the convenience of geographers they arbitrarily proposed the "water parting between Hunza and Gilgit rivers situated 10 miles east of meridian 74° as the dividing line between the Karakoram and the Hindu Kush". This arbitrary division at this particular location is, of course, not valid.

In his monumental book on the Karakoram, the most comprehensive one on this mountain range, Searle (1991) has shown the Karakoram to extend from Lake Pangong up to Afghanistan. According to him, "the Karakoram Range merges into the Hindu Kush Range and includes many high mountains, notably Tirich Mir (7,690 m) and Noshaq (7,485 m). The western Karakoram forms the main continental divide between the Oxus River drainage and its tributaries flowing through the Pamirs to the north and the Kunar, Yarkhun and Turikho Rivers to the south".

We are of the view that from the stand point of tectonics, geology, structure and orographic continuity, the Karakoram–Hindu Kush Mountains comprise one complex mountain arc and that there is no clear geomorphic or geological feature which divides this otherwise continuous chain of mountain ranges. This conclusion is based on the fact that though the Hunza, Karambar (Ishkuman) and Yarkhun Rivers appear to have cut across the entire width of the Range, there still remains to the north a relatively narrow, yet unbroken and very continuous, more than 700 km long, arcuate rampart, forming the back range of the Karakoram. This back-range extends from Afghan–Chitral border in the west to Kinzil Jilga in the east and beyond it. It does not contain any spectacular feature and its general altitude is about 5,500 to 6,000 m. West to east the Anoshan, Baroghil, Karambar An, Kilik, Mintaka, Aghil and Karakoram Passes cross this back range. It widens out eastward, where it is separated from the main, wider and loftier but relatively broken and discontinuous mass of the front range by the longitudinal Shaksgam Valley. There is a physiographic depression between the front and back ranges marked in patches by the east-west trending components of the Shaksgam–Shimshal, Chapursan–Upper Karambar–Yarkhun Valleys.

The front range of the Karakoram, between Lake Pangong and Hunza, comprises one continuous stretch of nearly 500 km long and over 200 km wide mass of the highest, most glaciated, remotest and the most inaccessible mountain ranges on Earth. Half of its surface area is about 5,000 m above the sea level and much of it is covered by snow and glaciers. It also contains the largest concentration of some of the tallest peaks on Earth (Fig. 2.3).

Westward the continuity of this magnificent mountain is broken by the gorges and valleys of the Hunza, Ishkuman and Yasin Rivers and its stature as well as its snow and glacial cover are gradually reduced. From the upper Ishkuman Valley, a portion of the range branches off westward in the form of the 160 km long Hindu Raj Range. Farther westward, across the Yarkhun Valley another small range, the Yarkhun Range, runs in a northeast-southwest direction parallel to the Hindu Raj Range (Photo. 6). The Yarkhun Range, however, is an offshoot of the Karakoram back range.

The northern and southern margins of the Karakoram–Hindu Kush Mountains have not been adequately defined in the available literature. According to Searle (1991) "the Karakoram lies immediately to the north of two major suture zones– the Shyok and Indus sutures– that mark the closing of Tethys and the collision of the Indian plate". For the sake of simplicity and from the standpoint of tectonics one may perhaps use Searle's definition, but this ignores the fact that the southern margin of the Karakoram is reasonably well defined by the Shyok River Valley, the Indus Valley (between Skardu and Gilgit), and the Ghizar and Gilgit Rivers, and that between the Karakoram and the High Himalayas there are significant orographic features that merit recognition as separate and distinct mountain ranges as discussed in the following paragraphs. The northern confines of the Karakoram–Hindu Kush mountains are not so well defined on physiographic maps or the Landsat imageries. The valleys of the Yarkand, the Dangabash and the Amu (Oxus) Rivers indicate the approximate northward spread of this mountain range (Photo. 6).

Koh-i-Ghizar and Ladakh – Deosai Mountains

Lying in between the Karakoram and the High Himalayas there are three en echelon, relatively narrow and less elevated mountain ranges. The Ladakh Range has a NW-SE orientation and is approximately 270 km long and 25 km wide whereas its altitude varies from 4,000 m to nearly 6,000 m. It is straddled by the Shyok and Indus Valleys and terminates near their confluence east of Skardu.

Westward, and en echelon to the Ladakh Range, and lying south of the Indus are the Deosai Mountains. Southward they are bounded by the High Himalayas. To the west they terminate near the northward trending spur of the Nanga Parbat Range and eastward they extend up to the Kargil Valley, beyond which lies the Zanskar Range. The Deosai

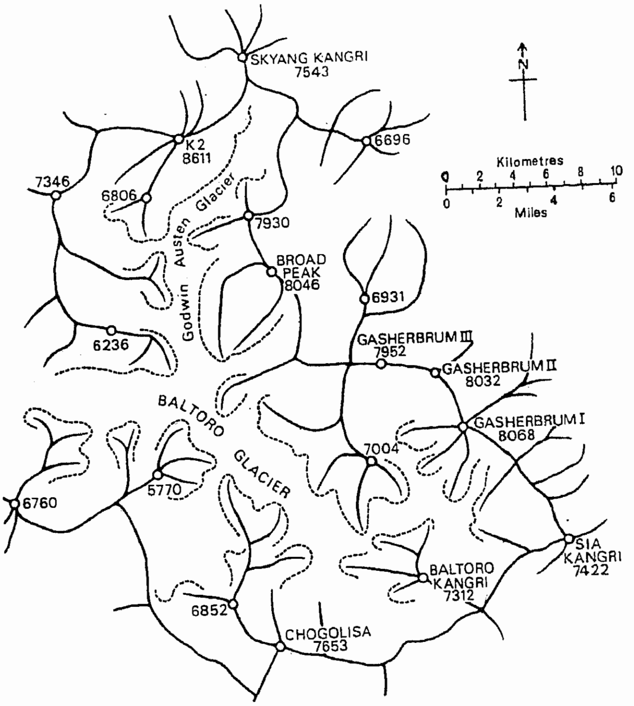
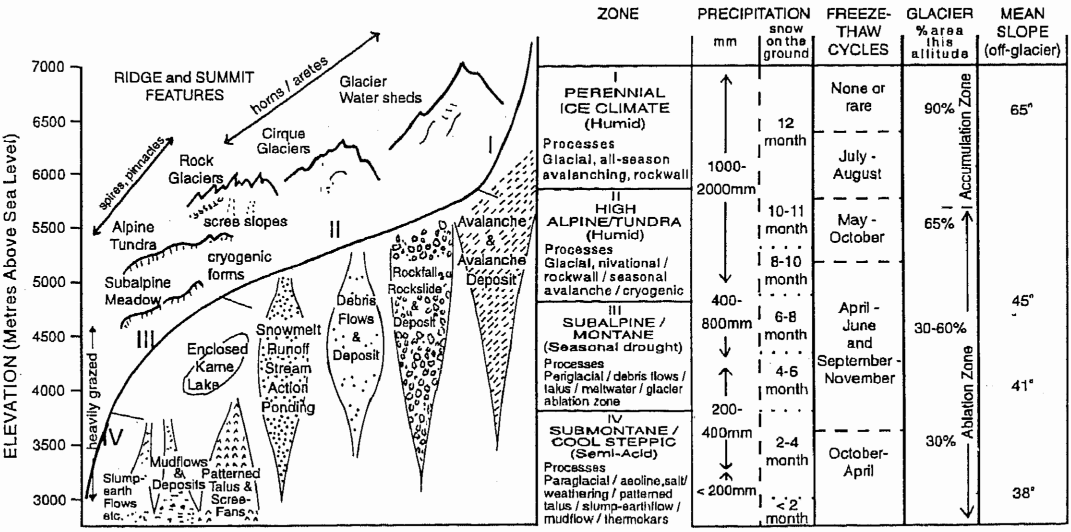


Fig. 2.3. High peaks of the Karakoram. K2 and its satellites comprise the greatest concentration of high peaks in the world. Heights in metres; for location see Fig. 2.1. (From Miller 1982).



ZONE	PRECIPITATION mm	PRECIPITATION snow on the ground	FREEZE-THAW CYCLES	GLACIER % area this altitude	MEAN SLOPE (off-glacier)
I PERENNIAL ICE CLIMATE (Humid) Processes Glacial, all-season avalanching, rockwall	1000-2000mm	12 month	None or rare July - August	90% Accumulation Zone	65°
II HIGH ALPINE/TUNDRA (Humid) Processes Glacial, nivational / rockwall / seasonal avalanche / cryogenic		10-11 month 8-10 month	May - October	65%	
III SUBALPINE / MONTANE (Seasonal drought) Processes Periglacial / debris flows / talus / meltwater / glacier ablation zone	400-800mm	6-8 month 4-6 month	April - June and September - November	30-60% Ablation Zone	45° 41°
IV SUBMONTANE / COOL STEPPIC (Semi-Acid) Processes Paraglacial / aeoline,salt/ weathering / patterned talus / slump-earthflow / mudflow / thermokars	200-400mm < 200mm	2-4 month < 2 month	October-April	30%	38°

Fig. 2.4. Diagram showing the altitudinal distribution of environmental variables in relation to prevalent geomorphic processes in central Karakoram. (From Hewitt 1989).

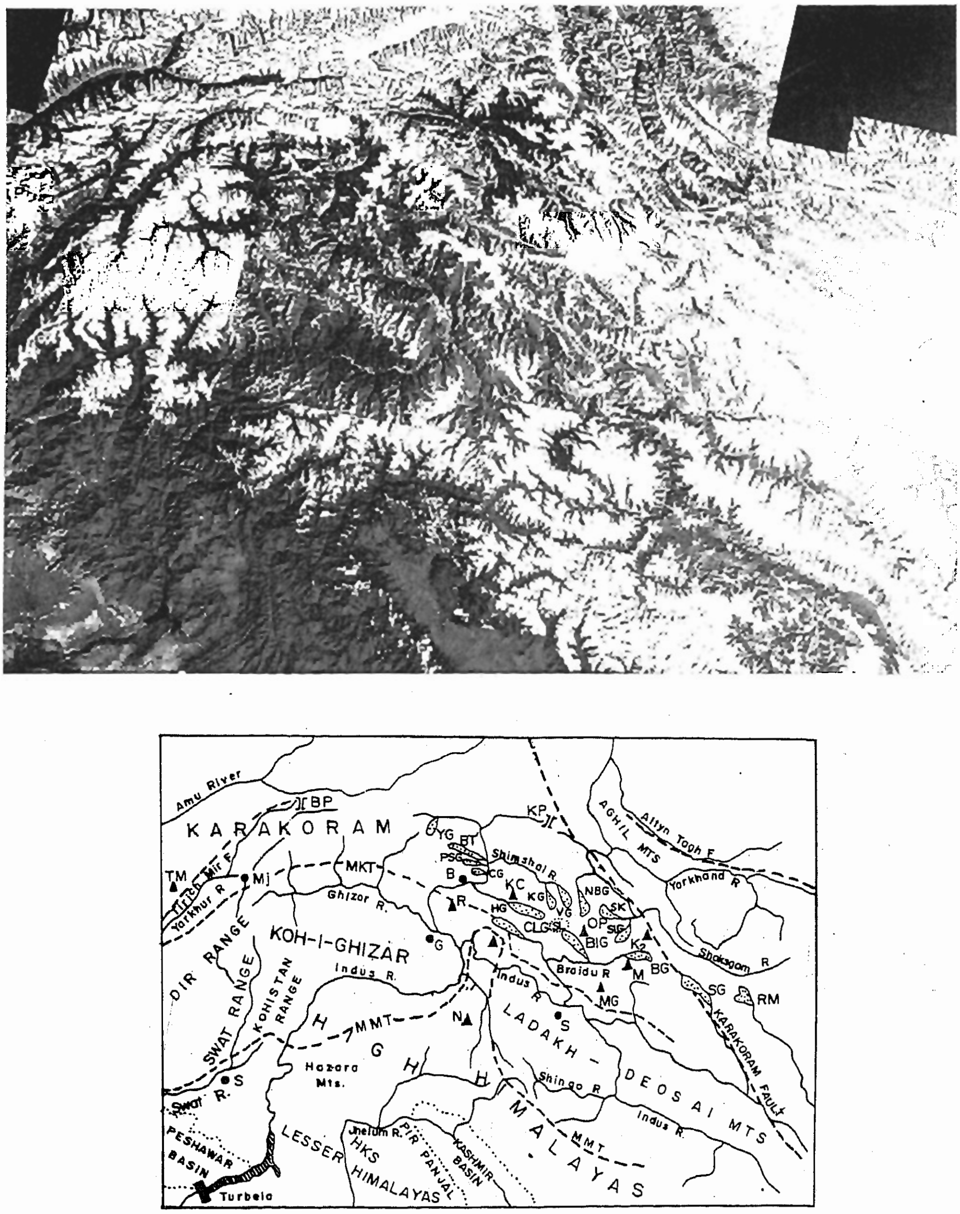


Photo. 6- Landsat photo mosaic of the Karakoram and NW Himalayas. BG–Baltoro Glacier, BP–Baroghil Pass, BIG–Biafo Glacier, BT–Batura Glacier, CLG–Chogo Lungma Glacier, GG–Ghulkin Glacier, Haramosh (7397 m), HG–Hispar Glacier, KC–Khunuang Chhish (7852 m), KG–Khurdopin Glacier, KP–Khunjerab Pass, M–Masherbrum (7821 m), MG–Mango Gusar (6264 m), Mj–Mastuj, N–Nanga Parbat (8125 m), NBG–N. Braldu Glacier, OP–Ogre (7285 m), PSG–Pasu Glacier, R–Rakaposhi (7788 m), RM–Remo Glacier, S–Skardu, SG–Siachin Glacier, SK–Skamri Glacier, SLG–Sarpo Lago Glacier, TM–Tirich Mir (7690 m), VG–Virgerab Glacier, YG–Yashkuk Glacier. See Figures 4.21 and 4.40. (Courtesy U.S. Geol. Survey).

Mountains extend southward for 160 km and are about 60–70 km wide, and their altitude is about 5,000–6,000 m.

West of the Indus loop between Bunji and Laspur (Chitral), there is another mountain range which is 190 km long and nearly 55 km wide. The Ghizar–Gilgit Valleys and the Indus Valley bound it to the north and south respectively. The north-western end of this range is known as Koh-i-Ghizar. However despite its sizeable extent, astonishingly enough the main range has not been given a name as yet. All the geographers, surveyors and explorers visiting this region have been so dazzled and mesmerised by the awesome and spectacular peaks and glaciers of the surrounding Karakoram and High Himalayas that they hardly paid any attention to this range which is completely dwarfed due to its relatively lower altitude of less than 5,000 m. Only its western end attains altitudes ranging from 5,700 to 5,900 m. We propose to name this range as the Koh-i-Ghizar, stretching the name of its loftier western end to its entire extent.

Owing to their lower altitude the Koh-i-Ghizar, Ladakh and Deosai Ranges have only small patches of their summits covered with perpetual snow and they contain very few cirques and corrie glaciers. Most of these, in any case, appear to be remnants of the last glaciation rather than new active ones. These mountain ranges are traversed by narrow transverse valleys and gorges with perennial stream-flow of sparkling, cascading fresh water largely derived from natural springs, unlike the meltwaters of the glacier fed drainage of the Karakoram - Hindu Kush Ranges.

The High Himalayas

The High Himalayas, previously referred to as the "Great Himalayas" (Burrard and Hayden, 1908), form a relatively wide and high mountain range (average crest-line altitude over 6,000 m), extending eastward from the Indus gorge near Besham up to the Brahmaputra gorge in Assam, a distance of about 2,400 km. Its upper reaches are snow-covered and the upper slopes have been dissected by numerous corrie and valley glaciers. The lower slopes are drained by deep transverse defiles and gorges with steep gradients and frequent falls. The western-most part of the High Himalayas within Pakistan is comprised of the Nanga Parbat Range. To the north this range is bounded by the Indus River and to the south by the Kishenganga River.

According to most authors, the Himalayan Range terminates at the Indus gorge because at this point the east-west Himalayan orographic axis terminates against the north-south axis of the Indus Kohistan Range (Photo. 6). All the geological formations here take a sharp hair-pin bend as if they were bent round a pivotal point obstructing them (Wadia 1931).

The Indus Kohistan – Swat – Dir Ranges

West of Nanga Parbat, separated by the valleys of the Indus, Swat, Panjkora, Drosh and Chitral Rivers, there are three north-south trending mountain ranges. They have remained nameless so far and we have herein labelled them (east to west) as Indus Kohistan, Swat and Dir Ranges (Fig. 2.2). The north-south axis of the Kohistan Range joins the east-west axis of the Koh-i-Ghizar in the north. The Kohistan Range is about 160 km long and 50 km wide, and has an altitude of 4,000–6,000 m in the northern part which decreases southward to about 3,000 m in the vicinity of Saidu. This range terminates near Ambela.

The Swat Range is about 120 km long, 40 km wide and has an altitude of about 4,500 m to 5,500 m in the north where it joins the axes of the Ghizar and Dir Ranges, near Paspat. This point is a kind of an orographic knot from where the axes of the Dir, Hindu Raj, Ghizar and Swat Ranges radiate to the northwest, east, southeast and southwest. The altitude of the Swat Range is reduced to about 2,000 m near its southern end, west of Mingora.

The Dir Range extends for about 140 km from near Mastuj to Dir, in a northeast-southwest direction. It is about 30 to 50 km wide and its altitude varies from 4,000 to 5,500 m.

The upper reaches of the Kohistan–Swat Ranges are covered with snow most of the year. They contain relicts and remnants of ancient cirques. The slopes are well-covered with vegetation and pockets of conifer forests. These ranges differ from the other ranges in the Karakoram–High Himalayan region inasmuch as their axes are transverse or oblique to those of the other ranges. They are relatively shorter, have lower altitudes and very little snow cover. Their southern ends taper off to very low altitudes. Their upper reaches are more like the ranges of the Karakoram and High Himalayas while their lower reaches are not unlike the Lesser Himalayas. Nevertheless we have lumped them in one orographic zone with the Karakoram and High Himalayan region because of the linkage of their axes with Hindu Raj and Ghizar Ranges. Their tectonic and geological features also conform to the broader geological aspects of this region.

Vertical relief in the Karakoram and High Himalayas

The rapid uplift of the Karakoram and High Himalayan region during the past 10 Ma (present Nanga Parbat uplift rate > 7 mm/yr, Zeitler 1985) has created stupendous vertical relief. In the main valleys the relief is rarely less than 4,000 m and even the tributaries contain elevational differences of 2,000 m in a horizontal distance of only 1–2 km. In Chitral the Tirich Mir (7,690 m) may be seen from the Yarkhun River, south of Reshun (2,000 m), exposing to the view a relief of over 5,500 m along a 23 km long precipitous slope. In the Hunza Valley, the Rakaposhi (7,788 m) exhibits a relief of about 6,000 m within a distance of only 15 km (Photo. 5) as seen from Hini (1,600 m). The greatest single free-falling precipitous cliff in the world (4,500 m) is the Rupal face of the Nanga Parbat along its southeastern flank as seen from the Astor Valley. The Raikot face of the Nanga Parbat peak (8,126 m) as seen from the ridge north of Gunar (1,000 m) on the Indus bank exposes the maximum vertical relief (about 7,000 m) exposed to view within a short distance (24 km). It is indeed the grandeur of these lofty peaks combined with this awesome relief that makes these mountains so distinct from others. It is impossible to comprehend their stature in one view. One has to pan the gaze, as it were, vertically to encompass the whole view. Speaking about the Nanga Parbat as seen from the Indus, Col. Tanner, one of the early explorers (1904–1905), remarked "it is a scene that is not grasped or taken in at once but after a while the stupendous grandeur of the view is appreciated. It is quite overwhelming in its magnitude; it is in fact one of the grandest spectacles that nature offers to the gaze of man" (Burrard and Hayden 1908).

Mountain Peaks

One of the most significant characteristics of the Karakoram-High Himalayan region is the high density of some of the tallest peaks in the world (Photos. 7,8 and 9). There are more than 44 peaks with an altitude of over 7,300 m, including five of the world's highest 8,000 m peaks– K2, Broad Peak, Gasherbrum I and II, Hidden Peak and

Nanga Parbat (Figs. 2.2 and 2.3, Table 2.1). On a comparative scale these peaks may be slight prominences of the lofty ranges with little or no geological significance, yet they are conspicuous and definite points that can be observed with accuracy from a great distance. For example K2 has been surveyed from distances of 106 to 220 km and Nanga Parbat from 73.6 km to 212 km (Burrad et al. 1907–1908). Since these peaks are lofty, isolated pinnacle of mountains observable from vast distances, they were of immense help in the topographic survey and mapping of this most rugged and inhospitable terrain.

Table 2.1. Major Mountain Peaks of Pakistan (Modified from Shams and Khan 1987).

No.	Name of Peak	Height (metres)	Latitude	Longitude	When conquered
1.	K2 (Godwin Austen)	8611	35° 53'	76° 31'	1954
2.	Nanga Parbat	8126	35° 14'	74° 35'	1953
3.	Gasherbrum I	8068	35° 43'	76° 42'	1958
4.	Broad Peak	8047	35° 48'	76° 34'	1957
5.	Gasherbrum II	8035	35° 46'	76° 39'	1956
6.	Gasherbrum III	7952	35° 46'	76° 39'	1975
7.	Gasherbrum IV	7925	35° 46'	76° 37'	1958
8.	Distaghil Sar	7884	36° 20'	75° 11'	1960
9.	Masherbrum V	7821	35° 39'	76° 19'	1960
10.	North Peak	7809	35° 15'	74° 36'	–
11.	Masherbrum VI	7806	35° 38'	76° 18'	1981
12.	Rakaposhi	7788	36° 09'	74° 31'	1958
13.	Hunza Kunji I	7785	36° 31'	74° 31'	1976
14.	Kanjut Sar	7760	36° 13'	75° 25'	1959
15.	Saltoro Kangri	7742	35° 24'	76° 51'	1962
16.	Tirich Mir	7690	36° 36' 15"	71° 51'	1950
17.	Bride Peak (Chogolisa I)	7654	35° 37'	76° 34'	1958
18.	Hunza Kunji II (Shispare)	7611	36° 27'	74° 41'	1974
19.	Masostang Kangri	7526	35° 19'	77° 38'	–
20.	Rakhiot	7510	35° 15'	74° 37'	1934
21.	Pumari Kish	7492	36° 12'	75° 15'	1979
22.	Noshaq	7484	36° 25'	71° 50'	1960
23.	Teram Kangri I	7464	35° 34'	77° 05'	1975
24.	Malubiting	7458	36° 00'	74° 53'	1971
25.	Sia Kangri I	7422	35° 36'	76° 45'	1934
26.	Haramosh I	7397	35° 50'	74° 54'	1958
27.	Istoro Nal	7389	36° 23'	71° 54'	1955
28.	Momhil Sar	7343	36° 20'	75° 03'	1964
29.	Baintha Brakk	7285	35° 57'	75° 45'	1977
30.	Passu Peak	7284	35° 29'	75° 37'	1978
31.	Diran (Minapin Peak)	7257	36° 07'	74° 40'	1968
32.	Apsarasa I	7245	35° 32'	77° 09'	–
33.	Singhi Kangri (Mt. Rose)	7202	35° 35'	76° 59'	1976
34.	Hachindar	7163	35° 27'	74° 29'	1982
35.	Kampire Dior	7143	36° 37'	74° 19'	1975
36.	Ghehishchish (Spantik)	7027	36° 03'	74° 58'	1978

In the Karakoram and the High Himalayas only a few peaks have names, some have been assigned numbers while most others have no names and are identified only by given heights. According to Burrard and Hayden (1908) "Peaks however posses in their heights an attribute which stars lack, and there is no more useful means of distinguishing peaks than by their heights".

Altitudinal zonation, geomorphic processes and land forms

The spectacular vertical relief of the Karakoram–High Himalayan region has resulted in well-defined altitudinal zones each with its characteristic microclimate, geological processes, geomorphic features, and land forms. Hewitt (1989) has delineated four altitudinal zones in this region (Fig. 2.4).

Zone I: It lies above 5,500 m altitude with 90% area under snow. It contains active glaciers. Erosion and degradation is mainly due to rock fracture from frost action and extreme variation in diurnal temperature. There are frequent avalanches and rock falls. The summits form glacial divides with sharp ridges, horns, aretes, spires and pinnacles.

Zone II: This zone lies between 4,000 m and 5,500 m altitude. It comprises a high, alpine, humid tundra and includes the upper and middle ablation zones. It has a heavy snowfall period of 6–10 months and a short summer. Glacial, nivational and cryogenic processes dominate, with abundant mechanical weathering and rock fragmentation resulting in avalanches, rockfalls and debris flow. Valleys contain scree cones, cirques and valley glaciers, glacial moraines or rock glaciers. Spires and pinnacles are common features.

Zone III: It extends between 3,000 m and 4,000 m altitude and has sub-alpine features with seasonal droughts, cold sub-humid winters, warm summers, rare grassy meadows and clumps of alpine trees. Mechanical rock weathering and periglacial processes are common features with glacial ablation, snow-melt run off, glacial lakes, glacial debris, talus cones, rock falls and land slides. Glaciers are located in this zone and they are covered with thick rock debris. Sharp, narrow and transverse ridges and intervening valleys having wide U-shaped sections characterise this zone.

Zone IV: It covers the area between 1,000 m and 3,000 m altitude and is climatically arid to semi-arid, with hot summers and frequent droughts. It is comprised of mountain valleys and basins which are in a continuous state of degradation resulting from paraglacial, fluvial and aeolian processes. Talus cones, alluvial fans, terraces, glaciofluvial, fluvial and lake bed deposits have formed along the wider stretches of valley flanks. Upper parts of the valleys exhibit U-shaped glaciated valley floor and glaciers descending from tributary valleys often have their snouts close to the confluence with main streams. In the lower reaches this zone contains abundant defiles, gorges and canyons.

In other mountain regions also, extending from the Lesser Himalayas to the Makran and Chagai Ranges, there are distinct altitudinal zones, each with its characteristic ecology, climate, geologic processes, geomorphic features and land forms. However in these mountains each altitudinal zone is much more wide-spread.

Snow cover and glaciers

Crests of the high ranges in the Karakoram-High Himalayan region are largely snow-bound (Photo. 6). The Karakoram has greater ice and snow cover (28% to 37%) than any other mountain system outside the polar region (Wissman 1959). In the Hindu Kush (western Karakoram) and the High Himalayas, ice and snow cover is relatively less extensive, and in the other ranges west of Nanga Parbat only the highest peaks are snow bound (Photo. 6). Snow line is at about 5,200 m along the southern and at 5,800 m along the northern aspects of the High Himalayas. It is at 5,100 m to 5,600 m in southern

Photo. 7 - West face of Gasherbrum I Peak (7925 m). Left of the summit, intrusive contact of diorite into limestones of North Karakoram terrain may be seen.(Photo: M. P. Searle).

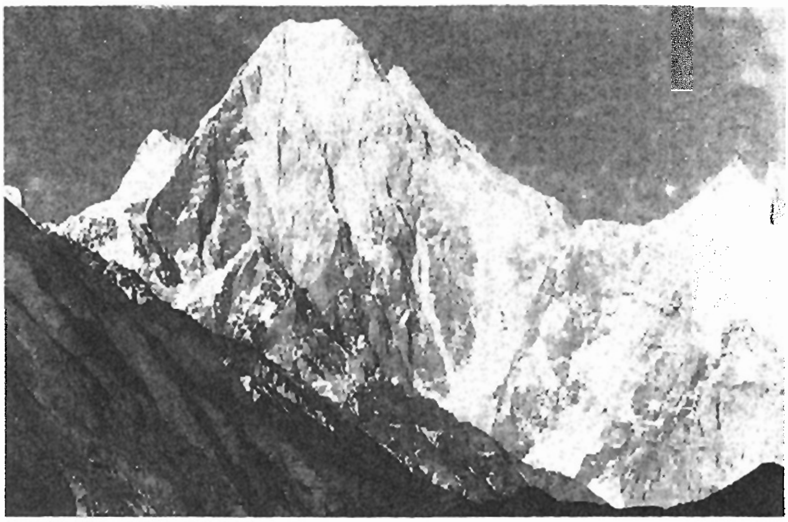


Photo.8 - Rakaposhi Peak (7788 m) seen from Dainyor Nala. Surgin Glacier and its ancient (Late Pleistocene) moraines may be seen in the foreground. (Photo: A. H. Kazmi).

Photo. 9 - North face of Haramosh Peaks (7392 and 6217 m), comprises the northern most outcrop of the Nanga Parbat Group Gneisses. Kupultung and Chongo Lungma Glaciers in foreground (Photo: M. P. Searle).

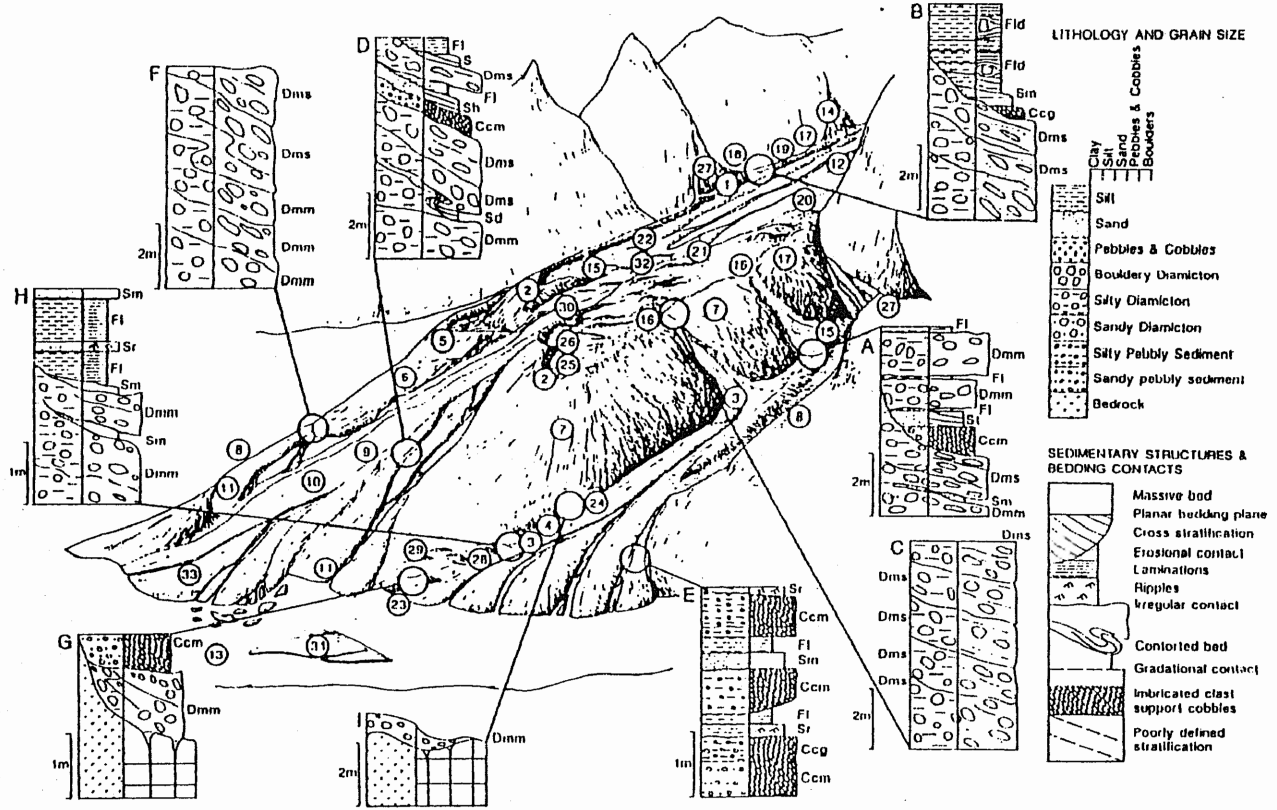


Fig. 2.5. Landform and lithofacies associated with a Ghulkin type glacier. (From Owen and Derbyshire 1992).

1. Truncated scree
2. Lateroterminal drum moraine
3. Outwash channel
4. Glaciofluvial outwash fan
5. Slide moraine
6. Slide-debris flow cones
7. Slide-modified lateral moraine
8. Abandoned lateral outwash fan
9. Meltwater channel
10. Meltwater fan
11. Abandoned meltwater fan
12. Bare ice areas
13. Trunk valley river
14. Debris flow
15. Flow side
16. Gullied lateral moraine
17. Lateral moraine
18. Ablation valley lake
19. Ablation valley
20. Supraglacial lake
21. Supraglacial stream
22. Ice-contact terrace
23. Lateral lodgement till
24. Roche moutonnée
25. Fluted moraine
26. Diffluence col
27. High level till remnant
28. Diffluence col lake
29. Fines washed from supraglacial debris
30. Ice-cored moraines
31. River alluvium
32. Supraglacial debris
33. Dead ice.

Karakoram and 4,700 m to 5,300 m in the northern part of the Karakoram (Mason 1929, Kick 1964). Farther northwards the snow line comes down to 4,700 m to 5,300 m elevation (Shi Yafeng et al. 1980).

In the Karakoram the mountain crests, locally known as Muztaghs (ice mountains), are covered with permanent ice and snow. They commonly enclose snow-fields which feed a number of glaciers (Photo. 10). The glaciers vary widely in size from small cirques lodged in the high mountain recesses to enormous Alpine type valley glaciers rivalling those in the Arctic (Table 2.2). Hanging glaciers are not uncommon. There are transverse glaciers, which are shorter, steeper, and descend to relatively lower levels. In contrast the longitudinal glaciers are larger and rarely descend to below 3,000 m altitude (Mason 1930). A characteristic feature of Pakistani glaciers is the extensive cover of superficial moraine and rock-waste, particularly near their snouts. Only a few glaciers are of the basin reservoir type whereas the incised reservoir and avalanche types with narrow, high gradient accumulation are more common (Washburn 1939).

Table 2.2. Some glaciers of the Karakoram-High Himalayan Region.

MOUNTAIN	GLACIER	LENGTH (km)	RIVER DRAINAGE
Himalaya	Rupal (S)	17.6	Astor
	Rupal (N)	16	Indus
	Phungatori	15.5	Indus
	Rakhiot	13	Indus
	Shafut	13	Indus
	Barmal	13	Wardwan
	Diamir	11.2	Indus
Karakoram	Siachen	72	Shyok
	Biafo	62.5	Braldo
	Hispar	61	Hunza
	Batura	59	Hunza
	Baltoro	58	Braldo
	Gasherbrum	38.5	Shyok
	Chogolungma	38.5	Shigar
	Malungatti	35	Hunza
	Ghainri	33.6	Nubra
	Barpu	25	Hunza
	Pasu	24	Hunza
	Remo	24	Shyok
	Daintar	21	Hunza
	Ghulkin	18	Hunza
	Rakaposhi	17.6	Hunza
	Hasanabad	17	Hunza
	Ghutalgi Yas	15	Hunza
	Pisan	11	Hunza
	Ghulmet	7	Hunza
Hindu Kush	Sakiz Jarab	30.4	Kunar
	Tirich Mir	22.4	Kunar
	Rich	16	Kunar
	Wasmu	11	Kunar

The glaciers in this region are high activity glaciers and have some of the steepest gradients in the world. For example, the avalanche fed Minapin Glacier (Photo. 5) beneath the Rakaposhi Peak (7,788 m) descends from about 5,300 m to 2,400 m over a distance of only 10 km (29% slope). The glaciers thus move through a wide range of climatic zones and are glaciologically complex. High summer radiation and steep barren slopes control the glacier ablation patterns. The maximum radiation balance measured on Batura Glacier was over 27.9 MWm^{-2} (August 1947, Zhang Jinhua et al. 1980). It is estimated that melting accounts for 80% of the heat loss whereas only 20% is due to evaporation and convection (Goudie et al. 1984).

The lower reaches of many glaciers are located in relatively hot mountain valleys and are thus provided with plentiful meltwater causing frequent surges. Glaciers north of K2 and in the Shimshal, Hispar and Braldu Valleys are known for their surge behaviour and are characterised by strongly convoluted moraines. Rock glaciers, which are comprised of slow moving ice, rock fragments and water under pressure, are common in the Himalayas, Karakoram and Hindu Kush (Giardino et al. 1987, Shroder 1992). The Siachen Glacier on the eastern slopes of Nanga Parbat is a good example. The upper part of the glacier is at about 3,500 m to 4,000 m altitude and a 2,000 m high cliff separates the glacier from its source of ice and snow perched above at about 6,000 m. It is however regularly fed by ice and snow avalanches and rock falls. Near its lower part the rock glacier flows through its lateral moraines into deep ablation valleys, nearly perpendicular to the main flow direction of the ice glacier. Another good example of a rock glacier, noted by the senior author, is in a hanging glacial valley perched above the Hunza River gorge southwest of Chalt.

According to their movement patterns, Mercer (1975) has grouped the Karakoram and Himalayan glaciers into three categories– (a) glaciers with steady movement (these are also the longer ones), (b) glaciers having cyclic advances (these have short steep crevasses) and (c) surging glaciers characterised by catastrophic advances. It has been observed that in this region the advancing glaciers flow most commonly towards east, southeast, northeast and north and rarely towards south and west.

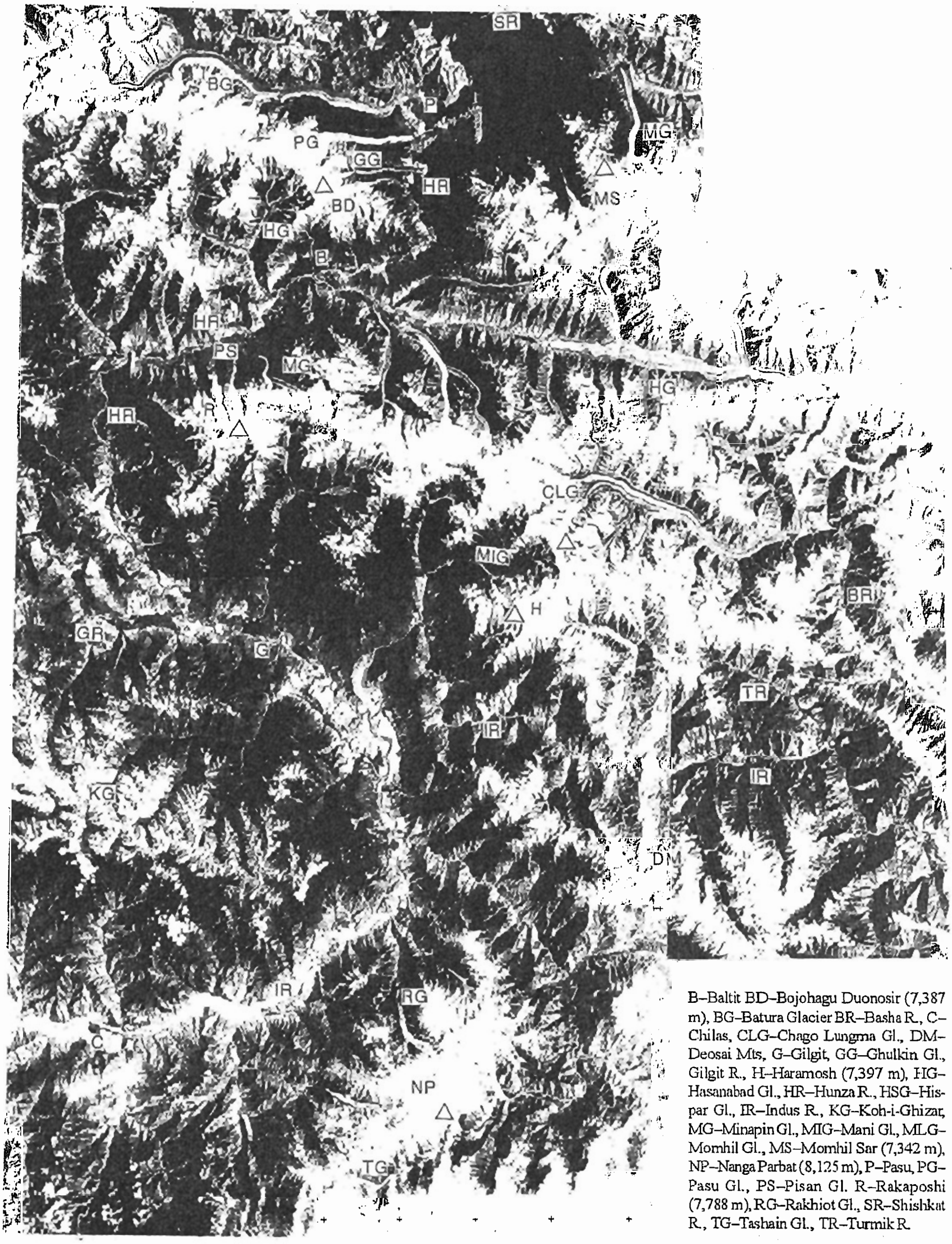
Based on sediment–landform association, Owen and Derbyshire (1989) have classified the Karakoram glaciers into two main types :–

1. *Ghulkin type:* In this type the glaciofluvial debris flow and debris slide processes dominate ice-contact fan sedimentation; ice contact cone is highly dissected and consists of meltout till (Fig. 2.5).

2. *Pasu type:* It comprises hummocky moraines, glaciofluvial outwash plains, small dissected end moraines, lateral moraines and subglacial tills (Fig. 2.6).

Glacial ice thickness: Ice thickness measurements using radio echo-sounding techniques reveal that the thickness of ice varies from 429 m to 590 m in the Hispar Glacier (Dong et al. 1984) and 111 m to 428 m in the Ghulkin Glacier (Francis et al. 1984). Gravimetric determination of ice thickness of Batura Glacier in Hunza indicates an average ice thickness of 310 m with a maximum of 432 m (Shi Yafeng et al. 1984). The depth of ice in the Dachingan, Zemu and Fedchenko Glaciers is stated to be 600, 200 and 548 metres respectively. Near the snout of the Baltoro Glacier the ice is about 122 m thick (Wadia 1957).

Glacial flow rates: Due to great thickness of ice the deeper parts of the glaciers are at or close to 0°C and they behave like temperate glaciers. Owing to relatively high activity indices, these glaciers have a relatively high flow rates ranging from 100 m to 1,000 m/yr (Goudie et al. 1984). Velocities of some of the Hunza glaciers are shown in Table 2.3.



B–Baltit BD–Bojohagu Duonosir (7,387 m), BG–Batura Glacier BR–Basha R., C–Chilas, CLG–Chago Lungma Gl., DM–Deosai Mts, G–Gilgit, GG–Ghulkin Gl., Gilgit R., H–Haramosh (7,397 m), HG–Hasanabad Gl., HR–Hunza R., HSG–Hispar Gl., IR–Indus R., KG–Koh-i-Ghizar, MG–Minapin Gl., MIG–Mani Gl., MLG–Momhil Gl., MS–Momhil Sar (7,342 m), NP–Nanga Parbat (8,125 m), P–Pasu, PG–Pasu Gl., PS–Pisan Gl. R–Rakaposhi (7,788 m), RG–Rakhiot Gl., SR–Shishkat R., TG–Tashain Gl., TR–Turmik R.

Photo. 10- Satellite imagery showing the Nanga Parbat-Haramosh Massif, and parts of the Karakoram and Kohistan-Ladakh terrains. See Figures 4.21 and 4.40. (Courtesy U.S. Geol. Survey).

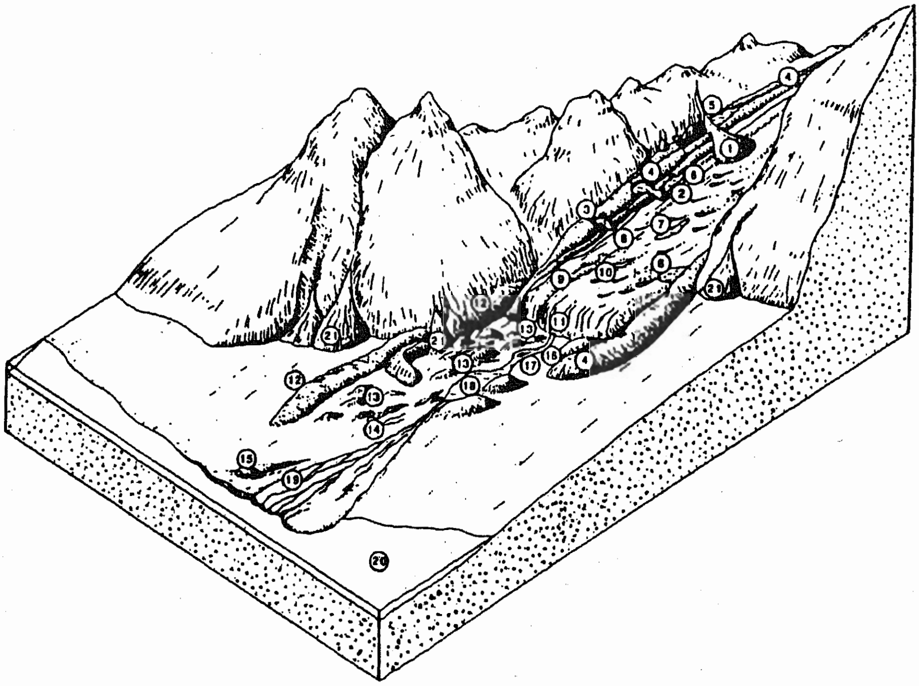


Fig. 2.6. Main landforms associated with a Pasu type glacier (From Owen and Derbyshire 1992).

1. Rock avalanche/flowslide deposit moraine.
2. Land sliding from a lateral moraine.
3. Ablation valley.
4. Lateral moraine.
5. Steep rocky valley sides.
6. Supraglacial marginal pond.
7. Supraglacial pond.
8. Supraglacial meltwater stream.
9. Surface of glaciers covered in supraglacial debris.
10. Crevasses and shears exposing fresh glacial ice
11. Subglacial meltwater tunnel.
12. Ancient lateral moraines.
13. Hammocky moraines
14. Fluted moraines
15. Roche moutonnée
16. Glacially eroded bedrocks with small deposits of subglacial tills
17. Meltwater stream
18. Periglacial lake
19. Sandur
20. Main valley river
21. Scree slope and talus cones

Table 2.3. Velocities of selected Karakoram glaciers (after Wadia 1957, Gaudie et al. 1984).

Glacier	Velocity (m/yr)	Glacier	Velocity (m/yr)
Baltoro	299.612	Hasanabad	329-449
Batura	1000	Kutiah	80-216
Biafo	18.7	Minapin	350-645
Chogolungma	296-366	Pasu	157
Fedchenko	169	Zemu	84
Ghulkin	117		

Glacial Fluctuations and Surges: Historical record of glacier fluctuations in the Himalayas and the Karakoram indicates that in the late nineteenth and early twentieth centuries the glaciers were generally advancing, followed by predominant retreat during 1910–1960 (Mason 1930, 1935, Goudie 1984). Fluctuation in the Batura, Ghulkin–Minapin and Hasanabad Glaciers in Hunza have been studied and changes in the position of the snouts of these glaciers since 1880 have been shown in Fig. 2.7.

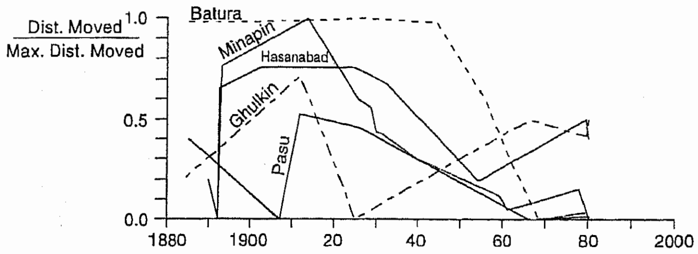


Fig. 2.7. Composite graph of changes in the position of the Hunza glaciers since the 1880's (From Goudie et al. 1984).

The Hasanabad Glacier has advanced and retreated more rapidly than almost any other glacier on Earth (except the Minapin). Around 1890 this glacier rapidly advanced between 9.6 to 11.2 km in one season, followed by a static snout position from 1903 to 1925. Between 1925 and 1980, repeated and rapid oscillations in the position of its snout have been noted as shown in Figure 2.8. The Pumari Chhish Glacier, on the northern side of Hispar, dramatically surged forward 1.5 km during 1989, fanning across the complete width of the Hispar Glacier (Searle 1991).

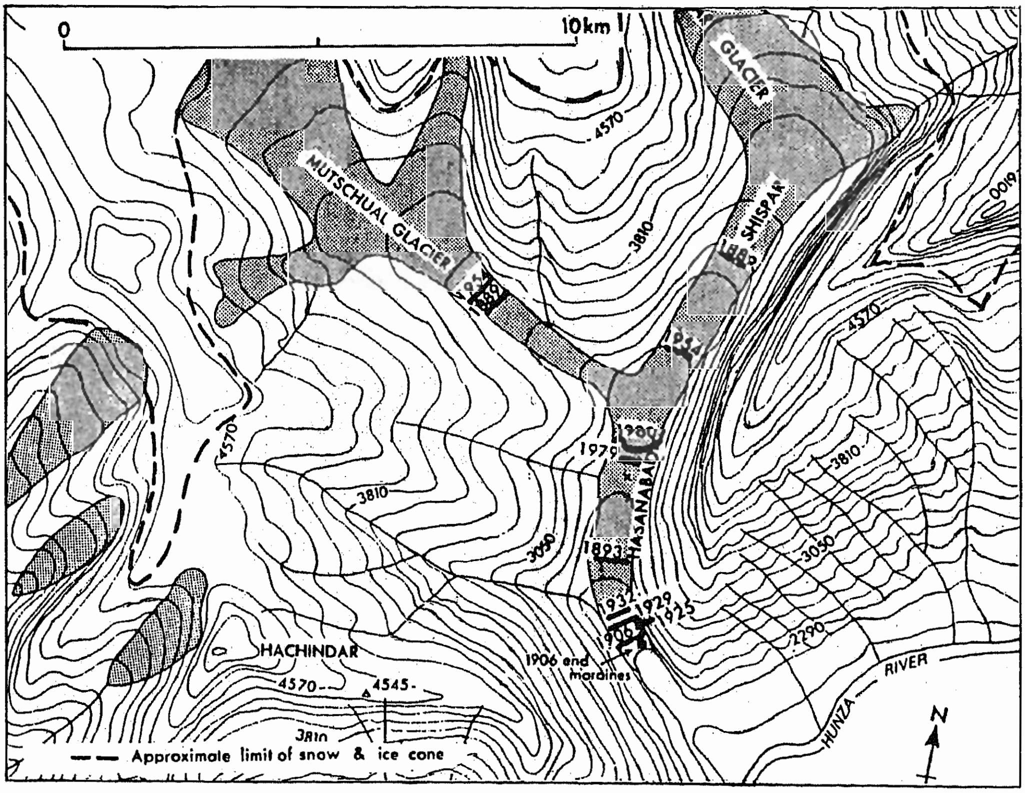


Fig. 2.8. Variations in the position of the snout of the Hasanabad Glacier, 1889–1980. (From Goudie et al. 1984).

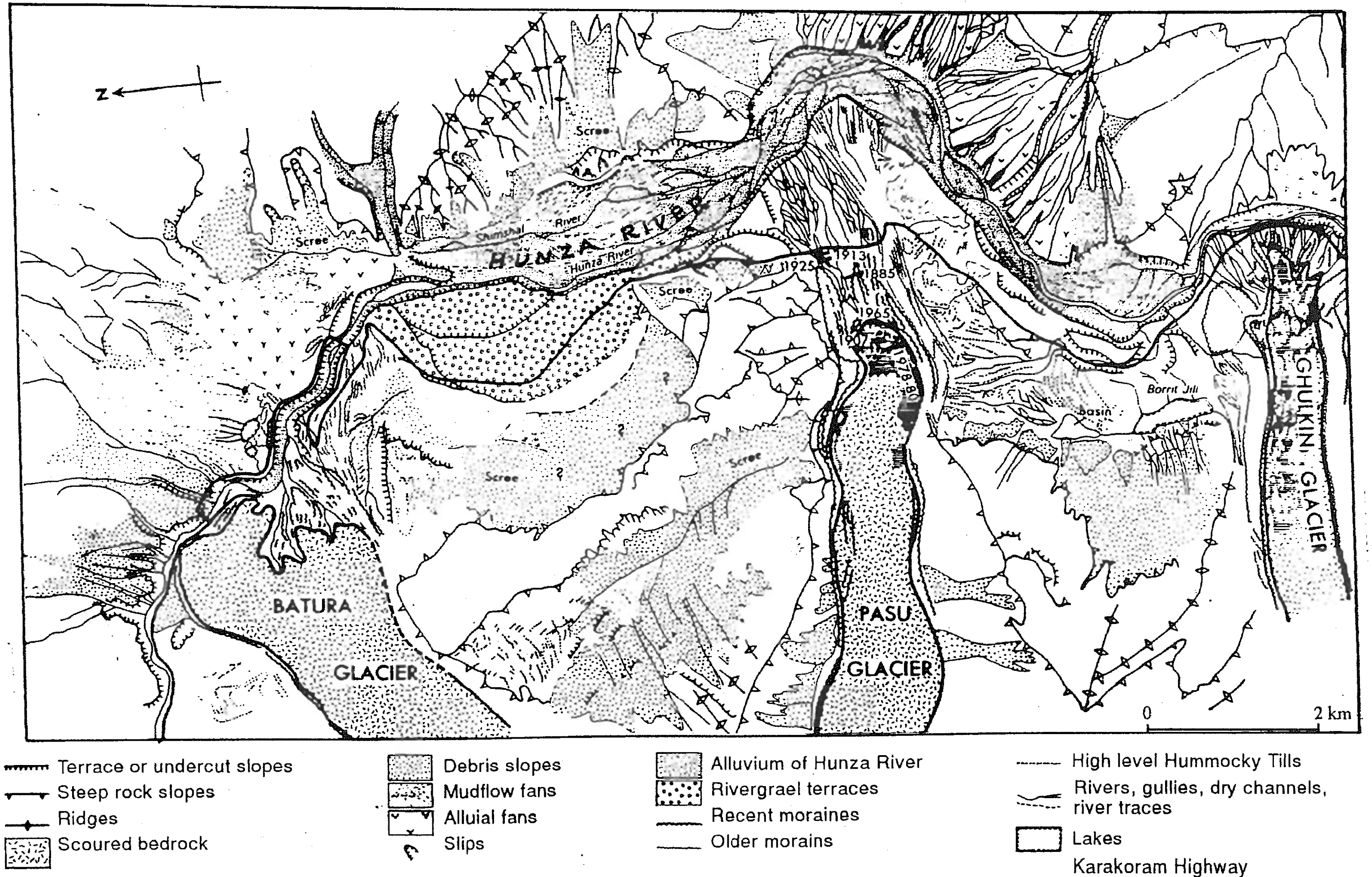


Fig. 2.9. Geomorphological map of part of the upper Hunza Valley near Pasu showing distribution, extent and form of mud flow fans and glacial deposits.(From Goudie et al. 1984).

At a number of localities, remnants of glacial deposits occur right across the main river valleys, indicating rapid glacial advances from tributary valleys, and consequent damming of the main valleys. The best known incidences of river blockade due to glacial surges are those of Pasu, Shishkat and Hini in the Hunza Valley (Derbyshire et al. 1990, Goudie et al. 1984).

Glacial erosion and deposition: The glaciers in Pakistan are also characterised by some of the highest glacial erosion rates on Earth owing to their high sliding velocities, abundant englacial and subglacial meltwater, and steep longitudinal gradients (Goudie et al. 1984). In the upper Hunza Valley the annual denudation is estimated at 5,000 to 10,000 tonnes/sq km or as high as 3 m every 1,000 years (Miller 1982).

The high activity, high gradient, low latitude valley glaciers of the Karakoram and the Himalayas have made deep erosional incisions. They have rapidly modified deglaciated surfaces and have formed well developed moutonnées, fluted tills, extensive lateral and end moraines. The glacial sediments comprise a gradual series of subglacial tills of lodgement type to subglacial and supraglacial meltout tills, angular ice-contact glaciofluvial gravels and proximal and distal glacio-lacustrine silts (Fig. 2.9). The glacial sediments exhibit complex interbedding and facies variations over short distances (Fig. 2.10).

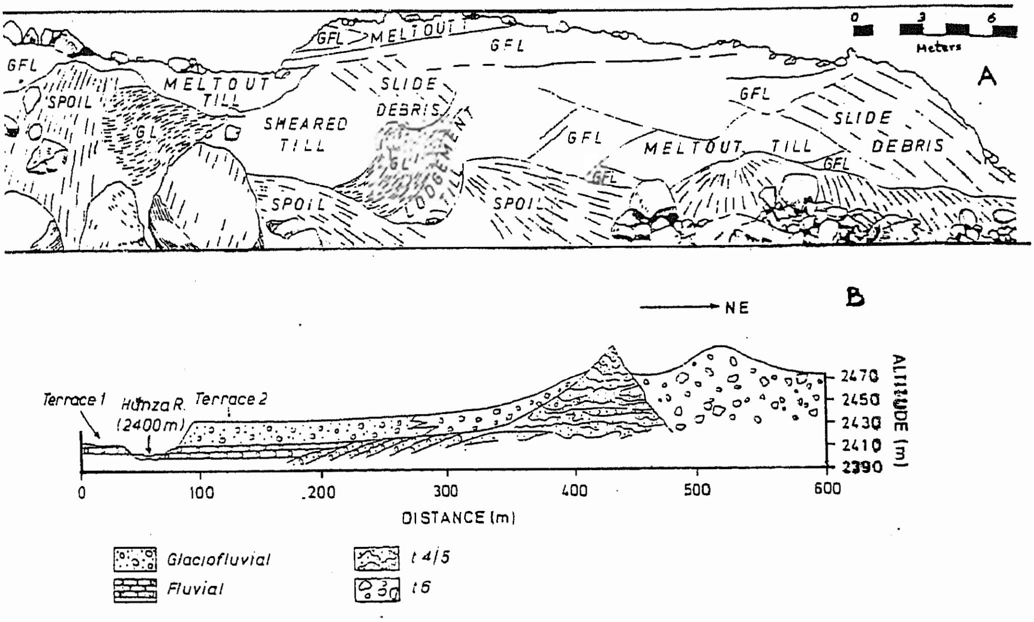


Fig. 2.10. Sections of glacial and glaciofluvial deposits of Batura Glacier. A. Exposure at Batura Glacier terminus; B-Section showing transition from Postglacial Batura (t6) stage till (right) to glaciofluvial terrace. (From Jijun at al. 1984).

Main streams, valleys and gorges

Main Streams: The Indus and its tributaries form the main drainage in the Karakoram-High Himalayan region. East to west, its main tributaries are Shyok, Shigar, Hunza, Astor, Gilgit, Ishkuman, Yasin, Ghizar, Yarkhun, Rick Gol, Arkari, Kunar, Panjkora and Swat Rivers (Fig. 2.2). They have steep gradients with frequent cataracts and flow through deep defiles and narrow valleys. A characteristic feature of the main valleys is the repeated alternation of their wider sections with intervening chasms and defiles. The valleys are commonly wide where they are longitudinal or parallel to the structural grain and where the rocks are relatively soft. They are narrower and form gorges where they traverse hard rocks and make abrupt sharp bends across the structure and the orographic axes.

Streams draining the northern slopes initially flow along longitudinal valleys or structural troughs, then turn abruptly and cut across the mountain range to join another major stream. They thus have a placid meandering or braided flow in the wider sections of the valleys followed by sudden torrential plunge through narrow and steep gorges. This pattern continues until the Indus emerges from its hilly sojourn.

Gorges: Some of the deepest gorges and canyons occur in this region. The most spectacular is the 5,000 m or more deep Indus gorge near Sassi, west of the Skardu basin. This gorge (base altitude 2,200 m) cuts through the Nanga Parbat–Haramosh Massif. Farther westward in the Patan–Dasu region, the Indus forms another 6,500 m deep gorge. This is indeed the deepest canyon in the world. Similar, but not quite as deep, are the Hunza River gorges upstream of Baltit and near Chalt, Gilgit River gorge near Henzal, and several gorges in the Ishkuman, Yasin, Mastuj and Chitral Valleys. These deep gorges and the sharp relief are due to the extraordinarily high rates of denudation and uplift of this region (Table 2.4).

Table 2.4. Exhumation, denudation and uplift rates for selected areas of the High Himalayas and Karakoram (from Owen et al. 1992).

	Method of Determination	**Region**
Exhumation rate (mm yr^{-1})		
0.1-1.5	Fission track	Nanga Parbat
0.7-0.8	Mineral cooling temperature	High Himalayas, India
1.4-2.1	Kyanite grade metamorphism	Suru valley, western Zanskar
1.2-1.6	Mineral cooling ages	Baltoro area, High Himalaya
Denudation rates (mm yr^{-1})		
2.0 Growth of Bengal Fan	Ganges/Brahmaputra watershed	
1.8 Sediment yields	Hunza watershed	
2.56-5.14	Sediment yields	R. Tamur watershed
1.43-2.50		R. Sun Kosi watershed
<0.65	Sediment yields	Annapurna range
Uplift (mm yr-1)		
5.0-12.0	Flexural modelling of the lithosphere	Northern Pakistan
0.1-1.5	Levelling of river terraces	Nepal

Valley slopes: Valleys in the Karakoram–High Himalayan region are faced with widespread slope failures and frequent mass movement activity. The valley slopes range from snow and ice-covered avalanche slopes of high peaks and ridges in the higher regions (Photo. 3), to low angled alluvial fans and terraces in the lower parts of the valleys (Photo.

5). The valley slopes of this region have been classified as follows (Goudie et al. 1984):–

a) Snow- and ice-covered steep avalanche slopes of the high peaks and ridges.
b) Seasonally snow-covered rock slopes, 40° – 90° steep, subject to frost and chemical weathering, with frequent rock-falls and avalanches.
c) High altitude shoulders (2,600–3,600 m), which are commonly covered with pastures and Juniper forests.
d) Scree slopes, ranging from small cones to huge compound debris accumulations with slopes 1,000 m long and a relief of up to 500–600 m. Overlapping scree cones often extend 3–10 km along valley sides. Slopes range 25°–35°, at places increasing to 40°. The scree slopes are the most conspicuous land-form in the valleys of this region (Fig. 2.11).
e) Mudflow debris cones and fans. Surface slopes range 7°–20°.
f) Alluvial fans with low angles (2°–7°).
g) Glacial deposits and valley fills.
h) Terraces (fluvial, glaciofluvial or lacustrine).

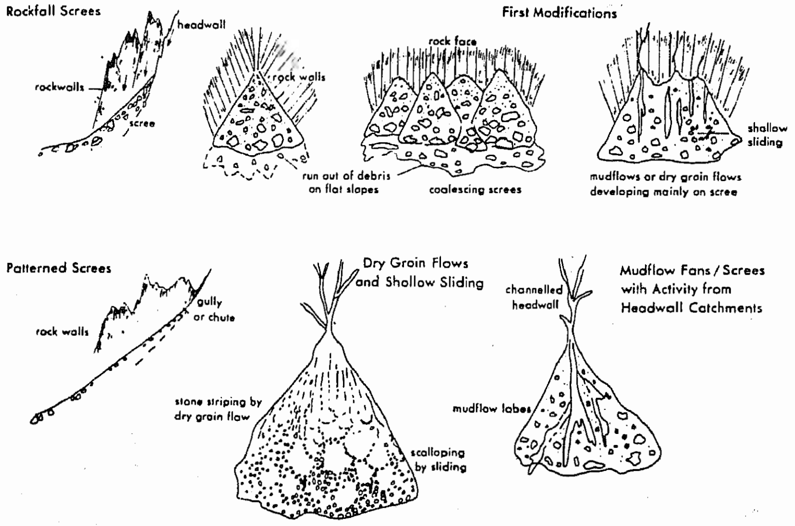


Fig. 2.11. Types of scree slopes commonly seen in the valleys of the Karakoram Mountains (From Brunsden et al. 1984).

Land forms: In their upper reaches the main valleys are characterised by glacial landforms such as ablation valleys, extensive lateral, medial and terminal moraines, ice contact fans and outwash fans (Fig. 2.9). Farther down in the valleys, glaciofluvial sediments on ice-contact fans, outwash plains of low sinuosity, braided channels, and fluvial sediments characterised by sand sheets and imbricated cobbles are common. Lacustrine sediments in ice dammed lakes or in lakes formed by massive landslides occur at many localities (Fig. 2.12). In some of the larger basins such as Skardu, extensive sand dunes have formed. The most common and the most spectacular features of these valleys, nevertheless, are the mass movement deposits – rockfalls, rockslides, debris flows and avalanches (Fig. 2.13).

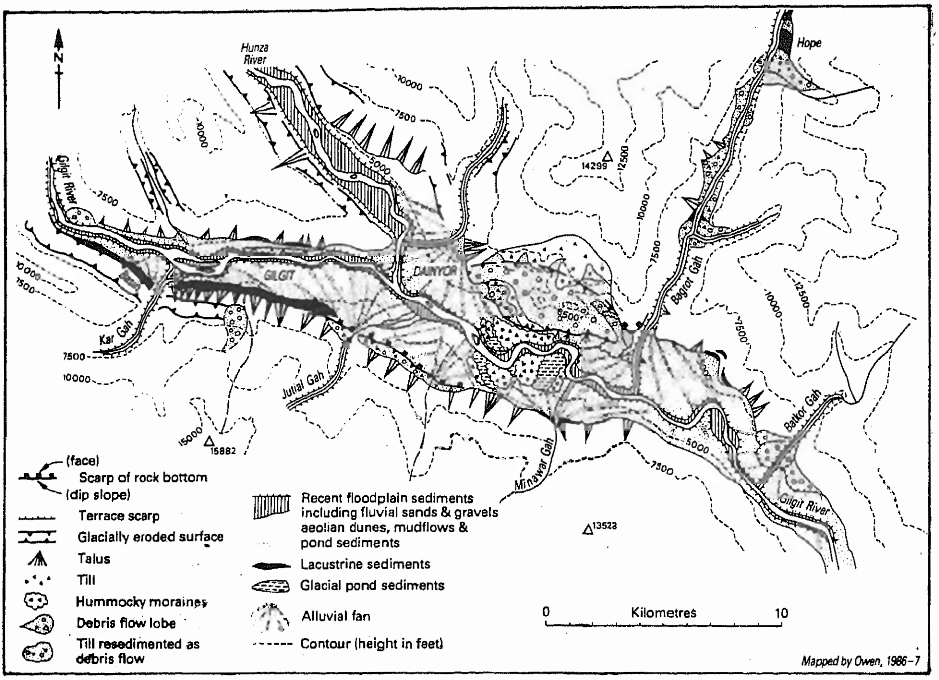


Fig. 2.12. Geomorphological map of the Gilgit Valley. (From Owen 1988a, 1989b).

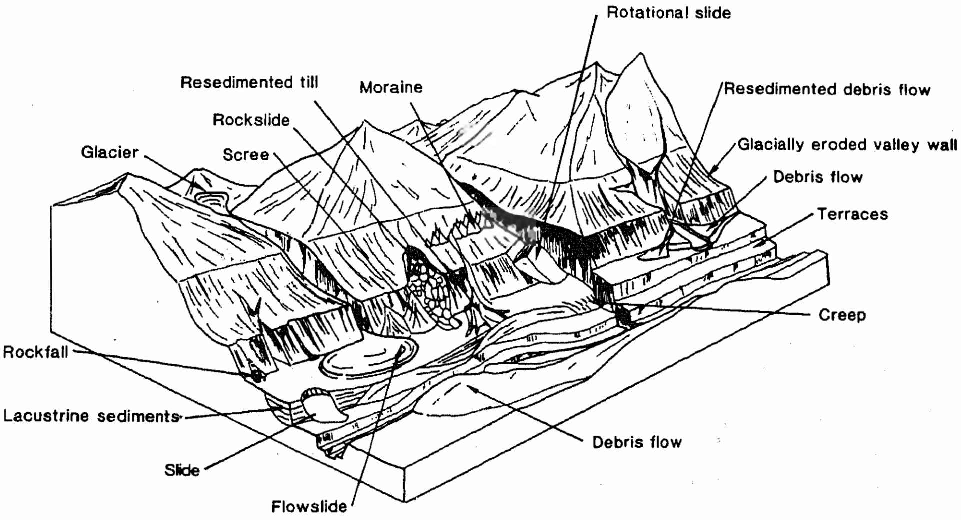


Fig. 2.13. Block diagram showing landforms and mass movement processes in Karakoram Valleys (from Owen 1989b).

Terrace remnants: Erosion of the valley slopes and the various land forms mentioned above has created a large variety of terrace remnants. Owen (1989b) has classified these terraces as follows:–

(a) *Morainic terraces*– These terraces have resulted from fluvial erosion of various glacial deposits which characterise the glaciated valleys of the Karakoram and Himalayas.

(b) *Glaciofluvial and fluvial terraces*– They encompass dissected ice-contact fans and outwash plain and are the product of allocyclic processes and climatic changes throughout the Quaternary.

(c) *Debris terraces*– Mass-movement deposits are widespread in the valleys of the Himalayas and the Karakoram and fluvial incision has formed extensive debris terraces along valley margins.

(d) *Lacustrine terraces*– The intermountain basins are characterised by fine, well sorted, silty and sandy lake deposits which formed in various types of lakes (formed due to damming by glacial ice, moraines or debris flow and landslides, or in glacially eroded, bedrock depressions). Upon infilling of the lakes or breaking of the dam, streams have incised the lake deposits to form these terraces.

(e) *Fan terraces*– According to Owen (1989b) these are polygenetic landforms comprised of debris flow, lacustrine, glacial, fluvioglacial and fluvial sediments and are one of the most common landforms in the Himalayan and Karakoram valleys.

Paleoplanation surfaces: Ancient planation surfaces are common in the Himalayas and the Karakoram Mountains. They occur in the form of flat, broad benches or rounded ice-smoothed slopes perched hundreds of metres above the present valley floors. One of these occurs at an altitude of >5,200 m in the Hunza Valley and is believed to be late Tertiary (Derbyshire el al. 1984). Another palaeosurface can be seen near Karimabad (Hunza) at an altitude of 4,100 m to 4,200 m and is believed to be a 'pre-Pleistocene' feature (Paffen et al. 1956). It has been named the 'Patundas surface' and according to Derbyshire et al. (1984), it was formed during the Shanoz Glaciation, which is the earliest.

Lakes: There are no large or significant natural lakes in the Karakoram–Himalayan region. However, small glacial lakes are numerous. The better known ones are the Saiful Muluk in upper Kaghan Valley, which fills up the depression formed by an old cirque, and the Satpura Lake near Skardu, which is the largest of these lakes and has formed in the depression of the floor of an ancient valley glacier. The upper reaches of the Hunza, Ghizar and other valleys contain small lakes in depressions formed in glacial deposits. The Borit Jheel near Pasu is a better known example of such lakes. There are numerous small tarns and paternoster lakes on the northern slopes of the Nanga Parbat near Babusar Pass and in other parts of this region.

Processes involved in mountain basin evolution

The various landforms discussed above are largely the product of complex cycles of weathering, and erosion and deposition through glacial, fluvial, lacustrine and aeolian processes. However, high magnitude-low frequency and catastrophic events have also significantly produced major landform modifications by triggering sudden mass movement activity (massive avalanches, rock and debris slides), erosion and resedimentation. According to Owen (1989b), the evolution of the Karakoram basins, their sediment fill and landforms has been controlled by (i) tectonics, (ii) climatic changes, (iii) allocyclic processes and (iv) catastrophic or high magnitude-low frequency events (Fig. 2.14).

The Lesser Himalayas and Subsidiary Ranges

The term "Lesser Himalayas" was first used by Burrard and Hayden (1908) to describe "a series of ranges closely related to the Great Himalayas". They divided the Lesser Himalayas into two classes, (a) those that branch from the Great Himalayas, e.g. the Pir Panjal, and (b) those that are separate folds. Wadia (1957) followed Burrard and Hayden and defined Lesser Himalayas as "a series of ranges closely related to the Great (or High) Himalayas but of lower elevation, seldom rising much above 12,000–15,000 ft. The Lesser Himalayas form an intricate system of ranges; their average width is fifty miles".

In Pakistan, the Lesser Himalayas comprise a series of ranges closely related to the High Himalayas but seldom rising to altitudes higher than 4,600 m. They form a complex system of ranges with sharp bends and abrupt changes in orographic axes, a feature described by such terms as arc, lobe, orocline, re-entrant, hairpin bend, syntaxis etc. The inner ranges are higher and parallel to the High Himalayas and comprise the northwest trending Pir Panjal Range, which links up with the southwest trending Hazara Mountains, through a sharp hairpin flexure (Photo. 11). This feature has been variously referred to as Northwest Himalaya Syntaxis (Suess 1904; Wadia 1931), "Punjab Orocline" (Carey 1958)" the Abbottabad Syntaxis" (Jones 1960), "Western Himalayan Syntaxis" (Gansser 1964), "Kashmir Syntaxis" (Calkins et al. 1975), "Hazara Syntaxis" (Desio 1976), "Punjab Re-entrant" (Johnson and Vohdra 1972) and "Jhelum Re-entrant" (Visser and Johnson 1978).

Parallel and subparallel to the NE-SW trending Hazara Ranges are the lower and smaller Margalla Hills, the Kalachitta Range, the Attock–Cherat Range, and the Safed Koh Range (Photo. 11). The north-south trending hill ranges of the Khyber Pass area are juxtaposed against the E-W trending Safed Koh Range. These mountains enclose the large, alluvial filled Peshawar Basin. South of the Margalla and Kalachitta Ranges there is a large intensely dissected plateau– the Kohat–Potwar Plateau – characterised by short, sharp, NE-SW or E-W trending ridges and intervening valleys. Their altitude ranges from about 300 m to 550 m and they are covered with thin vegetation, including patches and pockets of conifers at higher altitudes.

The scalloped Salt Range forms the southern margin of the Kohat–Potwar Plateau. Between the Jhelum and the Indus this plateau is about 150 km long and 90 km wide. It is traversed by numerous east-west or northeast-southwest oriented ridges, hogbacks and cuestas. The eastern component of the Salt Range, known as the Cis-Indus Salt Range, forms a northward sharp bend on either side (Photo. 11). The Trans-Indus Salt Range, to the west of the Indus, forms an S shaped flexure or syntaxis and is comprised of relatively small Marwat, Bhitani, Surghar and Khisor Ranges (Fig. 4.25). Steep cliffs and scarps form the southern margin of the Salt Range and stand nearly 600 m above the Indus Plain, whereas its northern flank is comprised of relatively gentler and wider dip slopes.

In Pakistan, the Lesser Himalayas and their subsidiary ranges are drained by the rivers Jhelum, Soan, Siran, Bara, Kohat Toi and Kurram, besides numerous smaller streams. They all eventually join the Indus River.

The Sulaiman-Kirthar Mountains

These mountains form the eastern part of Balochistan and extend for about 850 km in a north-south direction from south Waziristan up to the Arabian Sea (Fig. 2.2). Their width ranges from about 390 km in the north to about 190 km in the south. To

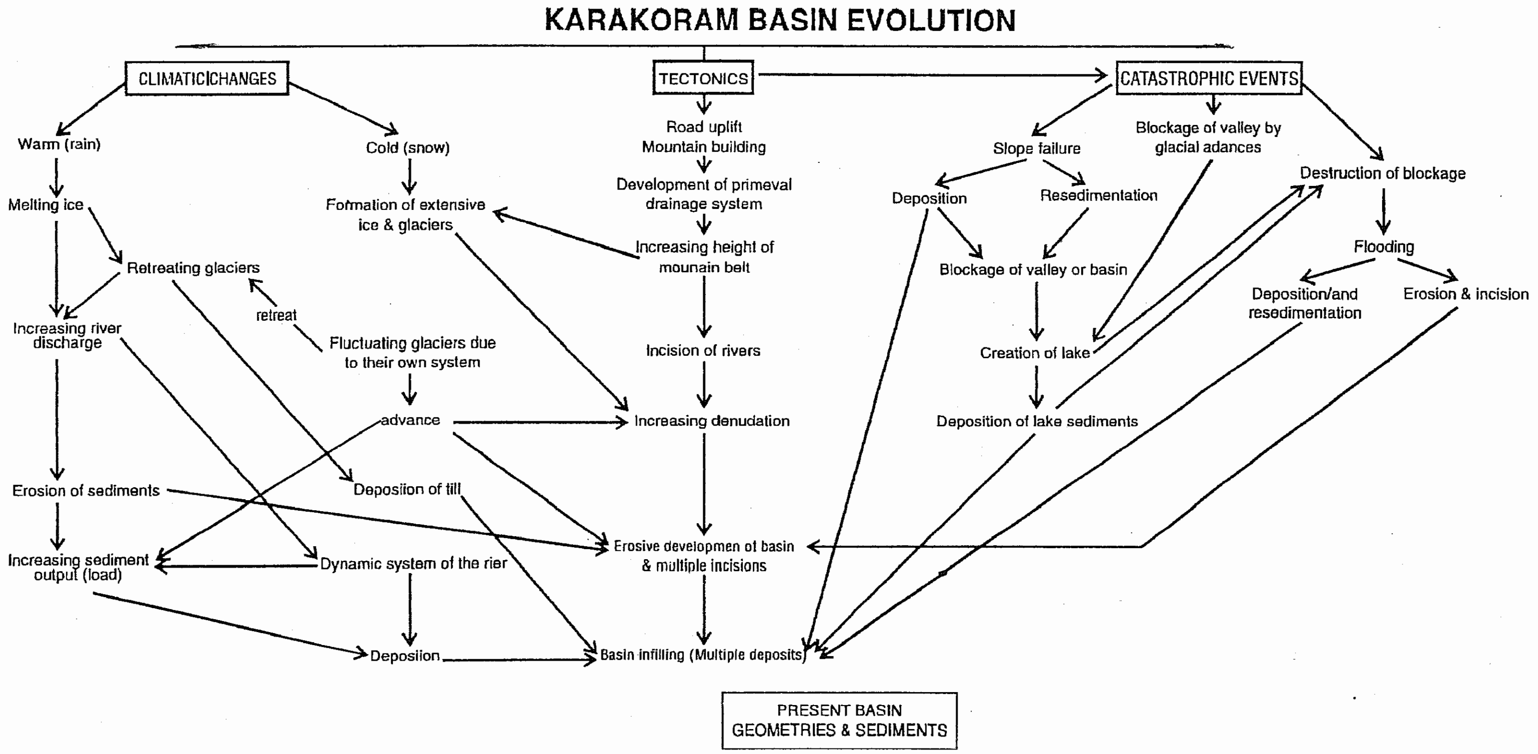


Fig. 2.14. Flow diagram showing the development of the Karakoram intermontane basins, and their sediment fills and terraces. (From Owen 1989b).

the west the Nal–Chaman faults, which run parallel and close to the 66th Meridian, truncate this mountain belt, whereas eastward the mountains overlook the vast Indus Plain. They are comprised of three distinct orographic zones: (a) the *Sulaiman Mountain Ranges* and their several arcuate branches form the northeastern part of this belt, (b) westward the hill ranges form the *Quetta Syntaxis* around the Zarghun Knot and (c) southward the *Kirthar Mountain Ranges* extend from the Zarghun Knot up to the Arabian Sea (Fig. 2.2 and Photo. 1)

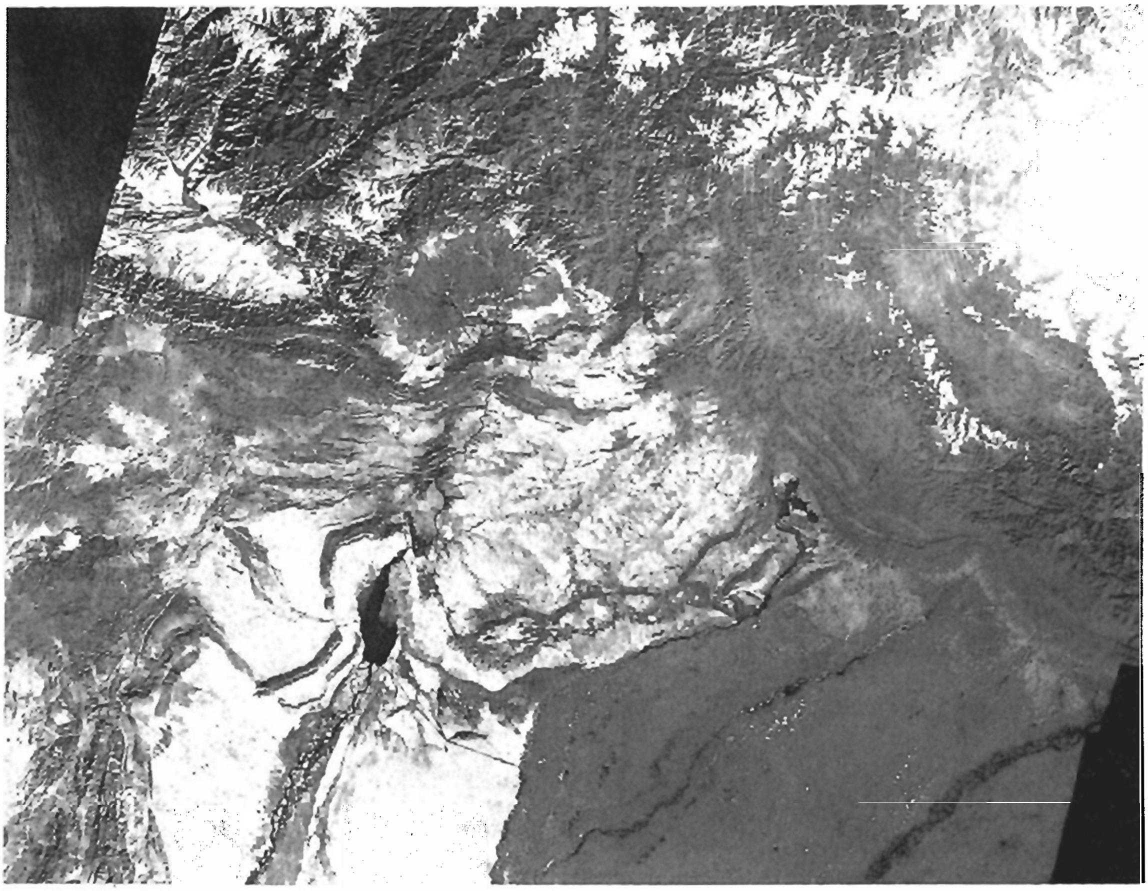
The Sulaiman Mountain Ranges

The Sulaiman Mountain Ranges have a lobate shape as a multistrand garland comprised of arcuate, parallel to subparallel and en echelon ranges a few tens of kilometres to over 300 km long. Their relief varies from a few hundred metres to nearly 3,600 m. The Sulaiman Mountains are thus comprised of the following orographic divisions (Photo. 12).

The Kakar Orocline: The northern-most strands of the Sulaiman system comprise the Kakar Oroclinal Ranges, namely the east-west trending arcuate Toba Kakar Range, followed to the south by the Kakar Khorasan Range which is about 240–280 km long. These ranges are relatively narrow (10–15 km) and are convex to the south and comprised of a series of parallel to subparallel en echelon hills interrupted by wide wind gaps or water gaps. Their altitude varies from about 2,200 m to 3,000 m. On either side of these hill ranges there are wide basin plains interspersed with small hillocks. Westward the Kakar Ranges merge with the north-south trending low ridges and hills of the Khawaja Amran and allied ranges close to the Afghan border. Eastward the Kakar Ranges gradually curve northwards and merge with the mass of hills in the Waziristan region on Afghan border.

The Sulaiman Composite Arc: South of the Kakar Orocline is the Sulaiman Arc which forms a nearly 300 km wide and 200 km long lobate mass of rather narrow, disjointed, parallel to subparallel, en echelon, arcuate and relatively low hill ranges and ridges which are convex to the south. Their altitude progressively decreases southward from about 3,400 m to 1,000 m. In the eastern part they have a southwesterly trend, then curve westward and finally swing northwestward. Four main but disjointed mountain arcs may be recognised south of Kakar Khorasan and the Zhob Valley. From north to south these are (a) The Shinghar–Khanozai Arc, (b) the Tor Ghar–Sanjawi–Mana Arc (c) the Musa Khel–Duki–Ziarat Arc and (d) the Kingri–Kohlu–Harnai Arc (Photo. 12). Forming the southern margin of the Sulaiman lobe and over-looking the Indus Plain, there are the eastwest trending Mari–Bugti Hills. They are about 120 km long and their altitude ranges from 600 m–1,000 m.

The Sulaiman Range: The most conspicuous mountain of the Sulaiman system of ranges, however, is the north-south trending Sulaiman Range itself. It comprises the eastern margin of the Sulaiman lobe and continues from South Waziristan to a point west of Rajanpur, a distance of nearly 400 km (Photo. 12). It is 20–25 km wide and its altitude ranges from about 3,440 m in the north to about 1,600 m in the south. This range is traversed by narrow, deep gorges and defiles of the Zhob, Drazinda, Domanda, Razanni, Vehowa, Luni, Kaha and Chachar Rivers. A slightly arcuate and disjointed chain of hills, the Drazinda–Pirkoh Range, flanks the Sulaiman Range on the east.



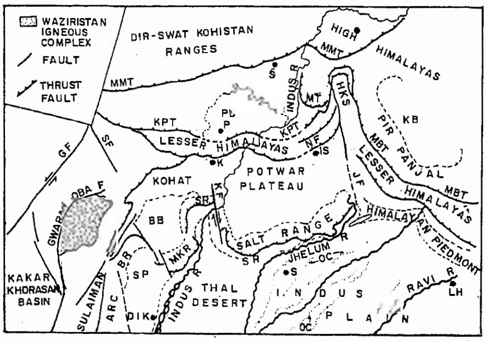


Photo. 11- Landsat photo mosaic showing the Lesser Himalayas and the subsidiary ranges. See Figure 26. (Courtesy U.S. Geol. Survey). BB–Bannu Basin, BR–Bhittai Range, DIK–Dera Ghazi Khan, GF–Gardez Fault, HKS–Hazara-Kashmir Syntaxis, IS–Islamabad, JF–Jhelum Fault, K–Kohat, KB–Kashmir Basin, KF–Kalabagh Fault, KPT–Khairabad-Panjal Fault, LH–Lahore, MBT–Main Boundary Thrust, MKR–Marwat-Khisor Ranges, MMT–Main Mantle Thrust, MT–Mansehra Thrust, NF–Nathiagali Fault, OC–Old Channels of Jhelum and Chenab Rivers, P–Peshawar, PB–Peshawar Basin, S–Saidu, SD–Sargodha, SF–Sarobi Fault, SR–Surghar Range, SRT–Salt Range Thrust.

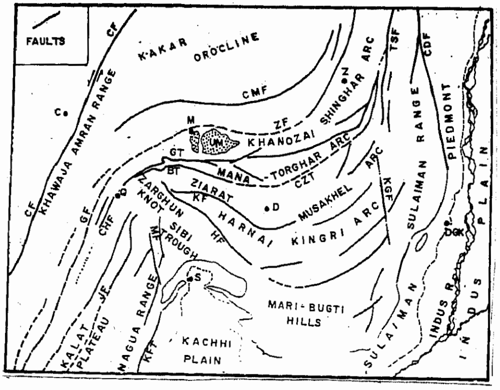


Photo. 12- Landsat photo mosaic showing the northern part of the East Balochistan fold-and-thrust belt. (Courtesy U.S. Geol. Survey). See Figures 4.20 and 6.12. BT–Bibai Thrust, C–Chaman, CDF–Chukhan Manda Fault, CZT–Chinjan-Zakriazai Thrust, DGK–Dera Ghazi Khan, D–Duki, GF–Ghazaband Fault, GT–Gogai Thrust, HF–Harnai Fault, JF–Johan Fault, JFT–Kirthar Fault, JF–Khalifat Fault, KGF–Kingri Fault, MF–Mach Fault, Q–Quetta, S–Sibi, UM–Ultramafic Rocks, Z–Zhob, ZF–Zhob Fault.

The Zarghun Knot and the Quetta Syntaxis

West of the Sulaiman system of ranges, between Quetta and Sibi, there is a triangular NW-SE to N-S trending mass of hills and ridges which are largely comprised of Neogene molasse sediments (Photo. 12). This orographic feature, about 190 km long and 50 km wide, forms a part of the Sibi trough (Sarwar and DeJong 1979). It has been also referred to as the Sibi re-entrant by Movshovitch and Malik (1965). Orographically it may be more appropriate to name this feature as the Zarghun-knot since the latter term, unlike "trough", "re-entrant" or "Syntaxis", has no genetic or structural implications and merely connotes a knot-like orographic feature. The most conspicuous part of this knot is the northeast-southwest trending 35–40 km long, 20 km wide Zarghun Range which bears the highest peak in Balochistan (Loe Sar 3,583 m). Unlike other ranges it is a mesa-like feature and instead of a sharp or well-defined crest it has a very wide, well-dissected and gullied flat top which is capped by massive conglomerates. It is, thus, often referred to as the Zarghun Massif. On either side of the Zarghun Range there are several parallel or en echelon, narrow ridges, punctuated by small oval or rounded hillocks. Southeast of Zarghun Range there is a NNE-SSW trending narrow anticlinal ridge that separates the Zarghun from a triangular synclinal basin. This basin is characterised by several parallel, sharp, narrow and inward dipping ridges.

The Zarghun Knot is like an orographic wedge that separates the E-W trending ranges of the Sulaiman lobe and the N-S trending ranges of the Kirthar belt. Near Quetta, however one can see the syntaxial bend of the Shin Ghar–Khanozai arc, changing from a westerly to southwesterly direction as it joins the Takhatu Range. The orographic axis then swings southward and follows the Chiltan Range. The term Quetta Syntaxis (Wadia 1957) has been thus used to denote the swing in the orographic axes of the hill ranges near Quetta.

The Kirthar Mountain Ranges

The Kirthar Mountains comprise a 560 km long and 130 to 220 km wide complex belt of north-south oriented mountain ranges, intervening valleys and basins (Photo. 13). From north to south its main physiographic divisions are as follows (Fig. 2.2).

The Kalat Plateau: It is a 1,600–2,000 m high, 250 km long and 110 km wide plateau, traversed by narrow hill ranges and wide intervening valleys and basins. The major hill ranges are the Unalath Range, Mashelakh Range, Chiltan Range, Koh-i-Maran, the Harboi Hills, the Brahui Range and the Central Brahui Range. Their altitude varies from 2,500 – 3,100 m. West of the Harboi Hills and the Brahui Range the main drainage runs parallel to the hill ranges. The major valleys in the region are Pahrod, Bhalla Dor, Shora Rud, Rej, Shirinab and Karanga Lora. However, east of Harboi–Brahui Range, the main streams (e.g. the Sarawan, Panch and Sukhleji) follow a transverse course across the hill range and flow through narrow sinuous gorges.

The Khuzdar Knot: South of Kalat Plateau, the N-S trend of the hills is abruptly terminated against east-west or southwest trending hillocks of the Khuzdar Knot, which covers a triangular tract about 75 km long and 60 km wide. The Khuzdar Knot contains a number of short narrow sinuous, V, U or W shaped sharp crested ridges, having E-W orientation in the northern part, followed by less distorted NNW-SSE trending ridges in the southern part. Wide expanses of flat, irregularly-shaped basin plains occur in between the hills. The altitude of the valleys varies from 1,000 m to 1,500 m, and of the hills from 1,800 m to 2,400 m. The knot is drained by the Soinda, Anjira, and Shahawar tributaries of the Mula River and the upper tributaries of the Kulachi River.

Las Bela Ranges: South of the Khuzdar Knot the hills divide into two groups. A westerly group follows a SSW direction and merges with the Makran Ranges. It comprises the Hazarganji, Kullit and Haro Ranges. The other group which lies to the east, is comprised of the SSE trending Khude, Pab and Mor Ranges. Between these two groups lies the wide, triangular, 120 km long and 80 km wide Porali Basin that extends southward up to the Arabian sea. The Bela Ranges are drained by the Kud, Porali, Kanrach, Windar and Hab Rivers (Fig. 2.2).

The Kirthar Range: The Kirthar Range (senso stricto) comprises the eastern part of the Kirthar Mountains. It is an approximately 400 km long and 30 km wide north–south oriented hill range and its altitude varies from about 1,000 m in the south to 2,400 m in the north. The northern part of the range (north of the Mula River) is known as the Nagau Range. It is drained by the Bolan, Panch, Sukhleji, Mula, Gaj, Nari and Anaai Rivers. Initially these rivers flow parallel to the hill range and drain its western slopes, but eventually they turn abruptly eastward across the hill range and flow through deep narrow defiles and gorges. Several small, dry channels dissect the eastern slopes of the range.

The Karachi Arc: This arc forms the southern-most part of the Kirthar Mountains and is comprised of a series of parallel to subparallel, short, narrow, serrate, arcuate (convex to east) en echelon ridges and wide, dome-shaped anticlinal hills. It forms a nearly 200 km long and 50 km wide zone between Karachi and Sehwan. The Bhit Range, Bhadra Range, Lakhi Hills, and Lakhra Hills are some of its more prominent components. The altitude of the hills varies from 250 m in the south to about 1,100 m in the north. The Naing, Baran and Malir Rivers are the main streams draining this region.

Characteristic features and land forms in the Sulaiman – Kirthar Mountains

In this region the mountain ranges are largely hogbacked anticlinal features with dip slopes on either side. The valleys and basins form wide synclines. The mountains in the north and west, however, comprise a thrust belt (see Chapter 4) and commonly have steep faulted scarps to the south or east and gentle dip slope on the opposite side. The synclinal basins and valleys are largely comprised of softer rocks and commonly contain parallel or concentric, low, scalloped, homoclinal ridges, hogbacks and cuestas. Often they contain mesas in their central parts. The Zarghun and Pishin basins are good examples.

The foot-hill region is commonly covered with gravel fans which form distinct piedmont zones. These are followed by sub-piedmont zones characterised by a gentler slope and finer sediments (sand and silt). Surrounded by gravel fans and well dissected and sloping apron of coarser sediments, in the centre there are valley plains, which are either flood plains, with entrenched streams or flat playas filled with fine silt and clay. The piedmont zone commonly comprises a series of four to five fan terraces. Some of the larger valleys, e.g., Mastung and Dasht, have scattered barchan type of sand dunes. These features are seen in the Makran Ranges and the Chagai–Ras Koh Ranges also.

The Makran Oroclinal Ranges

In the southwest corner of Pakistan, the Makran Ranges cover a vast area which is about 400 km long and 250 km wide (Fig. 2.2). This region comprises a monotonous series of parallel ridges and valleys (Photo. 14). The hill ranges form east-west to north-east trending arcs some of which (e.g., the Gokprosh-Garr Arc) are up to 500 km long. North to south the main hill ranges are the Siahan, Rakhshan, Central Makran, Gokprosh-Garr, Jhao, Rudia and the Makran Coastal Ranges. These ranges are largely comprised of a

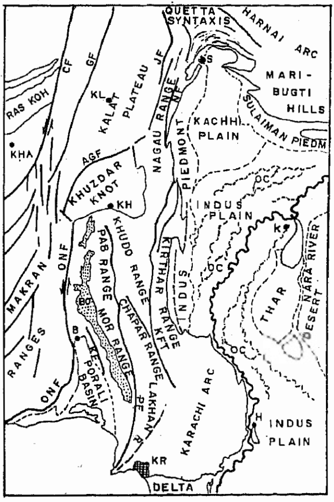


Photo. 13- Landsat photo mosaic showing the southern part of the East Balochistan fold-and-thrust belt. (Courtesy U. S. Geol. Survey). See Figures 4.4 and 6.11. AGF–Anjira-Gazan Fault, B–Bela, BO–Bela-Ophiolites, CF–Chaman Fault, GF–Ghazaband Fault, H–Hyderabad, JF–Johan Fault, K–Khairpur, KF–Kulri Fault, KFT–Kirthar Fault, KH–Khuzdar, KHA–Kharan, KL–Kalat, NF–Nagau Fault, OC–old channels of Indus River, ONF–Ornach-Nal Fault, PF–Pab Fault, S–Sibi.

series of parallel, sharp crested, serrate ridges capped by sandstones. Their southern slopes are commonly steeper. The altitude of the hills varies from about 2,000 m in the northeast to about 200–500 m in the south.

The main valley in this region are those of Rakhshan–Kap, Mashkai–Kolwa–Kech Kaur–Mand (over 450 km long), Nal Kaur and Dasht. These are relatively narrow features with entrenched streams. The main streams draining this region are the Mashkel, Hingol and Dasht Rivers. There are several smaller streams (Akara, Save, Shadi and Basol Rivers), which drain the coastal region only.

The Chagai-Ras Koh Ranges

These are relatively smaller, widely spaced and scattered mountain ranges which stand out conspicuously in a wide, open, flat expanse of an enclosed drainage basin (Photo. 15). This basin is covered by a stony-waste desert in the north and vast dry playas and extensive dune fields in the south. Apart from several dry nalas, the main streams draining into this basin are the Tahlab, Hamun Lora, Baddo and Mashkel Rivers. There are four main hill tracts in this region. The western most is the Mirjawa Range, which is a NW-SE trending, narrow, broken chain of low hills along the Iran-Pakistan border between Ribat and Hamun-i-Mashkel. It is about 240 km long, 5 to 20 km wide and its altitude varies from 700 m in the south to over 2,000 m in the north.

Another chain of low hills pockmarked by a few craters of extinct volcanoes e.g., Koh-i-Dalil (1,484 m), Damo Din (1,890 m) and Koh-i-Sultan (2,332 m), runs E-W along the Pak-Afghan border. Eastward it is disrupted by a 50 km wide wind gap filled with sand dunes. It continues eastward in the form of Chagai Hills which are 130 km long, 50 km wide and have an altitude of 2,000–2,400 m. They have a wide apron of sand dunes along their northern slopes and are surrounded by a vast (20–24 km) fringe of coarse stony piedmont zone or bahadas. Where this chain of hills is broken by water and wind gaps, the piedmont zone is intermittently covered with long, narrow, N-S oriented sand dune ridges which are parallel to the drainage. An impressive radiating drainage pattern surrounds the Koh-i-Sultan. A close-knit, dry, ephemeral, dendritic, parallel to subparallel drainage network extends southward into Hamun-i-Mashkel playa or flows eastward into the smaller Hamun-i-Lora playa.

South of Hamun-i-Lora there is the very narrow and low chain of E-W trending Rakhshani Hills, 180 km long, 2 to 5 km wide and about 50 m to 100 m high. Parallel to the Rakhshani Hills about 20 km to the south, there is the Ras Koh Range–the largest and the most conspicuous orographic feature of this region. It is 230 km long, 50 km wide and its altitude varies from 1,500 m to 3,000 m. It is like a wedge in the eastern part of the enclosed basin of Hamun-i-Mashkel and separates the Baddo River drainage from the Dalbandin Valley to the north.

INTERMONTANE BASINS

There are numerous intermontane basins and it is beyond the scope of this book to describe them individually. Most of them are structural or tectonic depressions. In the Sulaiman–Kirthar region the valleys and basins are synclinal whereas in the Makran region they are tectonic and follow large arcuate reverse or thrust faults. In the Himalayan region where erosion has been more active, valleys are anticlinal and were carved out of softer argillaceous rocks. Along the foot hills these valleys and basins have a sequence of four or five Quaternary terraces, followed by a gravelly piedmont zone and a relatively flat central alluvial plain. They have formed during the Neogene phase of the Himalayan Orogeny.

The Hamun-i-Mashkel, Peshawar and Potwar (Fig. 2.2) are exceptionally large basins with approximate areas of about 24,000, 5,000 and 16,000 km^2 respectively. The Hamun-i-Mashkel forms a part of the Makran arc-trench gap. The Peshawar and Potwar basins have been carried passively southward on backs of thrust sheets during the development of the Main Boundary Thrust during the Neogene (Burbank and Raynolds 1984) and consequent uplift of the Lesser Himalayas and the Salt Range (see Chapter 4).

THE INDUS PLAIN

The Indus Plain forms the eastern part of Pakistan. It is bounded by mountain ranges to the north and west, by the Arabian Sea to the south and the Thar Desert to the east (Photo. 1). It is about 1,200 km long and has a maximum width of about 400 km in the north. It is narrow (about 12 km) in its central part and again becomes wider (160 km) in its southern and lower part (Fig. 2.2). The most conspicuous relief features in the Indus Plain are a number of monadnocks between Shahkot and Chiniot and small hillocks near Khairpur, Hyderabad and Thatta. In the north these attain an altitude of about 250 m above sea level, their tops standing as much as 100 m above the plain.

The Indus Plain is a complex geomorphological unit and is composed of two broad, distinct, geographical and geological divisions, namely (a) the Piedmont zone, here referred to as the Indus Piedmont and (b) the Central Alluvial Plain, commonly referred to as the Indus Plain. Inasmuch as these features have formed through accumulation of a vast thickness of sediments brought down by the Indus and its tributaries, at this point a brief description of the general drainage would be appropriate. This is then followed by an account of the Indus Piedmont and the Central Alluvial Plain.

Drainage

The Indus Plain is drained by the Indus and its tributaries. Together they have a catchment area of about 440,000 km^2. There are three main types of streams in the Indus drainage system, the perennial streams, the misfit streams and the dry or ephemeral streams.

Perennial streams

The perennial streams vary in size from large rivers with discharge in thousands of cusecs to small streams with discharge ranging from 0.3 to 3 m^3 sec (Figs. 2.15 and 2.16). Northeast to southwest, along the mountain front, some of the more significant smaller streams are the Degh, Tawi, Bhimber, Poonch, Soan, Kurram, Gomal, Chakar, Beji, Nari, Bolan, Mula and Gaj Rivers (Fig. 2.2). Most of these streams drain arid or semi-arid regions with relatively small rainfall. Their perennial discharge is largely due to effluent seepage and their catchment areas vary from about 1,300 to more than 100,000 km^2. They are subject to flash floods and the flood water is heavily loaded with sediments (Fig. 2.16). The northern streams carry the sediment load to major rivers, whereas the southern streams deposit their load along the foothills.

The larger perennial streams in the Indus Plain are the Indus, Jhelum, Chenab, Ravi, Beas and Sutlej. In the upper reaches these rivers are moderately entrenched in their flood plains. The width of the river beds ranges from about 0.5 to 3 km, whereas the flood plains are 16 to 50 km wide. The Indus is one of the world's largest rivers in

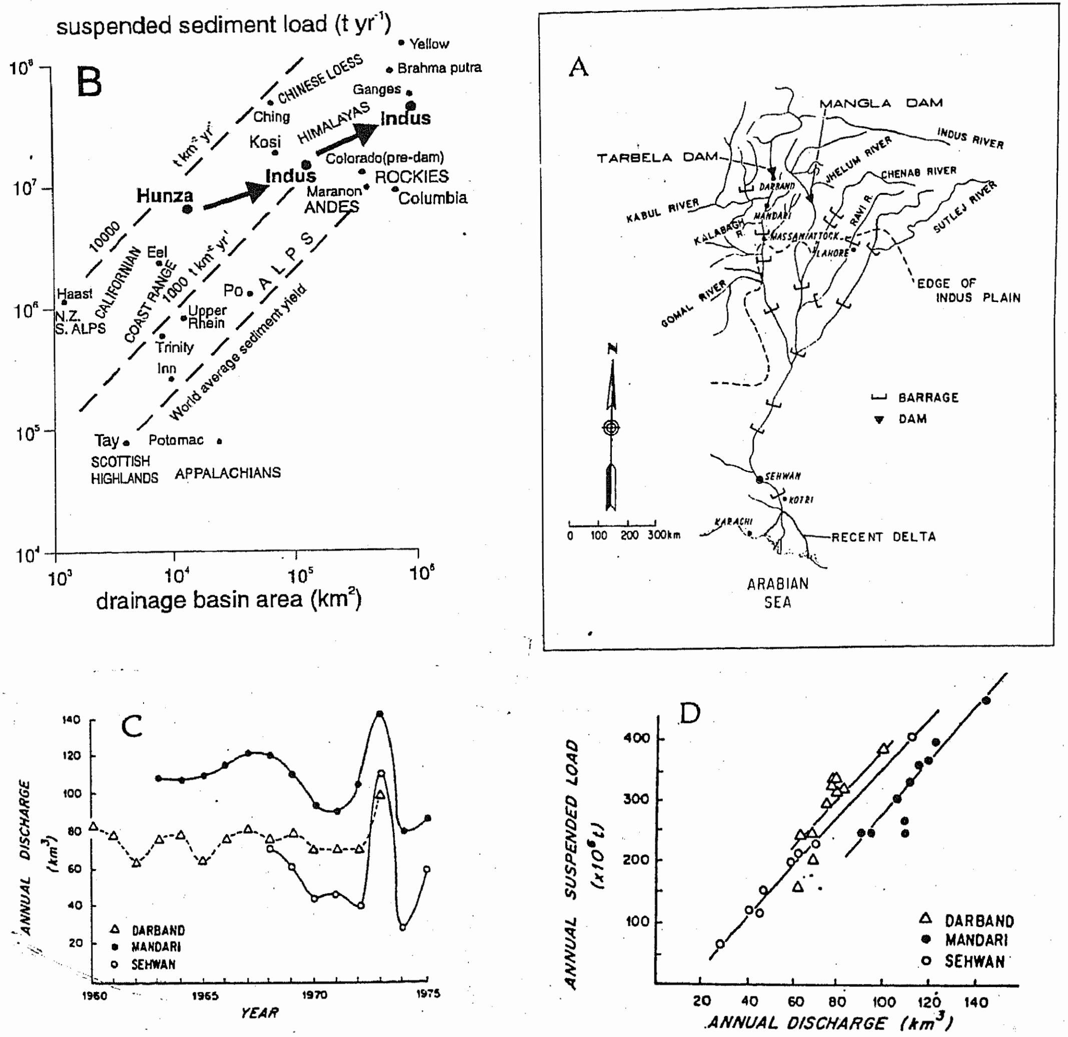
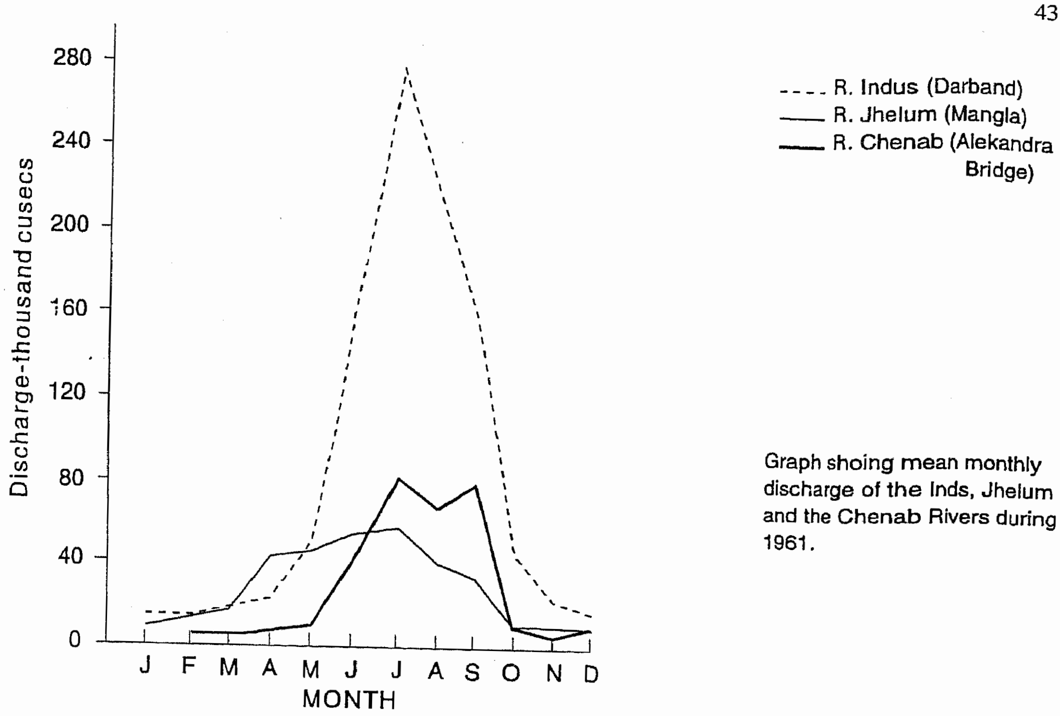
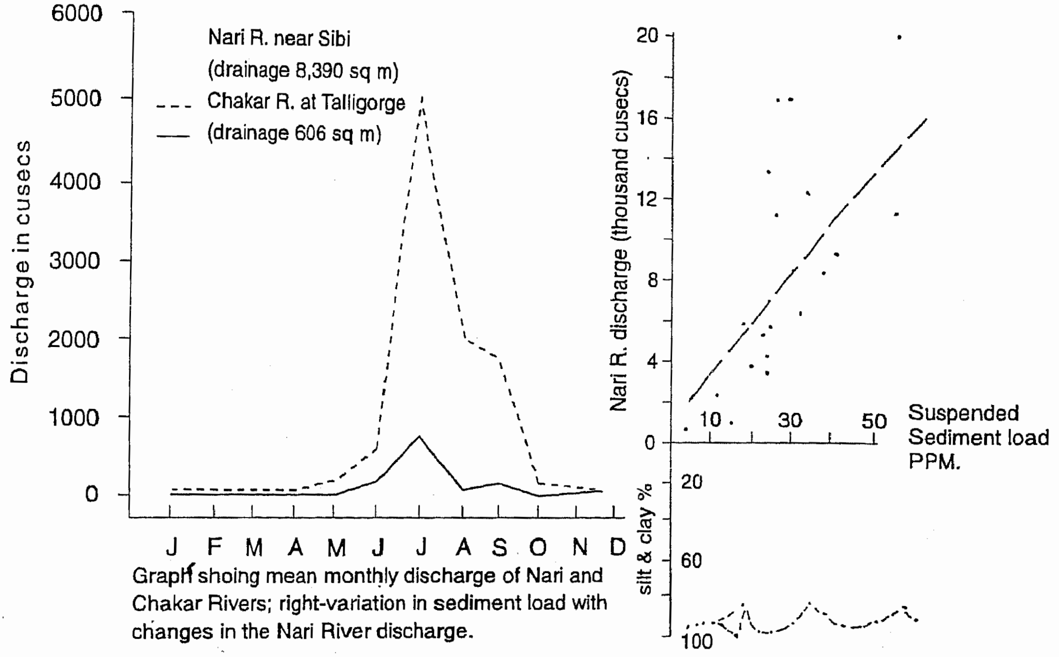


Fig. 2.15. Discharge and sediment load of River Indus. A–Map of the Indus Basin, showing location of dams and barrages. B–Sediment load of Hunza and upper Indus Rivers compared with other mountain rivers (from Ferguson 1984). C–Annual discharge of the Indus at Darband, Mandri and Sehwan. D–Annual discharge and sediment loads of the Indus at the above stations. (From Milliman et al. 1984; WAPDA 1979).

terms of drainage area, river discharge and sediment load. From its source to the delta it has a length of about 2,880 km. Its total drainage area is 970,000 km², out of which the hilly catchment covers about 264,000 km². Its average discharge at Mandori, 8 km below Attock bridge, for 92 years (1869 to 1961) has been computed to be about 3,528 m³/Sec (WAPDA 1970). However, according to some sources (Lisitzin 1972, Meybeck 1967), the mean discharge of the Indus varies from 5,550 to 7,500 m³/Sec. Its discharge, as it emerges from the mountains and before it joins the delta, is given in Figure 2.15. Its flood-discharge often exceeds 28,000 m³/Sec. After descending on the plain near Kalabagh, the river has a wide braided channel. However, a few miles down



Graph shoing mean monthly discharge of the Inds, Jhelum and the Chenab Rivers during 1961.



Graph shoing mean monthly discharge of Nari and Chakar Rivers; right-variation in sediment load with changes in the Nari River discharge.

Fig. 2.16. Graphs showing monthly discharge of Indus, Jhelum, Chenab, Nari and Chakar Rivers. (From Kazmi 1977a).

Table 2.5. Discharge, suspended sediment load and water quality of Indus and other selected rivers of Pakistan (from WAPDA discharge book).

	Gauging station	(sq. mile) Drainage area	(cusecs) Average Annual Discharge	Suspended sediment load			Ec x 10^6 at 25 ^{0}C	pH
				Average annual concentration	Sand %	Silt & clay %		
INDUS	Mandori	102,000	126,000	2,000 ppm	44	56	170 to 300	8
	Sehwan	-	73,000	3,540 "	15	85	210 - 470	8.2
CHENAB	Alexandra Bridge	12,600	21,600	2,120 "	28	72	150 - 370	7.5-8.2
JHELUM	Mangla	12,900	31,000	3,120 "	38	62	150 - 240	7.7-8.2
SOAN	Dhok Pathan	2,500	1,600	3,500 "	10	90	290 - 520	7.9-8.5
GOMAL	Kot Murtaza	13,900	536	3,800 "	5 - 8	92 - 95	920 - 1,600	7.8-8.4
CHAKAR	Talli Tangi	573	70	3,000 "	3 -12	88 - 97	850 - 1,510	8-8.4
BOLAN	Kundlani Bridge	1,560	73	3,670 "	2	98	690 - 900	8-8.45
MULA	Naulang	3,320	110	4,620 "	3	97	- -	
HAB	Bund Murad	3,640	390	-	-	-	1,720 - 1,950	8.2-8.4
PORALI	Sinchi Bent	1,560	126	-	4 -12	88 - 96	400 - 830	8.2-8.5

Table 2.6. Annual runoff and suspended sediment load from Karakoram River Basin.

River and site	Period of record	Area Km^2	Runoff		Sediment load per annum	
			m^3s^{-1}	mm a^{-1}	Mt	t. km^{-2}
1 Hunza (Dainvor Bridge)	1966 - 75	13,200	380	910 (640-1,170)	63 (28-101)	4,800
2 Gilgit (Gilgit Town)	1963-72	12,100	280	740 (690-830)	14 (4-28)	1,100
3 Gilgit (Alam Bridge)	1966/7/9-75	26,200	700	840 (620-1,020)	70 (33-104)	2,700
4 Shyok (Yugo)	1973-5	3,3700	310	290 (230-390)	34 (21-55)	1,000
5 Indus (Kachura)	1970-5	112,700	960	270 (240-370)	87 (57-133)	770
6 Indus (Partab Bridge)	1963-75	142,700	1,760	390 (290-530)	160 (81-201)	1,100

Source: WAPDA 1976, converted to SI units. Figures in brackets are ranges of values. Basins 1 and 2 lie within 3; 4 within 5, 3 and 5 within 6.

stream of its confluence with other rivers, it acquires the usual meandering character which may be seen all the way up to the sea (Photo. 1).

Relevant information concerning discharge of the Indus, Jhelum, Chenab, and other Rivers is given in Table 2.5. The largest rivers are fed by the melt-waters of the Himalayan glaciers. Consequently the lowest discharge of water is recorded in the winter months and the peak run off during the summer melt-water season and the monsoons (Figs. 2.15 and 2.16). A considerable amount of the sediments carried by the Indus is derived from glaciers or debris flow in the catchment (Table 2.6). As with discharge, the sediment load also varies with both time and distance along the river (Fig. 2.15).

During the winter months, the Himalayan rivers transport relatively fine suspended load. During this period there is little deposition of sediments on the flood plain. Most of the coarse debris is acquired through lateral corrasion and is transported only a short distance downstream where it is deposited along the meander beds. During the monsoon floods, large quantities of sediments are deposited on the flooded areas of the plain and vast tracts of land along the river beds are subjected to erosion.

Prior to the canals, the rivers were in a near equilibrium state and had a tendency to aggrade in the upper reaches. The natural regime of the rivers has been considerably modified due to the construction of flood protection levees and barrages. More than 70 percent of the annual discharge of the rivers is now diverted into the present canal-irrigation system. In the upper reaches, above the Indus-Sutlej confluence, the meander belts are very wide and in this region the coarsest sediments have been deposited. Downstream of the confluence, near Sukkur, the meander belt of the Indus abruptly becomes narrow as it flows through the water gap across the hill at Rohri. This constriction functions like the throat of a Parshalls flume. The velocity of water increases in the gorge and the latter is kept relatively free from sediments. Nevertheless it has caused active aggradation upstream of the gorge and slight degradation downstream.

The suspended load of the master streams reaches a peak of 6,000 to 9,000 ppm with sand content of 20 to 65 per cent (WAPDA 1970). The bed-load entirely comprises medium to coarse sand. Thus more than 90 per cent of the sand deposited on the Indus Plain during the Late Quaternary has come through the Himalayan rivers. The mountains to the west are mainly composed of limestone and shale with lesser amounts of sandstone and the debris in the flood discharge from this sedimentary cover does not contain any significant amount of sand. The smaller streams such as the Soan, Gomal, Chakar, Beji, Nari and Bolan which drain the mountains west of the Indus Plain, carry high sediment loads (20,000 to 190,000 ppm) during floods (WAPDA 1970). This load contains only 4 to 10 per cent sand which rarely reaches a value of 15 to 20 per cent (Table 2.5, Fig. 2.16). The high silt and clay load (commonly 80 to 99 per cent) is deposited in the subpiedmont zone. The extensive deposits of silt and clay along the margins of the Indus Plain, therefore, largely belong to this facies (see Chapter 5).

Misfit streams

The Indus Plain bears abundant scars of old abandoned courses of the present master streams. The old courses have been silted up and are at present under cultivation. Only a few of these still bear misfit stream channels which carry excess flood waters during high floods. These channels contain marshy tracts or oxbow lakes and are locally known as Wah, Rohi or Buddh nalas. Best known examples are the old channels of the Beas near Kasur, old Hakra River near Fort Abbas and the East Nara Canal. These streams have been described in some detail by Ahmad and Kazmi (1963), Kazmi (1966) and Flam (1992).

Ephemeral streams

Along the foot-hills numerous dry nalas descend on the Indus Piedmont zone from the outermost ridges overlooking the Indus Plain. Their catchments are commonly less than 16 km^2 in area and are largely composed of Siwalik siltstones, sandstones and conglomerates. After heavy rain, sediment laden flood water spreads over the piedmont and subpiedmont zones. This flood water contains more than 85 percent silt and clay

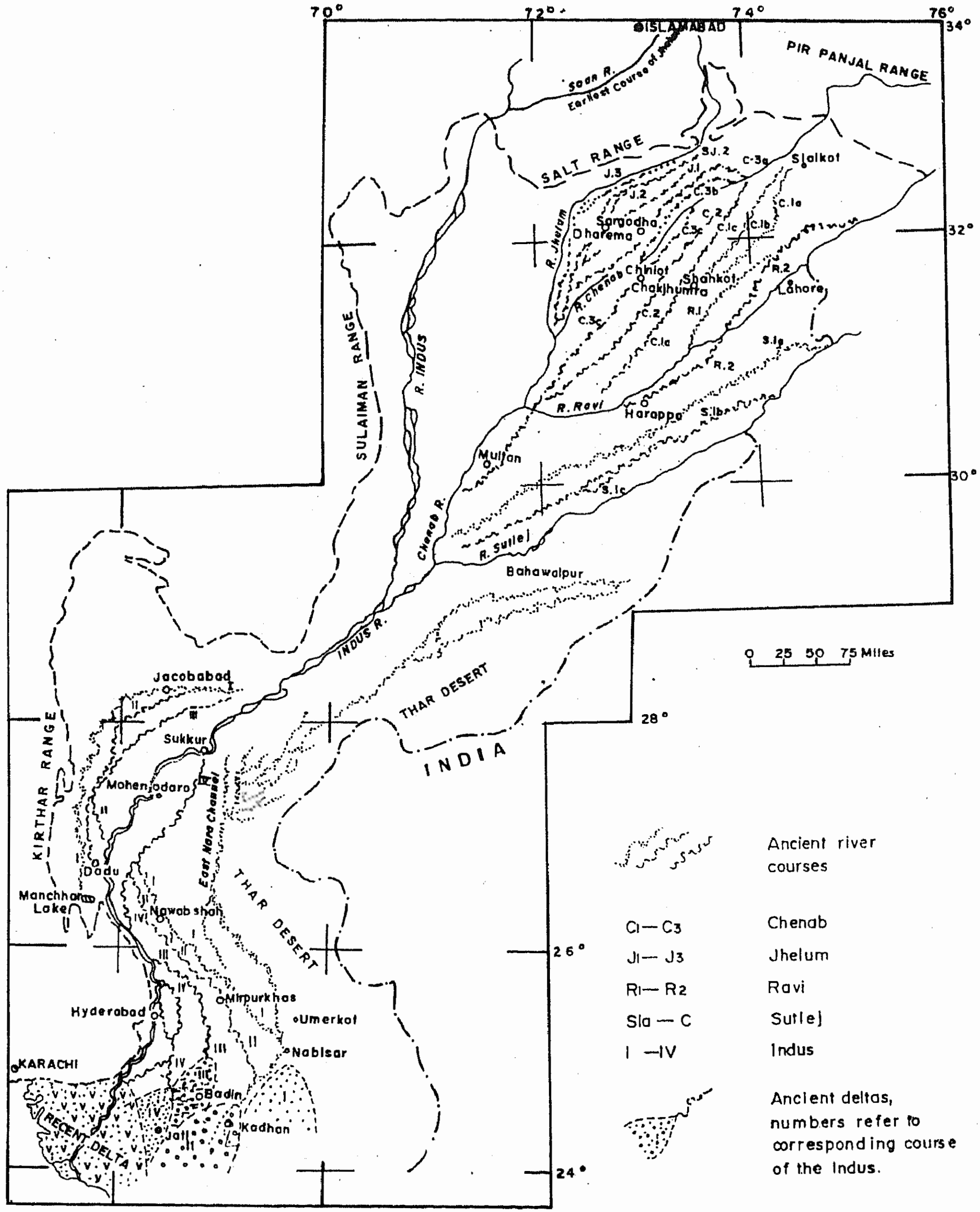


Fig. 2.17. Map of the Indus Plain showing courses of the extinct streams. (From Kazmi 1966, 1977a).

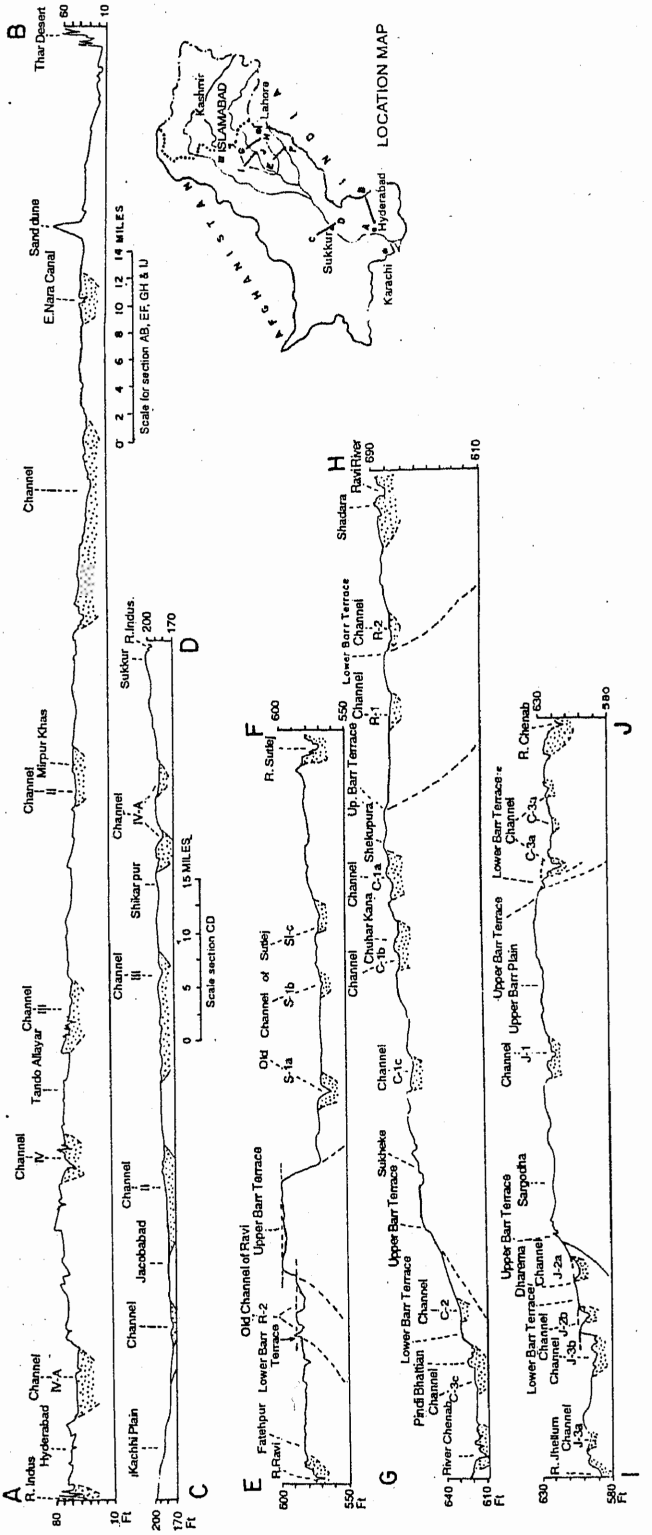


Fig. 2.18. Sections showing topography and geomorphic features of the Indus Plain. (From Kazmi 1966, 1977a and 1984).

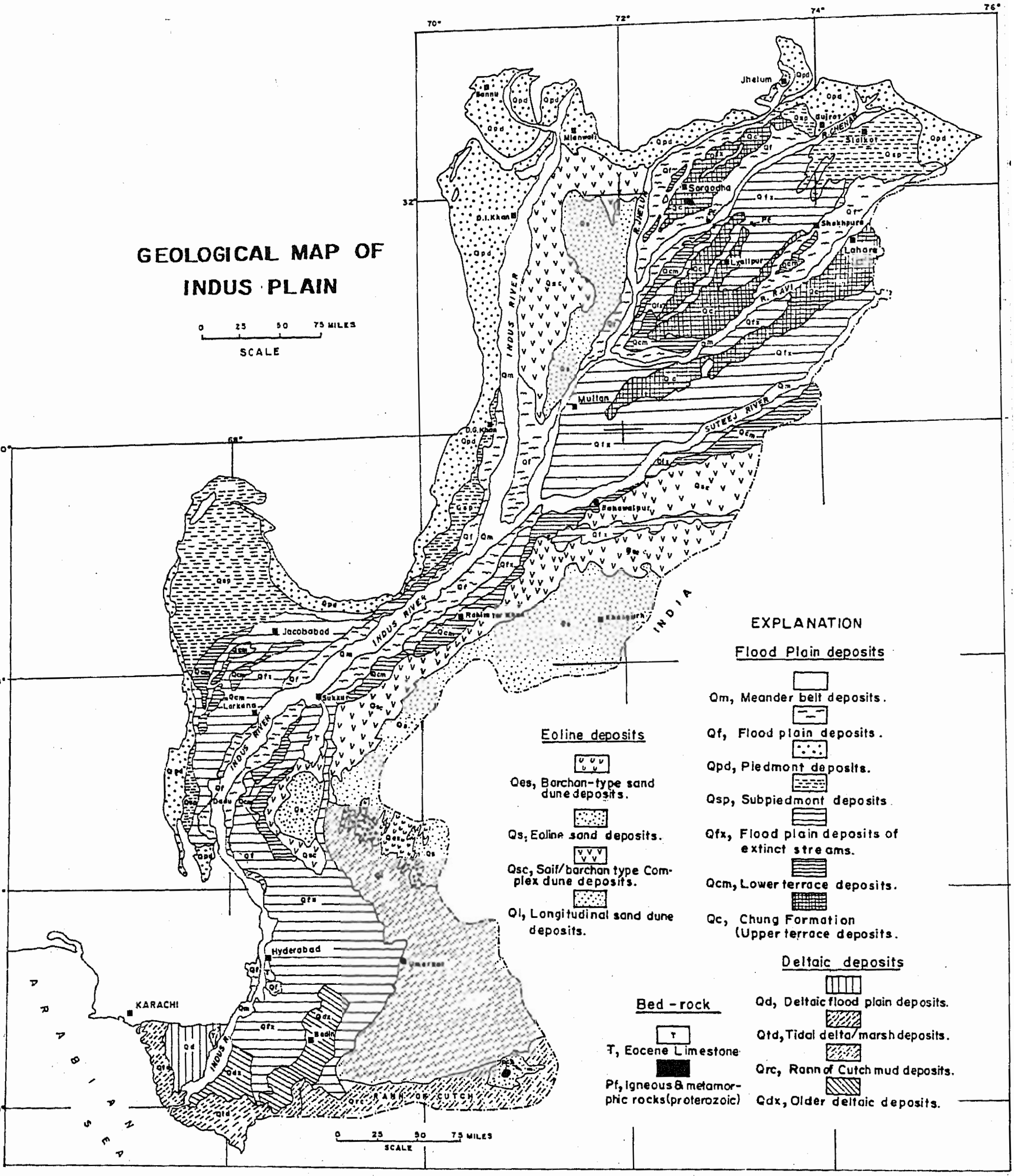


Fig. 2.19. Geological map of the Indus Plain. (From Kazmi 1966 and 1977a).

(Table 2.5 and Fig. 2.16). The fine silty sediments of the subpiedmont zone have originated in this way.

Drainage changes and ancient courses of the Indus and its tributaries

The old courses of the Sutlej, Ravi, Chenab, Jhelum and the Indus have been traced and mapped by Kazmi (1966, 1977) and these have been shown in Figure 2.17. The profile across Bari Doab (area between Sutlej and Ravi) shows that the old courses of the Sutlej are at a lower elevation than its present course (Fig. 2.18). Together with other features such as old levees, alluvial ridges and meander cut-offs, it suggests that the Sutlej has aggraded its flood plain in stages and has shifted eastwards. Its three successive old courses (S-1a, S-1b and S-1c) are shown in Figure 2.17. These courses are Recent in age and the alluvium deposited during this aggradational phase is only 3 to 16 m thick (Kazmi 1966, 1977).

Old stream courses of the Ravi

West of Lahore the wide flood plain of the Ravi is full of meander scars. It appears that the Ravi has been changing its course more frequently than the other streams and has not been confined to a single meander belt for any significant length of time. The meander scars indicate that unlike other rivers, changes in its course have been gradual. However, from the density of meander scars, sand bars and traces of oxbows, it may be inferred that its earliest course was about 16 to 19 km west of Lahore, close to Shekhupura, Nankana Sahib and Saiyidwala. It cuts through the Lower Barr Terrace and may be referred to as the Nankana course (Rl in Fig. 2.17). Eastward this is followed by the Sharqpur course (R 2). Finally, from a point near Narowal, the Ravi has shifted eastward again and has occupied its present course.

Old stream courses of the Chenab

The Chenab has long been the most active and powerful stream in this region after the Indus (Kazmi 1966, 1977). Its aggradational history is revealed by traces of its old courses and may be divided into three distinct stages. Initially the river flowed along the central part of the present interfluve in Rechna Doab and followed a course from near Sialkot, through Gujranwala area to Shahkot and beyond. It formed a wide flood plain between Hafizabad and Shekhupura. This period has been referred to as the Shekhupura aggradation stage by Kazmi (1966, 1977). At least three channels of this stage may be identified (C-1a, C-1b and C-1c in Fig. 2.17). The earliest channel is the easternmost and has the faintest meander scars. After abandoning this channel the river gradually moved westwards and between Shahkot and Samundri, it was entrenched on the Upper Barr Plain and formed a wide flood plain. The Barr Plain occupies a higher level above this entrenched flood plain and is an older feature. The flood plain deposits of the Shekhupura stage largely consist of sand, silt and clay and resemble the present deposits of the Chenab. The large meander scars (6 to 9 km across), suggest that the Chenab then must have been as big a stream as at present. It probably flowed under climatic conditions not very different from the present. This course therefore belongs to an interglacial or interpluvial period.

During the next period of aggradation, the Chak Jhumra stage, the river abruptly moved westward, occupying a lower level than it did during the Shekhupura stage, and flowed past Hafizabad and Chak Jhumra, towards Shorkot (C2 in Fig. 2.17). The present flood plain belongs to the third and final aggradational stage of the Chenab and it separates the Shekhupura and Chak Jhumra stages from most of its subsequent courses. During this last stage, which may be referred to as the Phalia stage, after the village of that name, the Chenab shifted farther westward, degraded its profile, and formed the slope of the Lower Barr terrace. It carved out an entrenched valley which was subsequently filled with sediments. Only the vestiges of the last phase of this aggradational stage can be seen on the surface in the form of the present flood plain, on which three abandoned channels of the Chenab may be seen (C3a, C3b, C3c in Fig. 2.17). The course C3a is the oldest amongst these and marks the westernmost limit of its migration.

Old stream courses of the Jhelum

Like the Chenab, old courses of the Jhelum also comprise three aggradational stages. During the earliest stage, the Bhalwal stage (J1 in Figs. 2.17), the river flowed along the centre of the Upper Barr Plain of the Chaj Doab. This stage may be correlated with the Shekhupura stage of the Chenab. Subsequently, during the second or the Dhrema stage, the Jhelum moved westward and occupied a shallow flood plain on the Lower Barr Plain of the Chaj Doab (J2a and J2b in Fig. 2.17). According to Kazmi (1966, 1977) the Dhrema stage of the Jhelum may be correlated with the Chak Jhumra stage of the Chenab.

During the last or the Shahpur stage, the Jhelum again shifted westward and formed the lower terrace near Dharema. It degraded its course in the shape of an entrenched valley and later filled it up to form the present wide flood plain (J-3a, J-3b in Fig. 2.17). The river has now become slightly entrenched in its present flood plain.

According to Kazmi (1966, 1977) the Bhalwal, Dhrema and Shahpur stages of Jhelum have the same history and bear the same position with reference to the Upper and Lower Barr Plains and the two Barr terraces respectively, as the Shekhupura, Chak Jhumra and Phalia stages of the Chenab and are probably of the same age. The two stages of the Ravi described earlier, correlate with the last two stages of the Chenab and the Jhelum. The old courses of the Sutlej probably correspond to the last stage of the other three rivers.

Old courses of the Indus and ancient deltas

In the upper part of the Indus Plain, sand dunes of the Thal Desert completely conceal the underlying alluvium and the former courses of the Indus. However, southward from Kashmor, five courses of the Indus can be seen on the Indus Plain (Kazmi 1966, 1977, 1984). Whereas in the north the rivers occupy lowest parts of the plain, in the southern region the present course of the Indus is on an alluvial ridge, which has a higher altitude than any of its former courses (Fig. 2.18). Its earliest courses are characterised by faint and poorly-preserved meander scars, channel and levee remnants, which occupy the low areas farthest from the present channel. Thus, four main aggradational stages of the Indus are found in this area. From the oldest to the youngest they are as follows :–

(I) *Jacobabad–Samaro stage:* During this stage the river flowed past Jacobabad, Usta Muhammad, Nawabshah and Samaro.

(II) *Shahdadkot–Jhudo stage:* There are channel remnants of this stage near Shahdadkot, Manahijo, Mirpur Khas and Jhudo.

(III) Qambar–Tando Allahyar stage: Meander scars, oxbows and levee remnants of this stage are traceable from Qambar to Dadu and Hala to Tando Allahyar and beyond.

(IV) Gambat–Tando Mohammad Khan stage: Relatively fresh channel scars, sand bars and swales, oxbows and levee remnants of this stage may be seen from near Khairpur to Daulatpur, Matiari, Tando Mohammad Khan and farther southward.

These courses can be traced southward and link up with remnants of ancient deltaic complexes. The earliest delta corresponding to the Jacobabad-Samaro stage is covered by the sand dunes south of Nabisar, as indicated by shallow bore hole data from that area (Kazmi 1966, 1977). With the westward migration of the Indus, the position of the delta has also shifted westward. Remnants of ancient deltaic flood plains corresponding to the ancient courses of the Indus are shown in Figure 2.17. These features represent the later part of the aggradational stage of the Indus following the last glaciation and correlate with the Phalia stage of the Chenab River.

Besides the Indus, there are traces of another ancient river - the Hakra, which once flowed east of the Indus, along the margin of the Thar and Cholistan Deserts. It has slowly dried up since the last 2,600 years (Wilhelmy 1969). The East Nara channel is a remnant of the ancient Hakra and its trace through the desert have been mapped (Kazmi 1966, 1984) and shown in Figure 2.17.

The earlier geological history of the Indus Plain has been discussed in Chapter 5.

The Indus Piedmont

This piedmont zone forms a continuous, sloping apron along the foothill region on the northern, northwestern and western margins of the Indus Plain (Fig. 2.19). It is about 16 to 24 km wide though in the Dera Ismail Khan area it widens to 45 to 55 km (Photo. 1). It covers an area of approximately 46,000 km^2 and has the steepest gradient amongst the land forms constituting the Indus Plain. It comprises two distinct physiographic units, an upper, more steeply inclined Piedmont Plain, which has developed from the coalescing of gravel fans, and a lower, gently sloping Subpiedmont Plain, which has resulted from the deposition of finer sediments. The slopes of the Piedmont and Subpiedmont Plains range from about 20 m/km to 10 m/km, and 5 m/km to about 2 m/km respectively. There is usually a perceptible break in slope between the two zones and their boundaries can easily be traced on ground as well as on topographic maps, aerial photographs and satellite imageries.

In the north (Shakargarh, Sialkot, Mianwali region) the Indus Piedmont has an elevation of about 240 to 270 m above the mean sea level and it decreases to about 120 m in the mid-western region near Sibi and to about 60 m in the southwestern region near Dadu.

The Piedmont Plain is characterised by gravel and sand deposits, braided stream pattern and bad-land topography. The Subpiedmont Plain contains fine textured soils, and is traversed by numerous small, meandering, dry channels, which have their origin in the Piedmont Plain. Downstream the Subpiedmont Plain merges with the Indus Flood Plain.

Associated with the Indus Piedmont there are three smaller, complex plains, the Daska Plain (between the Ravi and Chenab), Gujrat Plain (between the Chenab and Jhelum), and the Kacchi Plain (between Bugti Hills and Kirthar Range). The Daska and Gujrat Plains are extensions of the Subpiedmont Plain, but differ from the latter in having a gentler slope. The Kachhi Plain is essentially a complex alluvial fan formed by the accumulation of sediments brought down by the Beji, Chakar, Nari, Bolan and the Mula Rivers. These streams become dry as they reach the Kachhi Plain. This plain comprises two components, a relatively low-lying area subject to frequent floods, and the intervening higher ground which is free from floods. This higher ground is similar to the Subpiedmont Plain, and may represent an old Subpiedmont surface into which the braided flood plain has entrenched itself. The Kachhi Plain, has a slope of about 1 to 2 m/km and southward it merges with the Central Alluvial Plain.

The Central Alluvial Plain

The Central Alluvial Plain occupies the region between the Indus Piedmont and the Thar Desert (Fig. 2.19 and Photo. 1). It has a gentle southward slope ranging from about 1.0– 0.2 m/km in the north to 0.1 m/km or less in the southern part near the delta. Its elevation in the north is about 180 m to 210 m, in its central part near Sukkur it is about 54 m and near the delta only 3 m to 4 m above the sea level. It comprises the present flood plains of the Indus and its tributaries, their former courses and courses of the extinct streams, the Barr plains and the present as well as the old deltaic flood plains and associated tidal mud flats. The present flood plains and delta are geomorphologically active features, where processes of degradation, planation or aggradation are in progress. Most of the others are fossilised features or relics which, nevertheless, bear silent testimony to the frequent occurrence of the same geomorphic processes in the past.

The present flood plains of the Indus and its tributaries

The present flood plains of the Indus and its tributaries may be defined as low lying tracts of land along the rivers, frequently inundated during high floods. They are flat and have a relatively even surface. They bear scars of old river meanders, numerous oxbows, abandoned-channel fillings, and a number of small misfit streams, which are relics of the former courses of the present rivers. Laterally the flood plains terminate against low terraces, alluvial ridges or natural levees. Their width ranges from about 16 to 50 km.

The present streams are moderately entrenched in the flood plains and are contained in well defined, relatively shallow, 6 to 25 km wide meander belts. These belts occupy the lowest part of the flood plain and are separated from the latter by low, vertical banks or alluvial terraces. They are characterised by river meanders, cut-offs, oxbow lakes, abandoned channels, numerous point bars, swales and sand ridges. The stream bed itself is usually in the form of a narrow meandering depression, bounded by vertical banks two to five feet high. In exceptionally high floods, the water overflows even the flood plain zone and spills over the Barr lands or the old abandoned flood plain zone. The sediments being deposited on the present flood plains may be divided into flood plain deposits, meander belt deposits and stream bed deposits (Fig. 2.19).

The Barr Plains

The interfluves between the Sutlej, Ravi, Chenab and Jhelum form wide flat plains– the Barr Plains (Kazmi 1966, 1977). They are terminated towards the rivers by two sharply defined alluvial terraces which divide the Barr Plains into an upper and a lower zone. The Upper Plain is confined to the region upstream of the confluence of the five rivers. It constitutes the flattest, the most regular and even-surfaced feature on the Indus Plain. It bears scars of old and extinct stream channels. To the northeast it merges with the outwash plain of the Chaj Doab, the Subpiedmont Plain of the Bari Doab and the flood plains of extinct streams in the Rechna Doab (Fig. 2.19).

The terrace forming the Lower Barr Plain may be traced along the western margin of the Thar Desert as far south as Nawabshah. In this region it bears a relatively narrow terrace plain, one to three metres above the present and earlier flood plains of the Indus.

Flood plains and meander belts of extinct streams

Beyond the present flood plains, the remaining portion of the Central Alluvial Plain bears abundant traces of old meander belts which are traces of ancient courses of the present day rivers (discussed earlier in this Chapter). They provide significant information for deciphering the geological history inasmuch as each major shift in the course of the main streams, particularly when preceded by degradation, has been in response to definite climatic or geological changes. Abrupt change in the course of a river, therefore, indicates a definite time interval, in the same way as glacial advance, change in sea level, uplift, or terrace formation. Thus in Pakistan much of the aggradational history of the Indus and its tributaries, during the Late Quaternary Period, may be interpreted with the help of these old and abandoned stream channels (see Chapter 5).

The Indus Delta

The present delta of the Indus is typically triangular in shape and extends from Thatta up to the sea (Figs. 2.19 and 2.20). It covers an area of approximately 2,600 km^2. Before the construction of the canal irrigation system, the delta was advancing about 113 feet per year towards the sea (Pascoe 1964) and the Indus was transporting nearly 300 million tons or approximately 80,000 acre feet of silt to the delta. Since the construction of the canals and barrages across the Indus and its tributaries, the annual mean discharge of the Indus into the sea has been greatly reduced and the active delta has shrunk to a small 260 km^2 triangular zone, near Keti Bander.

The Indus Delta is one of the largest in the world. It ranks seventh in area, one step ahead of Mississippi, twelfth largest in its drainage area, seventh largest in river discharge and tenth largest in the length of its shore line (Figs. 2.21 and 2.22). The Indus Delta has a relatively straight coast line. With a ratio of less than two "shoreline kilometers" for each one kilometer of straight coastline, it stands in contrast to highly crenulate deltas such as the Mississippi and the Ganges. It receives the highest average wave energy of any major delta in the world at a water depth of 10 m (Fig. 2.22). At the shoreline the Indus Delta receives more wave energy in one day than the Mississippi Delta gets in one year and this wave energy is fourth highest as compared with other deltas (Wells et al. 1984).

The sediments discharged by the Indus settle on the continental shelf but some are also transported to the southeast by longshore currents. Much of the coarse sediment load is directly funneled through the Indus submarine canyon (Islam 1959), to the Indus Fan which is the largest physiographic feature in the Arabian Sea. Tongues of thick, bottom-water turbidites in the Indus Fan support the concept of deep turbid-layer flows (Kolla et al. 1981, Coumes et al. 1984).

The present day deltaic complex consists of the deltaic flood plain and the tidal delta. The deltaic flood plain is characterised by meander belt deposits of the distributaries

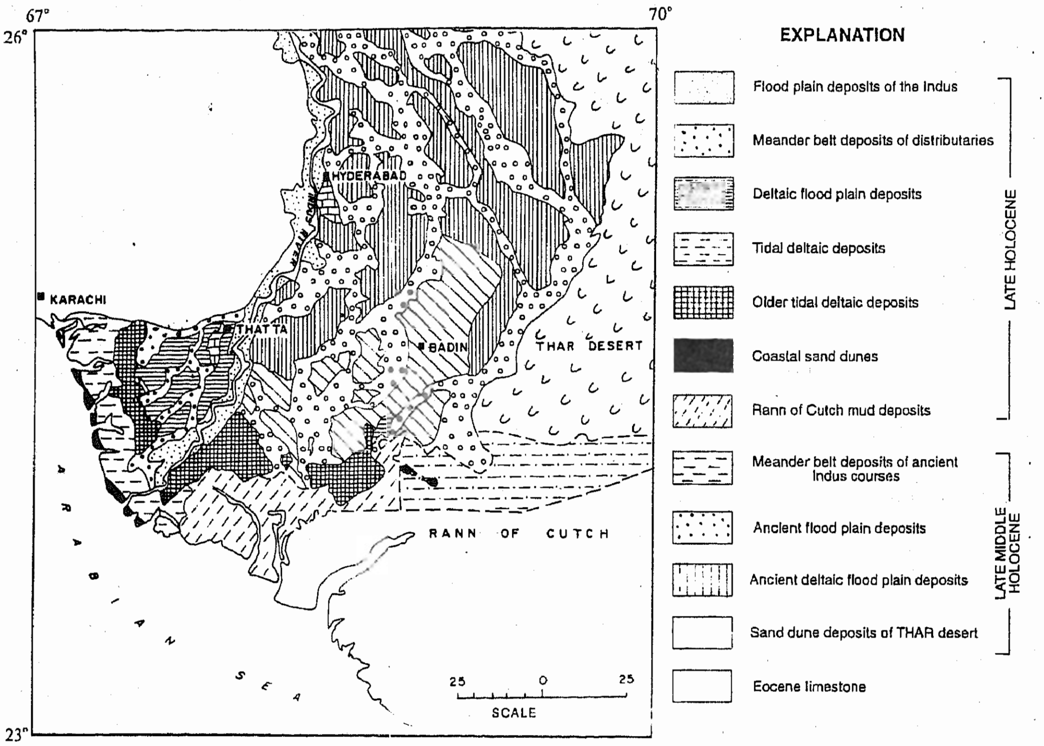


Fig. 2.20. Geological map of Indus Delta. (From Kazmi 1984).

of the Indus which have silted up since the confinement of the Indus within protection levees. In between the meander belts of the distributing channels, and located at slightly higher elevation, are the deltaic flood plains formed by the overflow from these distributaries (Fig. 2.20).

The lower margin of deltaic flood plain is rimmed by the tidal delta. Bordering the deltaic flood plain there is an arcuate zone of older tidal deposits, formed as a result of silting up of distributary channels. With the drying and silting up of major distributaries of the Indus, the tidal delta has been cut off from the alluvial process and has assumed the form of a tidal mud flat. Lower remnants of distributaries have turned into tidal creeks. Immunity from floods has caused growth of regular sandy beaches and a narrow zone of coastal sand dunes, some of which form small islands at the mouths of the major creeks.

The deltaic deposits mainly comprise interlayered deposits of very fine sand, silt and clay. Pits or shallow wells dug on the deltaic flood plain show the laminated nature of the sediments, which is unlike the common non-bedded, massive or cross bedded character of the flood plain deposits upstream of the delta. Deposits of silt and clay containing abundant mollusc shells underlie the present deltaic flood plain. Whereas the deltaic sediments are significantly finer than the flood plain deposits of the Indus upstream of the delta, they resemble the silty continental shelf deposits at the mouth of the delta (Fig. 2.23). The shelf deposits however are fine (mainly silt) and contain marine microfossils (mainly foraminifera).

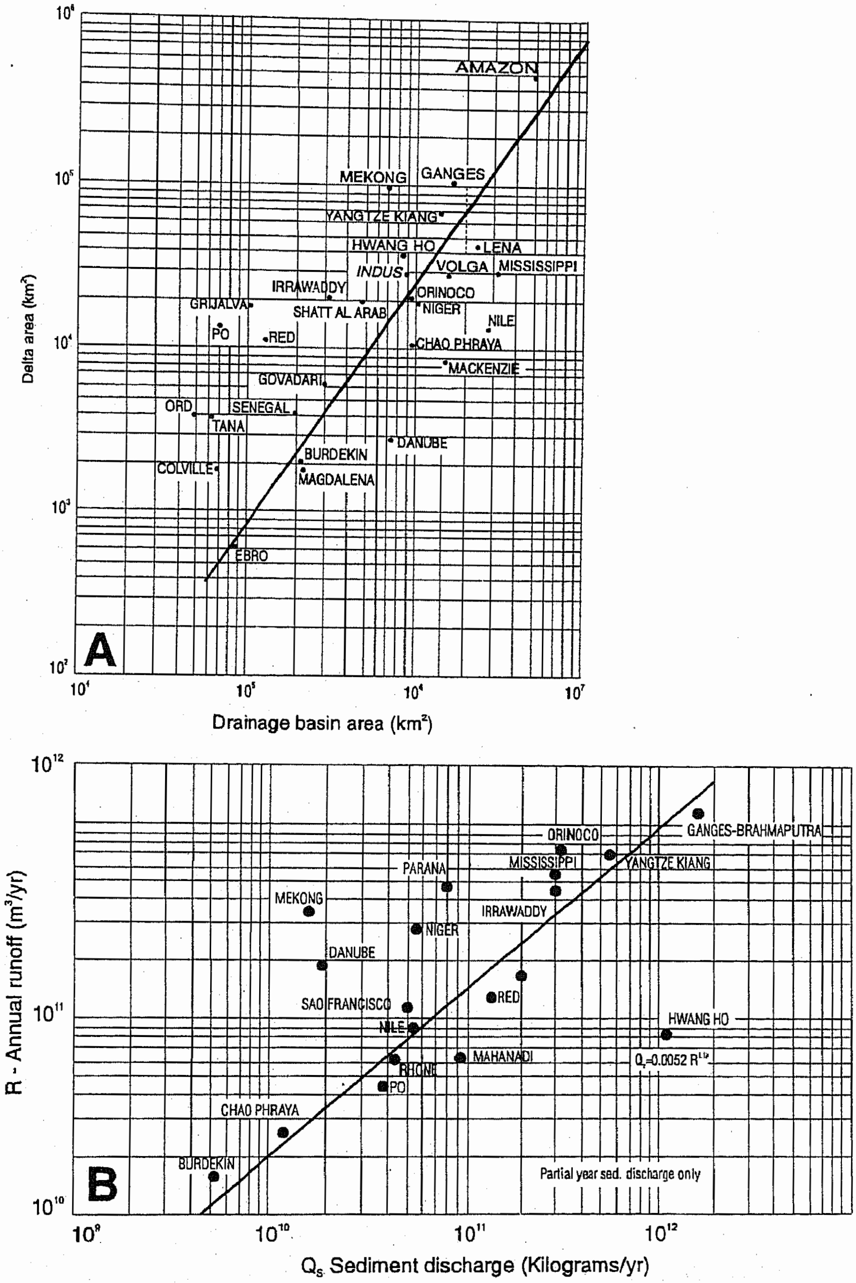


Fig.2.21. Area, runoff and sediment discharge of major deltas of the world. **A**–Plot of delta plain area versus drainage basin area. **B**–Plot of annual runoff versus sediment discharge for major rivers of the world. (From Wells and Coleman 1984).

Ancient deltaic deposits, similar to those of the present day delta, occur farther eastward, suggesting a westward shift in the position of the Indus Delta.

THE THAR DESERT

The Thar Desert is one of the significant geologic and geomorphic features of Pakistan and is located along its eastern borders (Photo. 1, Fig. 2.2). It extends beyond the international border into India. In Pakistan it covers an area of approximately 75,000 Km^2. It is mainly comprised of longitudinal, transverse, and barchan type of sand dunes as well as more complex dune forms, which are probably transitional between these three main forms. Based on the dune morphology, their relative stability and dimensions, the desert may be divided into four main zones (Fig. 2.19) which from south to north are (a) zone of stabilised longitudinal dunes, (b) zone of barchan dunes, (c) zone of transverse sand ridges and mixed dunes, and (d) the desert fringe zone (Kazmi 1977, 1985).

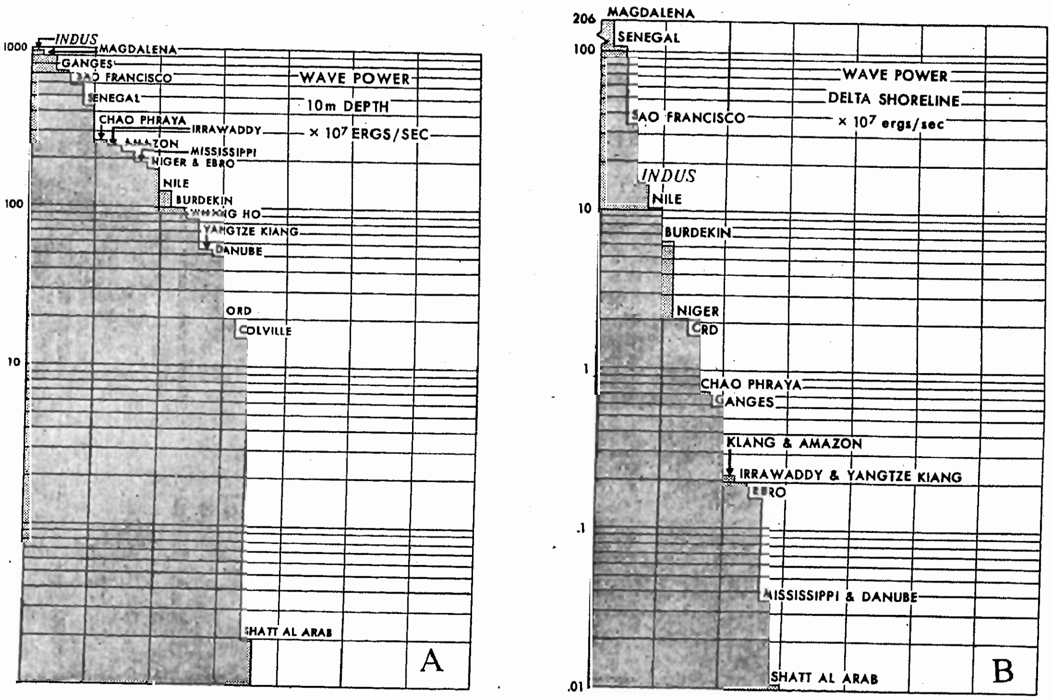


Fig. 2.22. Wave power of the Indus and other major deltas of the world. **A**–Bar graph of wave power per unit crest width at the 10-m contour. **B**–Bar graph of wave power per unit crest width at the delta shoreline. (From Wells and Coleman 1984).

Zone of stabilised longitudinal dunes

This zone covers the southern part of the desert and contains large longitudinal, northeast trending sand dune ridges 6 to 10 km long. These ridges occur in en echelon fashion and overlap or coalesce to form larger ridges as much as 32 km long. They are relatively narrow and barely exceed a width of about 200 to 250 m. They have steep slopes and the relief varies from about 20 to 100 m. Between the ridges there are small, narrow, elongated, flat depressions or basins covered by thin loamy sandy soil. In the southern part the interdunal depressions are filled with playa type sediments, salt deposits, or contain saline lakes and marshes.

Zone of barchan dunes

Along the northeastern margin of the zone of longitudinal dunes, there is a relatively small zone of active and moving barchans, (Fig. 2.19). Swarms of barchan dunes, with their pointed ends directed northeastward, are arranged along narrow sand ridges. Their relief varies from 3 m to 15 m and the larger ones are as much as 30 m to 65 m across. Southwestwards, these moving dunes are directly in contact with the longitudinal dunes which appear to be stabilized.

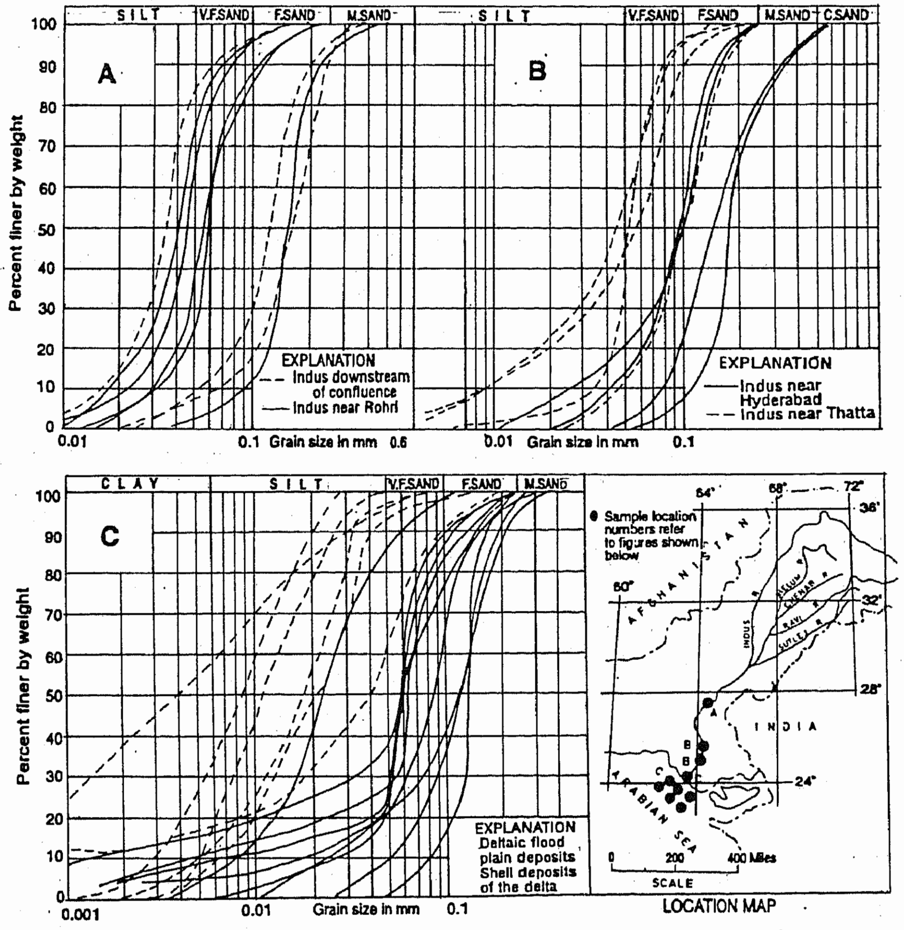


Fig. 2.23. Cumulative curves showing the grainsize frequency distribution in the Indus flood plain, deltaic flood plain and shelf deposits off the Indus Delta (from Kazmi 1984).

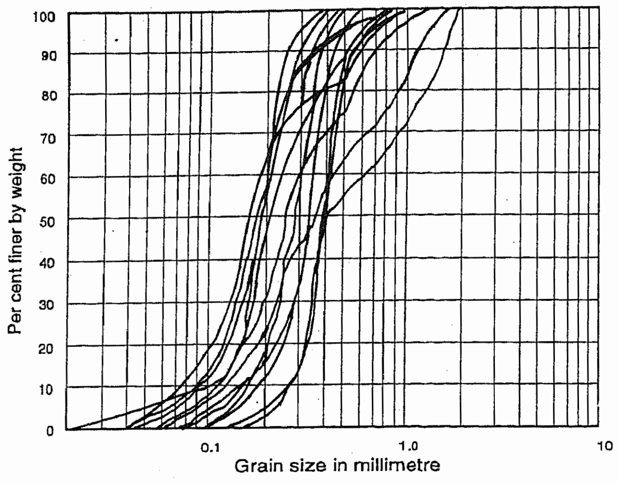


Fig. 2.24. Cumulative curves showing the grain size frequency distribution in the aeolian sand deposits of the Thar Desert (from Kazmi 1966).

Zone of transverse sand ridges

The zone of transverse ridges and mixed dunes extends from near Nawabshah, northeastward up to Bahawalpur. It is characterised by parallel ridges of transverse sand dunes. The ridges frequently have a beaded or serrate appearance. They have a long gentle windward slope and a narrow near-vertical leeward slope. On their leeward side there are small depressions occupied by mud flats. The ridges are about 3 to 6 km long and 3 to 16 m high. Most of the dunes are stabilized or in various stages of stabilisation. The continuity of the ridges is frequently broken by influx of moving sand, in the form of barchan dunes or, at some places, in the form of small longitudinal dunes. Northeastward the dunes become more complex.

The desert fringe

The desert fringe occupies a large area between the Thar Desert, and the Central Alluvial Plain. It is a part of the flood plain which has been invaded by the desert in relatively recent times. Near Fort Abbas there are remnants of the old channels of the Hakra River, which according to historical records flowed through the region as late as the seventeenth and eighteenth centuries. Several old channels of this extinct river have been mapped (Kazmi 1966) which indicate that before desertification this region was drained by a well developed river system (Fig. 2.17).

In this region the sand dunes occur in intermittent patches and in between there are large irregularly shaped mud flats. As a result of modern cultivation the supply of sand from the ploughed fields has triggered off the process of desertification and once again fresh barchan dunes have begun to form and move into this region.

Physical properties of the aeolian sands

The sand in the longitudinal dunes is commonly well sorted and the average coefficient of sorting is about 1.1. It is a fine to medium well rounded sand, with less than 5 per cent silt and an average median grain size of about 0.21 mm (Fig. 2.24). Some of the sand is bimodal and contains more than 50 per cent coarse sand grains. The coarse fragments are up to 2 mm in diameter. The desert sand contains less mica (less than 5 per cent as compared with 5 to 30 per cent in alluvial sediments), more quartz (80 per cent or more as compared with 50 to 75 per cent in alluvial sediments) and relatively smaller quantities of dark minerals. The sand in the barchan and transverse dunes is relatively finer and more micaceous than the sand in the longitudinal dunes. The former has been derived from the flood plain sediments.

THE THAL DESERT

A relatively small desert covers the interfluve between the Indus and Jhelum Rivers, south of the Salt Range. It is about 275 km long and its maximum width is about 100 km (Fig. 2.25). It is largely comprised of low longitudinal sand ridges, transverse ridges, barchan dunes and ridgy alveolar sands (Higgins et al. 1974).

THE COAST AND THE OFFSHORE

The Seacoast

The Pakistan shoreline stretches east to west for nearly 700 km. East of Karachi, 200 km of the crenulated deltaic coast is gullied by numerous tidal creeks. Fringing the subaerial delta there is the narrow 10–15 km wide sub-aqueous delta. Near Karachi, the coastline bends sharply northward and after a short distance forms a 60 km loop around the Sonmiani Bay (Fig. 2.2). At the head of the Bay there is the Miani Hor Lagoon. West of Sonmiani Bay the coast line is fairly regular, though it is slightly indented at Ras Malan (Photo. 14),

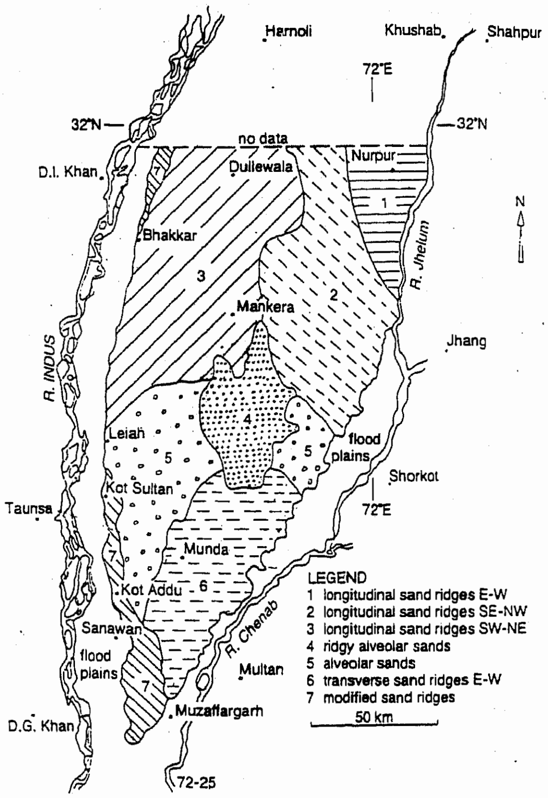


Fig. 2.25. Generalised map of sand ridge landforms in the Thal Desert interfluvial area between the Indus and the Jhelum and Chenab Rivers (modified from Higgins et al. 1974).

followed by a hammer-head shaped projection at Ormara (Ras Ormara). Near Pasni it forms a small loop ending in the Ras Jaddi cape and encloses the Pasni Bay. Farther to the west there are more hammer-head shaped projections of the coast near Gwadar (Ras Nuh, Ras Pishukan, Ras Jiwani and Ras Fastah), which enclose the Gwadar East Bay, the Gwadar West Bay and the Gwatar Bay. In this region of Makran there is an elevated 16 to 32 km wide coastal plain. The Makran coast is comprised of a 15 m to 65 m high rocky cliff standing above a very narrow beach. Above the cliff there is a 16–32 km wide coastal plain which is dotted with several small hills and ridges and is extensively covered with sand dunes. Numerous small, ephemeral streams drain the coastal ranges and traverse this plain. They drain into the sea through short narrow creeks.

Uplifted coastal terraces

A number of isolated terraces, characterised by wave-cut platforms, occur on Makran coast between Karachi and Jiwani (Fig. 2.26). Some of these terraces rise 260 m above the coastal plain (HSC 1960). They are largely comprised of calcareous Chatti mudstone and covered with beach and near-shore sediments including coastal dune deposits. The sediments capping the terraces have been deposited on wave cut platforms commonly with shelly sandstone and conglomerate at base and the sequence contains abundant remains of corals, clams, oysters and other wave resistant forms. Some of the terraces, capped by these Quaternary deposits, stand as much as 500 m above sea level (Haq and Milliman 1984). This indicates the extent of recent uplift in the coastal area. Some of the raised-terraces of the Makran coast have been described in Table 2.7.

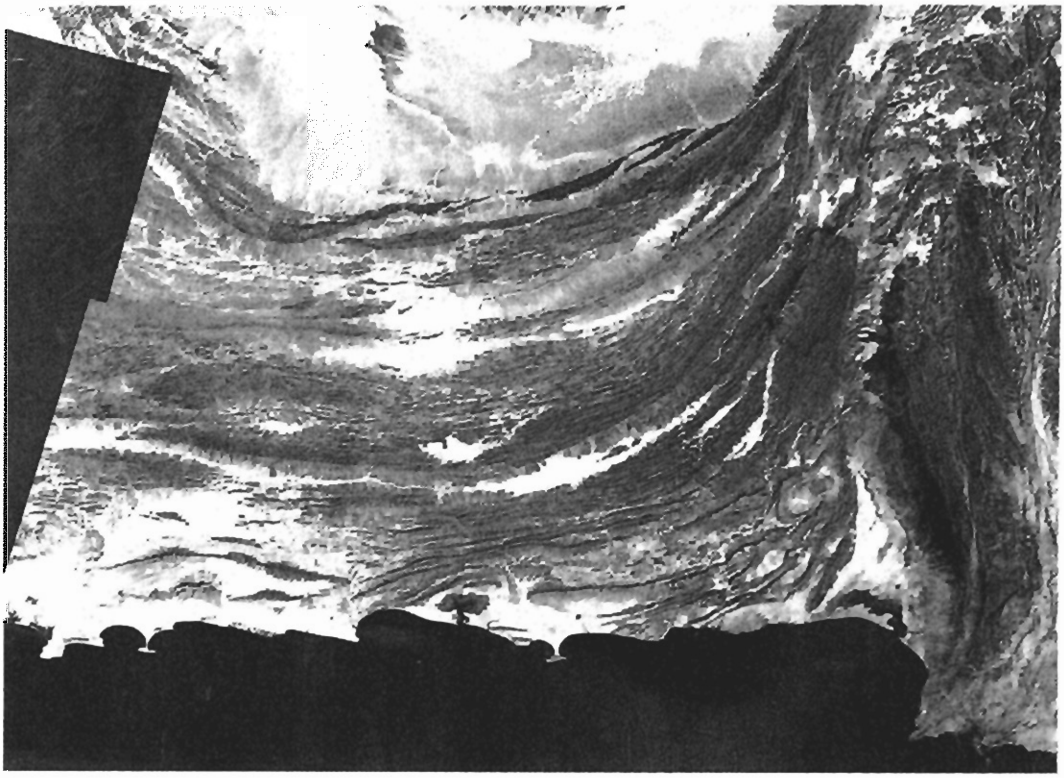
Besides the uplifted marine terraces, the following three types of terraces can also be seen on the Makran coast :-

1. *Structural terraces:* These are flat uplands, capped or protected by hard resistant beds and formed due to erosion.
2. *Stream terraces:* A sequence of four fluvial terraces occurs above the valley floors of the present streams.
3. *Submarine terraces:* There are offshore, submerged terraces resembling the coastal terraces and they are believed to have formed by either-faulting (HSC 1960) or lower sea levels (Stiffe 1873).

The uplifted marine-terraces have been dated by ^{14}C from shell material (Table 2.8). According to Snead (1992) a transgression occurred 30,000–25,000 year BP, when the sea stood approximately at today's level. It was followed by a regression that reached its lowest level (at least 90 m lower than present sea level) about 20,000–17,000 year BP (Hoyt et al. 1968, Bull 1984). Progressive development of the marine terraces has been shown diagrammatically in Figure 2.27.

Oman Abyssal Plain and submarine ridges

The sea shelf is about 130 km wide off the Indus Delta but along the Makran coast it narrows down to 20 km to 50 km. The sea slope is about 250 km wide off the delta and 60 km to 90 km wide in the Makran region. The sea-floor is characterised by a number of physiographic features which are significant from morphotectonic or plate tectonic stand point. South of Makran coast there is the triangular, nearly 350 km long, 120 km wide and over 3,200 m deep, Gulf of Oman Abyssal Plain. The Oman promontory of the Arabian Shield



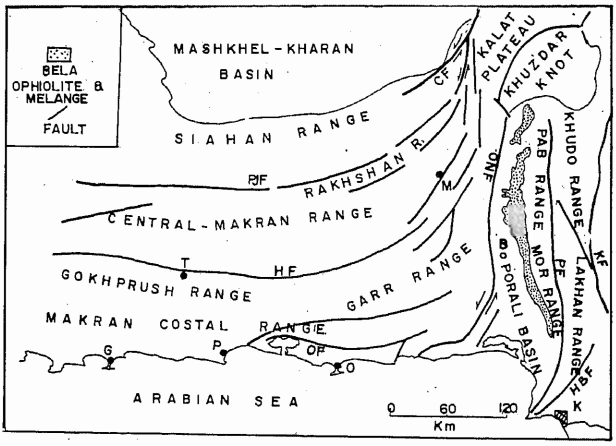


Photo. 14. Landsat photo mosaic showing the Makran Ranges, the coastal region and the Bela ophiolite and thrust belt. (Courtes U. S. Geol. Survey). See Fig. 4.43. B–Bela, CF–Chaman Fault, G–Gwadar, HBE–Hab Fault, HF–Hoshab Fault, K–Karachi, KF–Kirthar Fault, M–Mashkai, O–Ormara, OF–Ormara Fault, ONF–Ornach-Nal Fault, P–Panjgur, PF–Pab Fault, PJF–Panjgur Fault, T–Turbat.

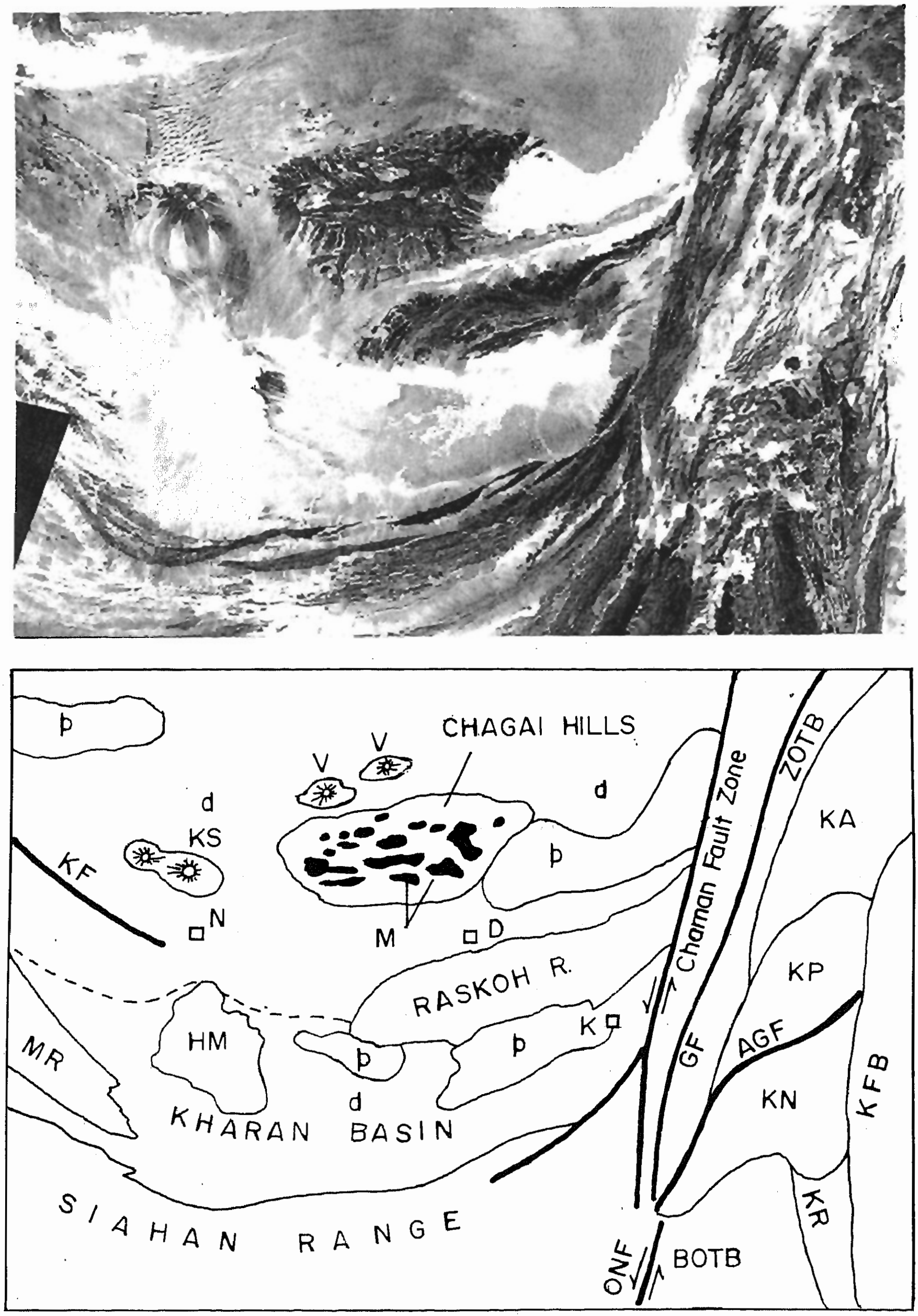


Photo. 15. Landsat photo mosaic showing the chagai-Ras Koh Ranges and adjacent areas. (Courtesy U. S. Geol. Survey). AGF-Anjira Ghazan Fault, BOTB-Bela ophiolite and thrust belt, D-Dalbandin, d-sand dunes, GF-Ghazaband Fault, HM-Hamun-i-Mashkhel, K-Kharan, KA-Kalat anticlinorium, KF-Kirtaka Fault, KFB-Kirthar fold belt, KN-Khuzdar Knot, KP-Kalat Plateau, KR-Khude Range fold belt, KS-Koh-i-Sultan, MR-Mirjawa, Range, N-Nok Kundi, ONF-Ornach-Nal Fault, p-playa, V-volcano. ZOTB-Zhob ophiolitic Thrust Belt.

Table 2.7. Characteristics of Makran coastal terraces and platforms (modified from Snead 1992).

	Location	Elevation (m)	Elevation of wave-cut cliffs (m)	Age	Characteristics
1.	Clifton Hills	29	3-6, 0.8 km inland	Pliocene	Wave cut scarps, sea caves, bedrock conglomerate and sandstone.
2.	Manora Rocky Headland	6	9.14, on present beach	Pliocene	Flat, wave-cut, small caves and sea arch, capped by conglomerate.
3.	Oyster Rocks	2	9.1	Pliocene	Small rocky islands 1.6 km offshore Karachi harbour.
4.	Hawks Bay Cape Monze	12-15	6-9	Pliocene-Pleistocene	Dip 3° seaward (E), comprised of sst, lst and shale, oyster beds at 2.4-3.0m, trace of uplifted marine terraces at 15- 46m.
5.	Ras Malan	611	91-152 (East) 183-244 (West)	Pliocene-Pleistocene	Perched rocky platforms, gentle westward dip.
6.	Ormara	474	Inner cliffs 427; seaward side 183-305	Pliocene-Pleistocene	Dip 3-5° seaward(S), comprises a horst with mst, sst, cgl and shelly lst, connected by large tombolos to the mainland.
7.	Ras Sakanni Kamagar Hills	318	12-15	Pliocene-Pleistocene	Rocky, flat, marine platform, comprised of mst & sst.
8.	Ras Zarain Ras Jaddi	127	91	Miocene-Pleistocene	Uplifted bedrock platform, comprised of mst. cgl & shelly lst.
9.	Ras Shamal Bandar Poin	195	140-183	Miocene	Marine platform dipping to SW, comprises mst & sst.
10.	Jabal-i-Mehdi Platform	410	152-183	Pliocene-Pleistocene	Marine platform, 5-15° SW dip, with mst, cgl and shelly lst.
11.	Gwadar	145	Inner cliffs 107-22m; outer cliffs 85-21m	Pliocene-Pleistocene	Flat, rocky, marine platform; 3° SW dip; connected to mainland by tombolo.
12.	Ras Pishukan	15-18	7.6	Pleistocene	Rocky headlands comprising two uplifted platforms. composed of cgl and shelly lst.
13.	Jiwani	135	26-30	Pliocene-Pleistocene	Rocky headland formed by Uplifted, tilted marine platform, 5° W-NW dip, comprised of mst,sst,cgl and shelly lst; eastern side deeply dissected.

Table 2.8. Location, height and age of some of the uplifted coastal terraces (modified from Snead 1992).

Location	Elevation (m)	14C Age (yr BD)
Gavatar (Iran)	30	5,000
E. of Chah Bahar (Iran)	15	15,000
Gwadar	45	20,000
Sonmiani	?	25,000
Gwadar	75	30,000

forms its western margin. Eastward it is bounded by the submarine Murray Ridge (Fig. 2.28). This is a slightly arcuate ridge which extends for about 750 km southwestward from near Karachi. Its relief varies from 2,000 m to 3,500 m. It consists of a series of

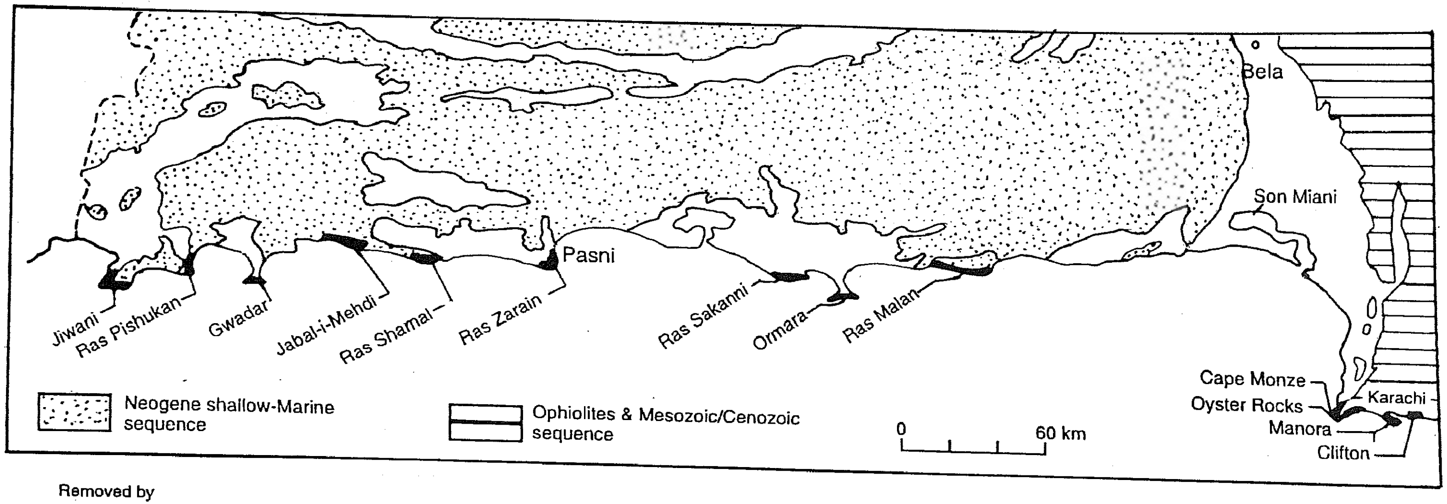


Fig. 2.26. Map showing terraces along the Pakistan coast (from Snead 1992).

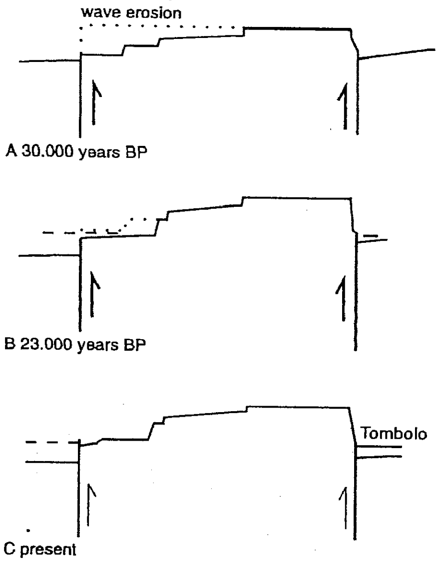


Fig. 2.27. Progressive development of marine terraces (from Little 1972).

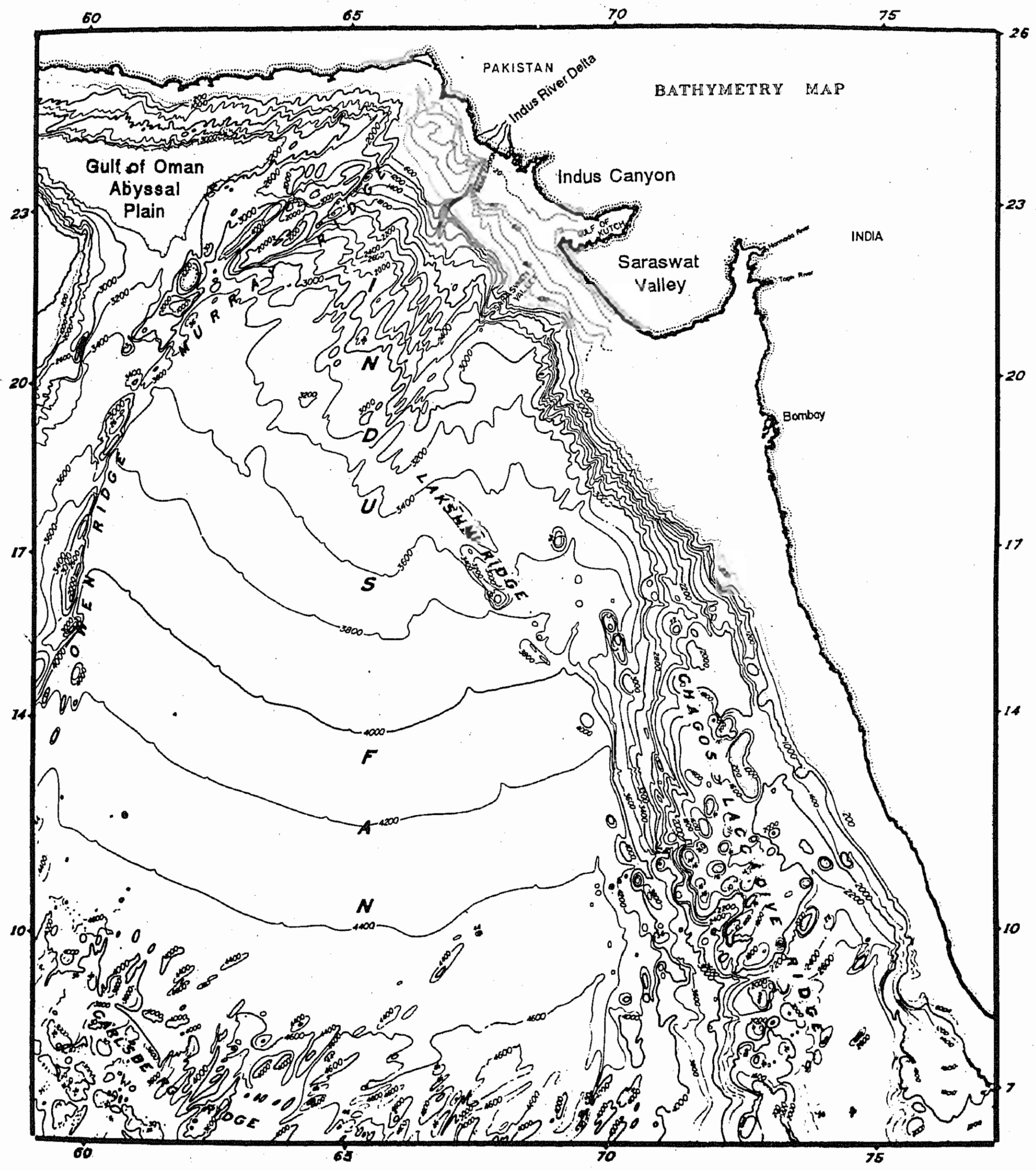


Fig.2.28. Bathymetric map of Arabian Sea showing offshore features and the Indus Fan (from Coumes and Kolla 1984).

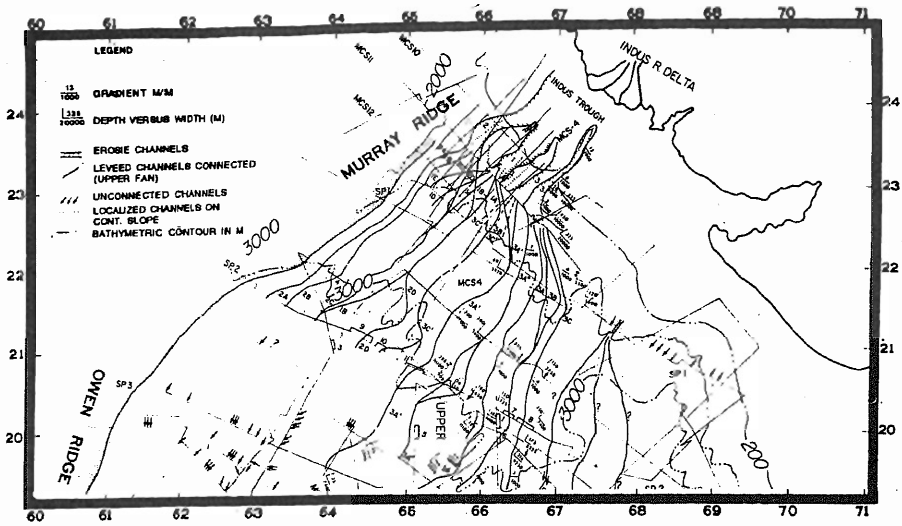


Fig. 2.29. Map showing patterns of submarine canyons (1, 2 and 3) and channels (1A, 1B.., 2A, 2B..., 3A, 3B...etc) on the Indus fan and shelf (from Coumes and Kolla 1984).

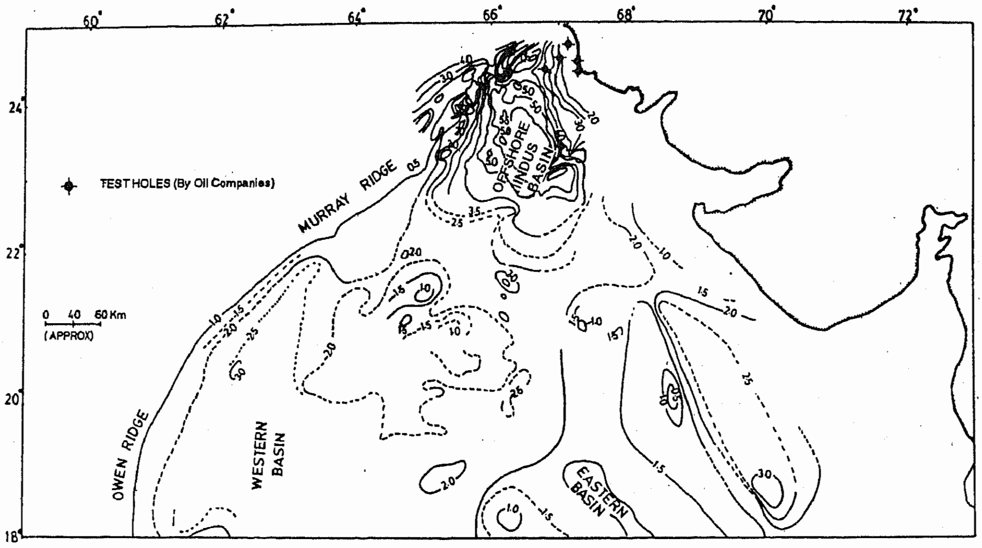


Fig. 2.30. Sediment-thickness distribution in seconds of two-way travel time. Note three major depocentres: Offshore Indus Basin, Western Basin and Eastern Basin (from Coumes and Kolla 1984).

ridges, seamounts, and small basins arranged in a linear pattern. Southwards the Murray Ridge links up with the Owen Fracture Zone.

Indus Submarine Fan and the Indus Canyon

East of the Murray Ridge lies the vast Indus Submarine Fan which is indeed the most pronounced feature off the coast of Pakistan. After the Ganges Fan, it is the second largest submarine fan in the world. It extends 1,500 km southwards from near the Indus Delta up to the Carlsberg Ridge. The chain of Chagos–Laccadive–Laxmi ridges forms its eastern boundary (Fig. 2.28). The water depths of the fan range from 1,400–1,600 m in the north to 4,500 m near its southern margin along the Carlsberg Ridge. Off the Indus Delta, in the shelf–slope region of the Fan, there is the Indus Canyon. It is a spectacular offshore feature, 170 km long, about 8 km wide and commences from water depths of 20–30 m on the shelf and ends at 1,400 m depth at the foot of the continental slope. At its distal end the canyon is 20 km wide with a relief of about 325 m. According to Coumes et al. (1984) seismic surveys have revealed relicts of older submarine canyons west of the present one (Fig. 2.29). These canyons lack overbank-levee deposits and were erosional with respect to the surrounding strata when they were active. Headwards the canyons merge into one extensive erosional zone "the Indus Trough." Seaward they lead into many channels in the upper fan. Seismic surveys indicate that more channels exist in the subsurface. Three sedimentary basins separated by a ridge complex can be distinguished in the upper Indus Fan (Fig. 2.30). These are the offshore Indus Basin, the Eastern Basin and the Western Basin (Naini and Kolia 1982).

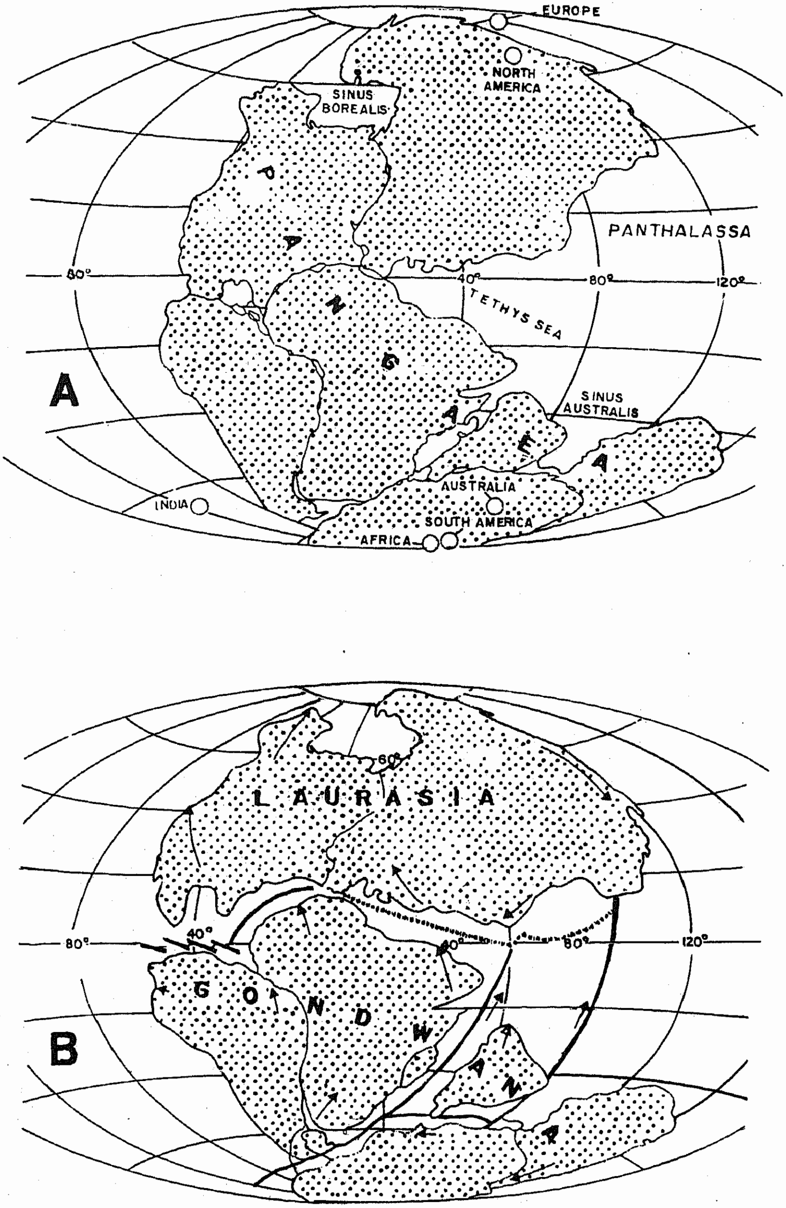


Fig. 3.1. Assembly and break up of Pangaea and Tethys (modified from Dietz et al. 1976).

A. An outline of Pangaea 200 million years ago. Panthalassa was the ancestral Pacific Ocean. The Tethys formed a large bay separating Africa and Eurasia. Circles show relative pole locations in Permian times.

B. Northern group of continents–Laurasia separated from the southern group–Gondwana after 20 million years of drift (i.e., 180 m.y. ago). India has been set free by rifting. Arrows indicate movement along plate margins.

3

Regional Geological Setting

A BRIEF REVIEW OF THE GEOLOGY OF THE REGIONS SURROUNDING PAKISTAN

Eurasia is comprised of three broad geological divisions which, from north to south, may be referred to as the Laurasian, Tethyan and Gondwanian domains. Their origin may be traced back to the Late Paleozoic, when all the continents had drifted to form a continuous land-mass, the supercontinent of Pangaea (Wegner 1924, Smith and Hallam 1970, Smith 1981, Irving 1979), which remained intact for about 100 Ma. It was surrounded by a universal ocean, the Panthalassa. An arm of this ocean, the Tethys (Suess 1893) formed a wedge between the northern and southern parts of Pangaea. By Late Triassic, Pangaea had split into two supercontinents, Laurasia to the north and Gondwanaland to the south (Fig. 3.1), separated by the Tethys seaway (Du Toit 1937).

It has been recognised now that Eurasia is comprised of many crustal blocks of various sizes sutured along ophiolite belts. Several of these blocks were accreted to the Laurasian continent since the Carboniferous (Burret 1974, Chang et al. 1973, Sengör 1984, 1988). Most of these are fragments of Gondwanaland, which detached from the mother continent, drifted northward and successively collided with the Laurasian Domain (Fig. 3.2). Present day Eurasia is thus comprised of the initial Laurasian landmass–the Laurasian Domain–and the accreted assemblages of the former fragments of Gondwanaland. Following Sengör (1988) we have herein designated this accreted terrain as the Tethyan Domain. Southward this domain is followed by the Arabian and Indian peninsular shields, both of which also have fragmented from the former Gondwanaland and are the last ones to be accreted to Eurasia. However they differ from the other Gondwanian blocks inasmuch as they bear extensive outcrops of Archean and Proterozoic metamorphic and igneous rocks. For want of a better name we here refer to them as the Gondwanian Domain. Pakistan is located at the junction of the Gondwanian and Tethyan domains. The geology of these two domains has been briefly reviewed in this chapter.

GONDWANIAN DOMAIN

This domain is characterised by a continental crust and crystalline basement consolidated in the Precambrian and a platform type development in the Paleozoic. In Asia it is represented largely by the Arabian and Indian shields. The latter forms the Indo-Pakistan subcontinent and its northern margin comprises the crystalline thrust sheets of the Himalayan orogenic belt. This belt continues east- and westward in the form of a series of contorted, oroclinal, marginal fold belts in the Arakans, the Sub-Himalayan Siwalik Range, the Salt Range and the Potwar Plateau, and the Sulaiman–Kirthar Ranges (Fig. 3.3).

Traditionally the Indo-Pakistan subcontinent has been divided into three principal physiographic and geologic divisions, namely (1) the Peninsular Region, (2) the Himalayan Foredeep which lies between the Himalayan mountain ranges and the Indian Peninsula and (3) the Himalayas (Wadia 1957). The Peninsular Region is comprised of

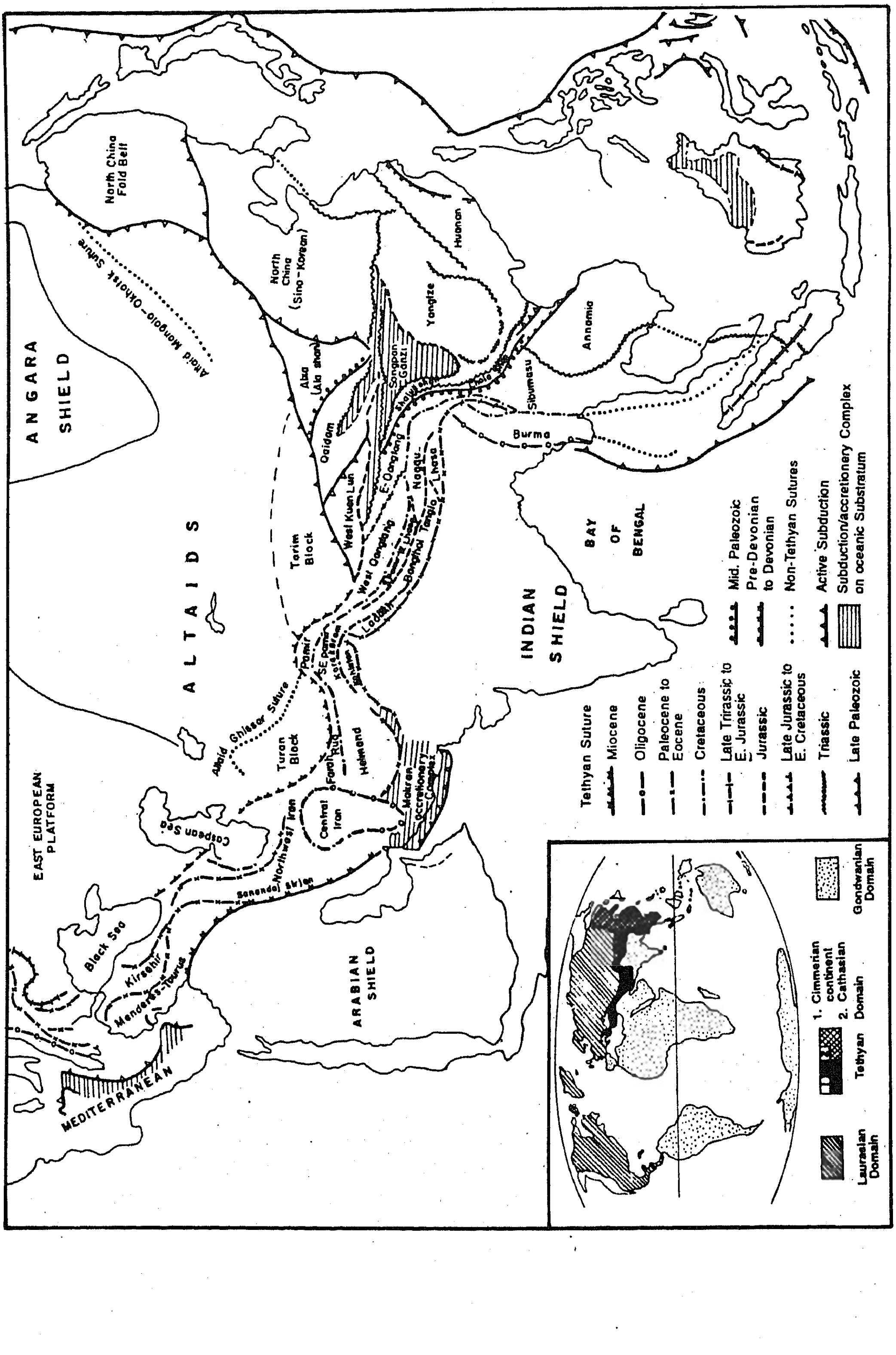


Fig. 3.2. Map showing broad tectonic zones of Eurasia and the major suture and accretionary blocks of the Tethyan Domain (modified from Sengör et al. 1988).

a series of metamorphosed plutonic, volcanic and sedimentary rocks ranging in age from Archean to Late Proterozoic. These Precambrian rocks crop out in the form of Archean cratons associated with supracrustal enclaves and surrounded by various Proterozoic mobile belts. Along the northern margin of the Peninsular Region, the wide plains of the Indus and the Ganges, underlain by Cenozoic to Mesozoic sedimentary cover, mask the Indian Peninsular Shield. Farther northwards, the Himalayas and its subsidiary ranges have been formed on the subducted margin of the Indo-Pakistan crustal plate.

The Peninsular Region

This region is mainly comprised of elements of the Indian Shield (Fig. 3.4) and it is characterised by :-

(1) Granite-greenstone terrains (Karnataka, Jeypore-Bastar, Singhbhum) comprised of Archean to Proterozoic magmatic rocks (Peninsular Gneisses) with enclaves of older supracrustal rocks (Gorur and Sargur Groups), Archean Banded Gneiss and granites (Bundelkhand).
(2) Late Archean reactivated basement rocks (Dharwar Supergroup) which envelop the older gneisses.
(3) Proterozoic mobile belts (Eastern Ghat, Satpura, Aravalli-Delhi) which are tectonically wrapped around the earlier Archean and Proterozoic rocks. Middle to Late Proterozoic granites intrude these belts.
(4) Middle Proterozoic rifting which has formed vast grabens filled with Late Proterozoic (Cuddapah) and Paleozoic to Mesozoic (Gondwana) sediments.
(5) Late Proterozoic sequence of thick and extensive, shallow-marine to deltaic deposits of coarse clastics and carbonates interspersed with volcanics (Vindhyan Supergroup).
(6) Late Cretaceous flood basalts (Deccan Trap) covering vast tracts of the Peninsular Region.

The Deccan Trap flood basalts cover about 500,000 km^2 area on the western flank of the Peninsular Shield (Fig. 3.4). Scattered outcrops of Deccan Trap-related lava flows also occur on the east coast of the Peninsula, in the Rajmahal Hills, Gujrat, Rajasthan, Sindh and Balochistan. These volcanic rocks attain a maximum thickness of about 3,000 km along the Bombay coast and become thinner towards east and north. They unconformably overlie the eroded surfaces of rocks ranging in age from Archean to Cretaceous. The individual flows range in thickness from a few metres to 30 metres and are commonly separated by thinner partings of ash and tuff. The beds are almost flat. The most common rock is a monotonous greyish-green, vesicular, augite basalt though at places there are variations in its colour, texure and composition. The petrography of the lava flows is fairly uniform, though the phenocrysts may vary from 0 to 50 modal percent between flows (West 1958, Najafi et al. 1981).

At places, the lava flows are locally separated by fossiliferous, lacustrine or fluviatile sediments, more commonly in the upper and lower traps. Plant remains, mollusc shells, insects, crustaceans and vertebrate fossils have been found in the intertrappean beds (Wadia 1957). In Gujrat and in the vicinity of Narbada–Son Lineament and parallel to it, the Deccan Traps have been extensively invaded by E-W trending dyke swarms.

Stratigraphic, geochemical and structural data supports the hypothesis that the Deccan

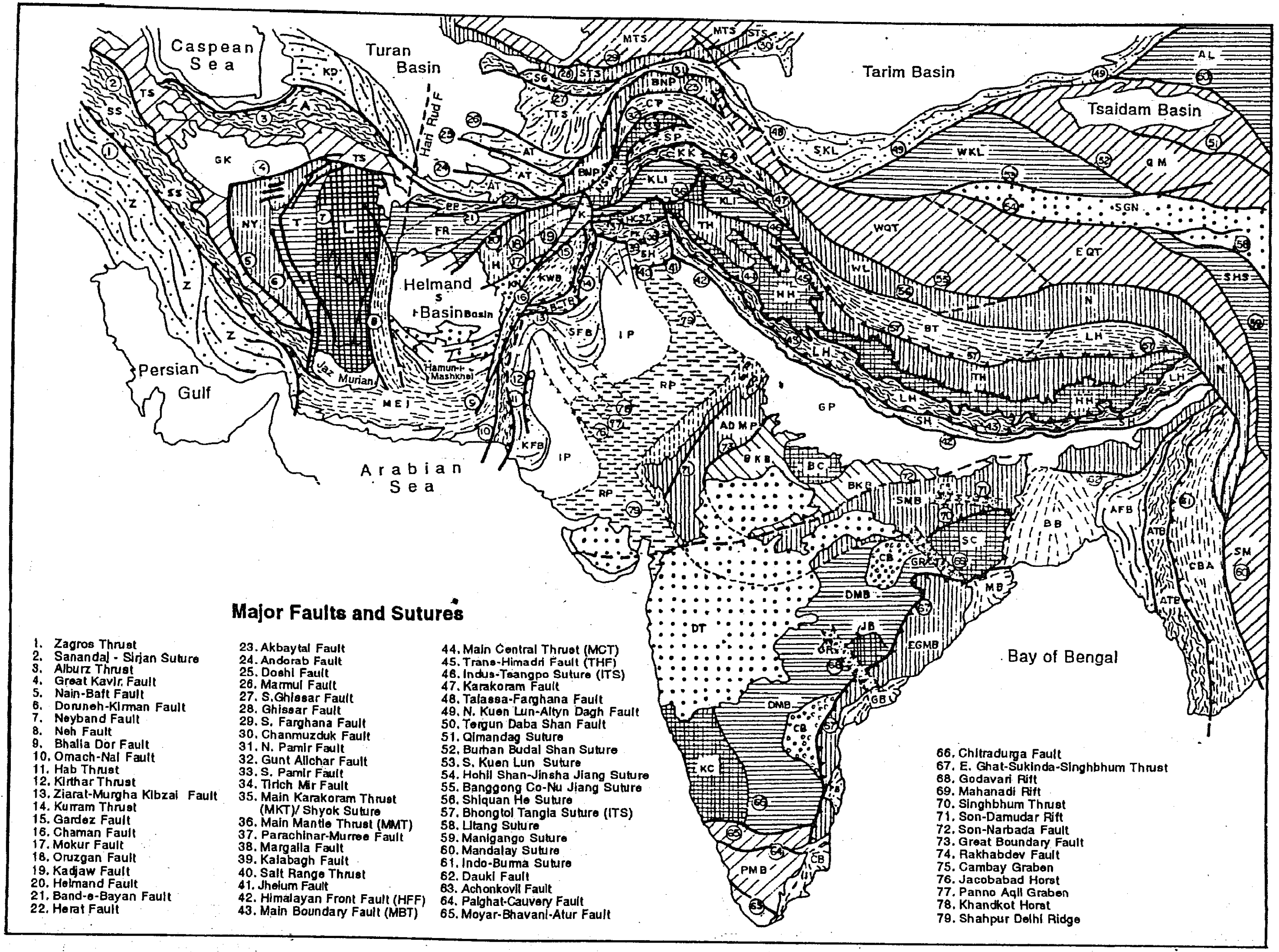


Fig. 3.3. Map showing tectonic setting of Pakistan and surrounding regions (from Gansser 1964, Stöcklin 1977, GSI 1977, Kravchenko 1979, Kazmi and Rana 1982, Drury et al. 1984, Radhakrishna 1986, Boulin 1988, Sengör et al. 1988, Sugden 1990, Ramakrishnan 1993).

(Fig. 3.3.)

EXPLANATION

GONDWANIAN DOMAIN

INDO-PAKISTAN SUB-CONTINENT

▦ KC- Karnataka Craton
Jb- Jeypore-Bastar Craton
Sc- Singhbhum Craton

▤ DMB- Dharwar Mobile Belt

▨ PMB- Pandiyan Mobile Belt

⁘ CB- Cuddaphah Basin

▥ EGMB- Eastern Ghat Mobile Belt
SMB- Satpura Mobile Belt
ADMB- Aravalli Delhi Mobile Belt

(Vertical dashed lines show covered areas of the belt)

▧ East Coast Sedimentary Basins
BB- Bengal Basin
MB- Mahanadi Basin
GB- Godavari Basin
PB- Palar Basin
CB- Cauvery Basin

▧ GR- Gondwana Rift Basins

▧ BKB- Bundelkhand Block

▦ BC- Bundelkhand Craton

▤ RP- Rajasthan Platform

⁘ DT- Deccan Trap

☐ Gp- Ganga Plain } Himalayan
IP- Indus Plain } Foredeep

▧ SH- Sub-Himalayan Fold Belt
SFB- Sulaiman Fold Belt
AFB- Arakan Fold Belt
KFB- Kirthar Fold Belt

≋ LH- Lesser Himalayan Thrust Belt
PK- Parachinar- Kalachitta Thrust Belt
BZTB-Bela-Zhob Thrust Belt
ATB- Arakan Thrust Belt

▦ HH- High Himalayan Crystalline Schuppen Zone

▥ TH- Tethyan Himalaya Fold and Thrust Belt

ARABIAN PLATFORM

▧ Z- Zagros Fold and Thrust Belt

TETHYAN DOMAIN

CATHAYSIA-TRANSHIMALAYA

(Tethyside microcontinental blocks of Gondwanic affinity)

▤ AL- Alxa
WKL- West Kuen Lun
SHS- Shashuli Shan

▨ QM- Qaidam
WQT- West Qangtang
EQT- East Qangtang
SM- Sibumasu

▥ WL- West Lhasa
N- Naggu

▤ BT- Bhangtol Tangla
LH- Lhasa
CBA- Central Burma

PAMIR-KARAKORAM

≋ CP- Central Pamir
SP- Southern Pamir

▦ NSWP- Nuristan- SW Pamir

≋ KK- Karakoram Fold and Thrust Belt

▤ KLI- Kohistan-Ladak Island Arc

PAKISTAN-AFGHANISTAN-IRAN

≋ BB- Band-e-Bayan Block
A- Alburz Fold Belt
SS- Sanandaj-Sirjan Microcontinental Block

▤ FR- Farah Rud Flysch Zone
T- Tabas Block
K- Kabul Block

▥ H- Helmand Magmatic Block
NY- Nain-Yazd Folded Zone

⁘ KN- Kandhar Volcanic Arc
CVA- Chagai Volcanic Arc

≋ MEI- Makran-East-Iran accretionary arc-trench system (Volcanic and Flysch Belt)
KWB- Katawaz Basin

▨ TS- Tabrez-Sabzevar Volcanic Belt

☐ GK- Great Kavir

LAURASIAN DOMAIN

CENTRAL ASIA-AFGHANISTAN-IRAN

▨ MTS- Central Tien Shan Fold Belt

▥ STS- South Tien Shan Fold and Thrust Belt
BNP- Badakhshan-N. Pamir horst and Thrust Belt

≋ SG- South Ghissar Suture Zone

▧ TTS- Trans-Tien Shan Fold Depression
AT- Afghan Turkestan Platform and Fold Belt
KD- Kopet Dagh Fold Belt
SKL- Sub-Kuen Lun, Tarim

▲ Thrust Fault

⊥ Rift

⊥ Horst

╱ Undifferentiated Fault

Trap Basalts formed with the northward drift of India over a hot spot (Morgan 1972, Devey et al. 1986). Subsequently the hot spot formed the Chagos–Laccadive Ridge (Fig. 4.4). Presently this hot spot is located at the volcanic Reunion Island in the Indian Ocean. Some authors have proposed that the flood basalts are the result of rising diapirs which become trapped at the crust-mantle boundary and feed the surface eruption (Richards et al. 1989), while others relate volcanic flows to an extra-terrestrial, cataclysmic impact event (Rampino et al. 1988).

The Himalayan Foredeep

This Foredeep occupies the gently sloping continental platform. Its basement is comprised of Archean and Proterozoic metamorphic and plutonic rocks. It contains a series of upwarps and intervening depressions and horst and graben structures which have been clearly defined by geophysical surveys (Fig. 8.5). The western part of the Foredeep is covered by the Indus sedimentary basin. This basin contains a thick prism of westward sloping Precambrian to Holocene sediments which peter out eastward. In the upwarp zones the bed-rock crops out or occurs at shallow depths (e.g., Shahpur–Delhi buried ridge, Jacobabad–Khairpur high, Khandkot–Mari high). Southward and westward the sedimentary cover thickens to more than 5,000 metres (Fig. 3.5).

The Gangetic Plain covers the northern part of the Foredeep. In this region the alluvium is up to 1,700 m thick and it is underlain by the Siwalik molasse. In East Punjab the Siwaliks unconformably overlie the Eocene to Early Miocene rocks. The latter cover the basement rocks. The Cenozoic sequence is up to 5,000 m thick and the Proterozoic Vindhyan sequence has been encountered at depths of 4,400 m (Shastri 1976). Eastward the Ganga Basin is mainly comprised of a thick pile of Siwalik molasse (900–4,000 m) which un-conformably covers the Pre-Tertiary sequences, mainly the Gondwanas or the Vindhyans (Shastri 1976).

The Himalayas

Bordering the Indus-Ganga Basins, the Himalayas form a 2,500 km long, 160–400 km wide mountain belt, comprised of a series of en echelon mountain ranges with extensive intervening valleys. This range forms sharp oroclinal loops or syntaxial bends near its eastern and western extremities, where its various off shoots form north-south trending, less elevated and not-so-spectacular mountain ranges (Figs. 3.3 and 3.4). The Himalayas have formed along the northern margin of the Indo-Pakistan crustal plate. This zone is covered by a nearly complete Phanerozoic sedimentary sequence up to the Eocene. Continental collision of India with Eurasia formed this vast mountain range with a thick pile of intensely deformed Phanerozoic sediments, interspersed with ripped up and thrusted masses of Proterozoic basement, intruded by a series of magmatic rocks and affected by various metamorphic events. The geology of the Himalayas is thus quite complicated.

Northward the Himalayan terrain terminates along the Indus-Tsangpo Suture Zone. Geophysical data (Molnar 1984, 1988) indicates that the continental crust in this region (High Himalayas and Tibetan Plateau) is between 50 and 80 km thick which is twice the normal crustal thickness. This feature has been attributed to the continued under-thrusting and underplating of India beneath Asia. To explain this anomaly Power et al. (1973) and Le Fort (1975) have proposed different models as shown in Figure 3.6.

According to Powell et al. (1973) the subducted mass of the Indian Shield has broken into two slices or blocks. The frontal block has collided with the Asian plate along the Indus-Tsangpo Suture, whereas the rear block of the Indian plate has continued to be

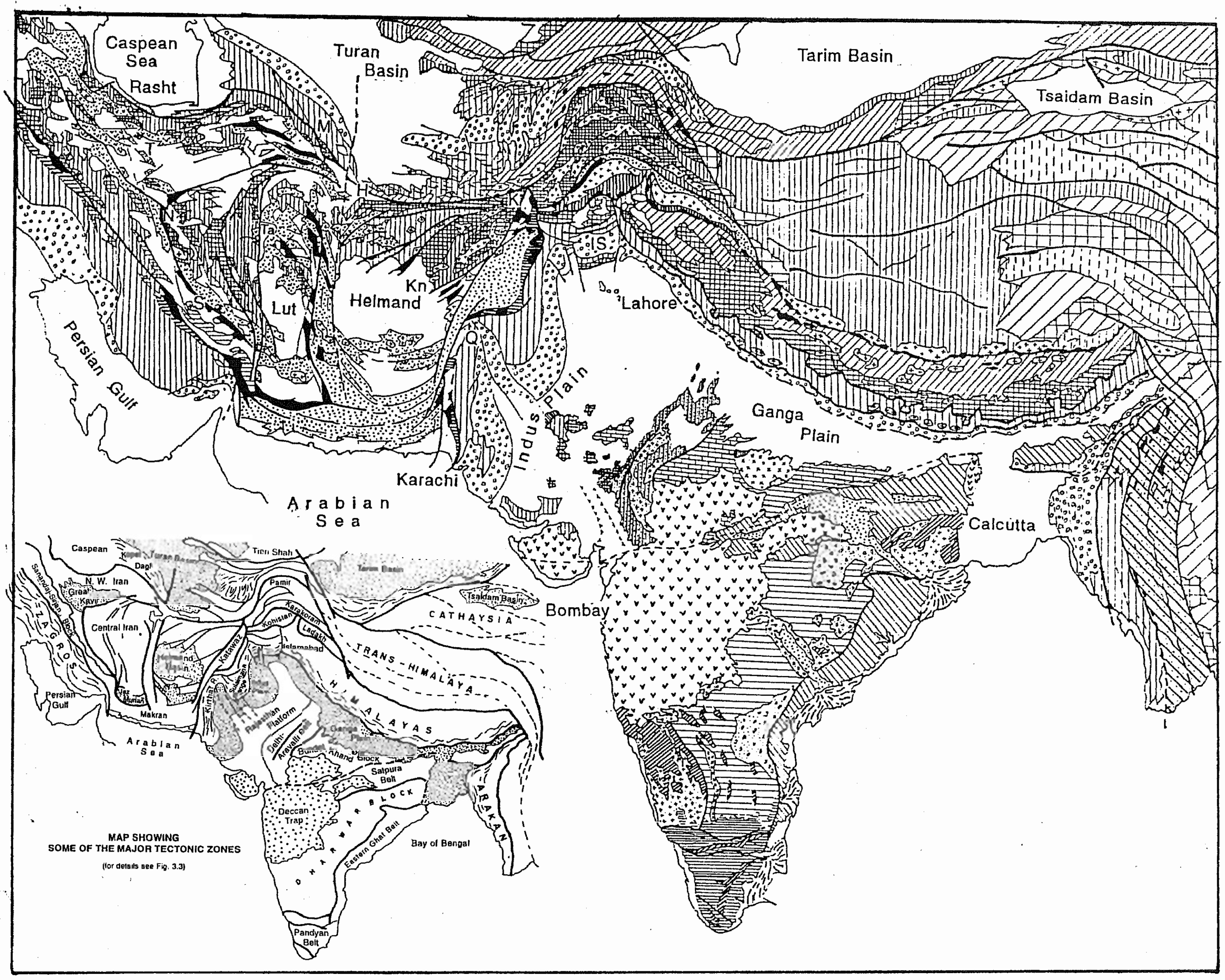


Fig. 3.4. Geological sketch map of Pakistan and surrounding regions (compiled from Kazmi and Rana 1982, Stöcklin 1977, Gansser 1980, CGMW 1990, GSI 1977, Drury et al. 1984, Radhakrishna et al.1986, Sugden et al. 1990, Ramakrishnan 1986, Kravchenko 1979).

EXPLANATION

GONDWANIC DOMAIN
Himalayas, Zagros, Arakan

- □ Quaternary Deposits
- ⁘ Neogene-Quaternary Molasse
- ▧ Cenozoic shelf deposits
- ✛ Paleozoic to Cenozoic granites
- ▤ Mesozoic oceanic sediments
- ▥ Mesozoic-Paleozoic pericratonic shelf deposits. Lesser Himalaya sequence in India
- ⁞ Mesozoic-Paleozoic Tethys Himalaya sequence
- ▦ Proterozoic-Cenozoic Crystalline Complex of High Himalaya

Indian Peninsular Region

- ▧ Cenozoic shelf deposits
- v Late Cretaceous Deccan Trap cover
- ▥ Mesozoic sedimentary cover
- ▨ Carboniferous-Neogene East Coast basinal deposits
- ⁘ Carboniferous-Cretaceous Gondwana Super-group
- ▦ Late Proterozoic Malani, Erinpura Granite
- ▦ Mid. Proterozoic Vindhyan and Marwar Super-groups
- ⁘ Mid. Proterozoic Cuddapah Super-group
- ▨ Early to Mid Proterozoic Gwaliar Fm. and L. Archean to E. Proterozoic Bijawar F. Aravalli & Bhilwara Formations
- ▧ Early to Mid Proterozoic sequence of Eastern Ghat and Satpura mobile belts (Charnokite, khondalite, granulite, anorthosite etc.)
- ▤ Archean-E. Proterozoic remobillised gneisses (Peninsular gneiss) granitoids and charnokites
- ▤ Late Archean granulites, metasediments, layered anorthosite, gabbro and ultramafic complexes
- X Late Archean granites closepet, Bundelkhand, Berach
- ▨ Archean greenstone belts, supracrustals, Dharwar, Sargur Groups
- ✛ Archean Peninsular Gneiss, Banded Gneiss complex

TETHYAN DOMAIN

- □ Quaternary Deposits
- v Tertiary-Quaternary Volcanics (locally Cretaceous)
- ▧ Cenozoic sedimentary and volcanic sequence
- ✛ Mesozoic-Cenozoic granite, diorite
- ■ Cretaceous-Paleogene ophiolite, melange, greenschist
- ⁘ Cretaceous-Neogene Flysch
- ▥ Cretaceous-Paleogene Kohistan-Ladak Island arc sequence
- ▥ Mesozoic sedimentary and volcanic sequence
- ▥ Triassic sedimentary sequence
- ▨ Paleozoic sedimentary sequence
- ▦ Precambrian sequence
- ▦ Ancient cores with Precambrian crystalline basement and Paleozoic to Triassic platform cover

LAURASIAN DOMAIN

- □ Quaternary deposits of intermountain basins
- ⁘ Neogene-Quaternary deposits of unfolded foreland to marginal fold zones
- ▤ Jurassic- Paleogene intercratonic basinal deposits
- X Hercynian and Early Cimmerian granites
- ■ Hercynian-Cimmerian ophiolites
- ▨ Paleozoic-Triassic sedimentary and volcanic sequence
- ▦ Ancient cores with Precambrian crystalline basement

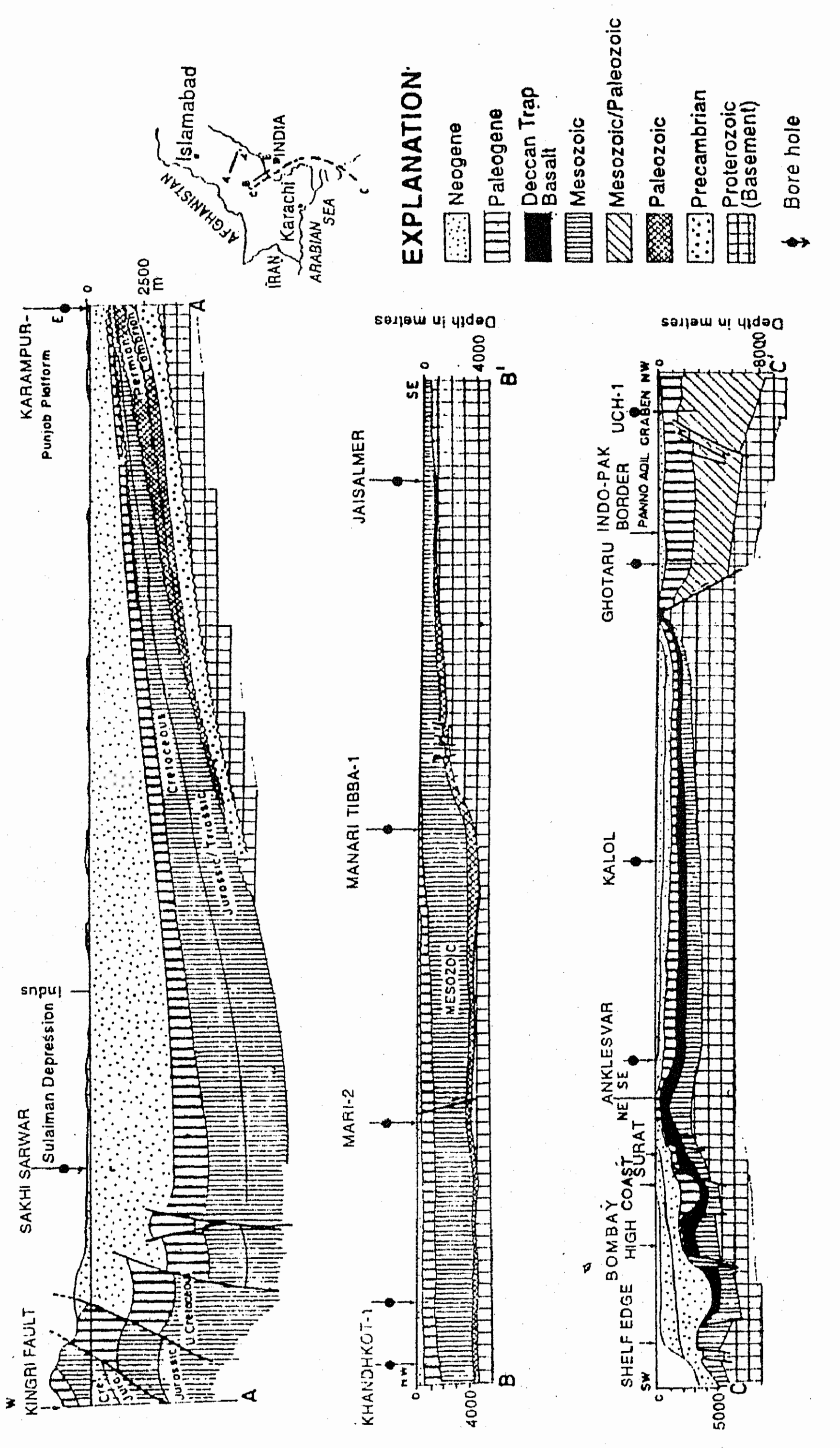


Fig. 3.5. Geological cross-sections across the Himalayan Foredeep (Sulaiman depression) and the Panjab-Rajasthan platform (from Ahmad et al. 1991 and Hiller et al. 1990). AA', section across the Panjab platform and the Sulaiman depression; BB', section across Khandkot-Mari High and Rajasthan platform; CC' section through Panno Aqil graben, Rajasthan platform, Cambay Basin and Bombay High.

subducted and has formed an underthrust beneath the frontal block and the Asian plate. Collision of the block has obducted widespread masses of melanges, exotic blocks and ophiolites in the suture zone. It has formed south verging folds and thrust belts of Phanerozoic sediments and volcanics over the basement rocks of the frontal block (Tethyan Himalayas). Continued under-thrusting of the rear block has scraped off Phanerozoic sediments and volcanoclastics from the sea floor as well as some of the basement rocks and they have been shoved southward as highly tectonised and metamorphosed masses, nappes and thrust slices. These imbricated metasediments form a distinct morphotectonostratigraphic belt commonly known as the Lesser Himalayas or the Himalayan Schuppen Zone.

Burial and friction has caused reactivation and remobilisation of the crust, particularly in High Himalayas, which has been often referred to as the Central Crystalline Axis. Southward the crystalline rocks form an extensive thrust, the Main Central Thrust (MCT), which is the sole thrust of the Cenozoic gneisses and schists developed along the deep crustal fracture. In some sections, south of the MCT, the crystalline rocks of the Central Crystalline Axis form extensive nappes and their remnants may be seen as klippes in the Lesser Himalayas (Fig. 3.7).

The rocks of the Lesser Himalayas have been thrust southwards and overlie the Neogene Siwalik molasse. This thrust zone, known as the Main Boundary Thrust (MBT), runs along the Himalayan foot-hills (Fig. 3.3). Farther southwards, there is another fault, the Himalayan Frontal Fault (HFF), along which the folded Siwaliks have been faulted against the less deformed Siwalik sequence of the Ganga Basin (Nakata 1972, 1989).

Table 3.1. Tectonostratigraphic zones of the Himalayas (listed from north to south).

Ladakh/Tibet Block
Indus - Tsangpo Suture Zone
Tethyan Himalayas
---Fault---
High Himalayas
Main Central Thrust (MCT)
Lesser Himalayas
Main Boundary Thrust (MBT)
Sub-Himalayas
Himalayan Frontal Fault (HFF)
Himalayan Foredeep

Le Fort's model (Fig. 3.6) suggests that India - Asia collision broke off the frontal lobe of the Indian plate into three slices (instead of two as envisioned by Powell et al. 1973) and they successively underthrust each other, resulting in surficial tectonic setting which is more or less the same as in Powell's model. Though these models provide a logical basis for dividing the Himalayas into various tectonostratigraphic zones, long before the advent of plate tectonic concepts, geologists working in the Himalayas (Burrard and Hayden 1907-1908, Burrard and Heron 1932, Wadia 1944, Gansser 1964) had adopted a realistic morphotectonostratigraphic subdivision of the Himalayas. This classification which is now accepted by most workers, has been shown in Table 3.1 and Figure 3.4.

Sub-Himalayas

The Himalyan foot-hills form the Sub-Himalayan zone and from the Punjab to Assam these hills are comprised of a narrow belt of folded Neogene molasse type sediments (Siwaliks). Southward the folded Siwalik sequence is covered by the Indo-Gangetic alluvium.

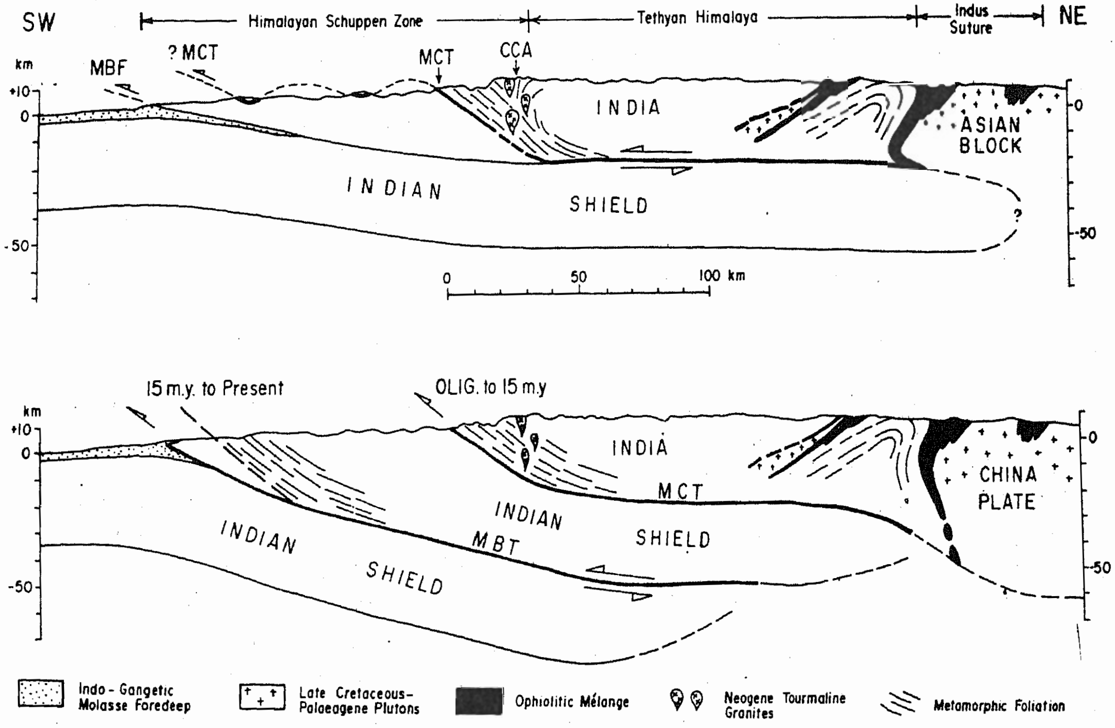


Fig. 3.6. Two interpretation of the gross structure of the Himalayas. Upper section from Powell and Conaghan (1973) and lower section from Le Fort (1975). Sections are drawn to fit the surface geology and topography along a NE-SW line in the Kumaon Himalayas. Coupled arrows represent the postulated present-day zones of underthrusting.

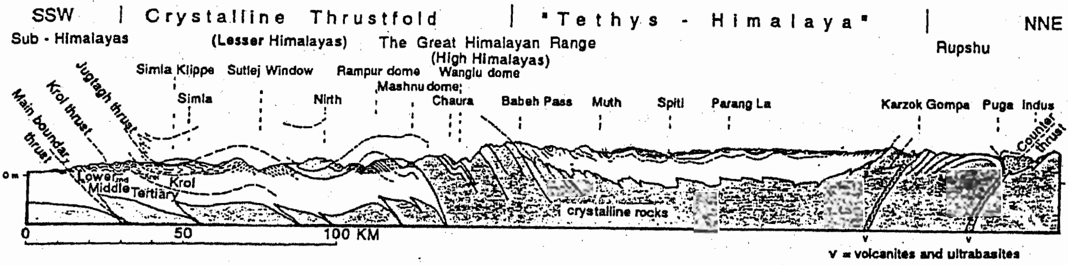


Fig. 3.7. Generalised section of the Himalayas between Simla and Rupshu (from Gansser 1964).

Traces of an active fault– the Himalayan Frontal Fault (HFF), cut the alluvium in the foothill region. According to Nakata (1972, 1989), this fault comprises a series of en echelon faults which run parallel to the strike of the hills between Punjab and Assam. Northward this sequence is terminated by the Main Boundary Thrust (MBT), which comprises a set of north-dipping faults and forms the boundary between the Sub-Himalayas and the Lesser Himalayas (Fig. 3.3 and 3.7).

Lesser Himalayas

This zone is bounded to the north by the Main Central Thrust (MCT) and to the south by MBT. In its western part, this zone is comprised of Precambrian to Late Paleozoic metasediments (including carbonates with Riphean stromatolites), and Paleozoic sedimentary and volcanic rocks. These metasediments have been over-ridden by thrust nappes of high grade gneisses derived from the Central Crystallines (Valdiya 1980a,b, Sinha 1981). The Late Proterozoic sections, such as the Almora nappe in Kumaon, have been intruded by 510–520 m.y. S-type granites (Jaeger et al. 1971). Various cross-sections in Figures 3.8 provide a general picture of the structure of this zone. The highly complicated structure of the Lesser Himalayas and the difficulties in stratigraphic correlation has led to different interpretations by various workers and this is well exemplified by the three cross-sections of the Himalayas in the Everest region as shown in Figure 3.8.

Main Central Thrust (MCT)

This thrust was initially defined by Auden (1937) and Heim et al. (1939) as a thrust fault which had brought the high grade crystalline rocks over the lower grade metasediments. Valdiya (1980a) has shown that it lies at a higher level than originally proposed and occurs within an inverted metamorphic sequence. According to some Himalayan geologists (Le Fort 1975, Valdiya 1980a) MCT forms a ductile zone associated with the main phase of Himalayan deformation. At several places the MCT lies above a more brittle and younger thrust known as Jutogh or Mansiari Thrust which is responsible for over-thrusting of some of the crystalline outliers of the Lesser Himalayas (Le Fort 1989).

Pecher (1978) introduced the notion of a "Main Central Thrust Zone" considering the fact that MCT is not only a thrust plane but reflects a whole column of rocks with ductile deformation extending for several kilometres on both sides of MCT. Crustal scale thrusting associated with MCT has produced extensive inverted metamorphic zones. The inverted metamorphism passes upwards from the chlorite-biotite zone through a Barrovian-type sequence (garnet, staurolite, kyanite and sillimanite zones), with migmatites and leucogranites in the sillimanite zone (Windley 1988). According to the thermal model of Le Fort (1975), the metamorphic zones are related to the post-collisional intracontinental thrusts which resulted in crustal thickening and reached a peak 20 Ma after collision. The MCT was formed about the same time (Miocene). Inverted metamorphism could also result from the thrusting of a hot slab of central crystalline gneisses, over a cold slab of lower Himalayan sediments (Windley 1988).

Some of the Himalayan leucogranites are contemporaneous with MCT and have formed around 14 to 25 Ma (Deniel et al. 1987). The MCT is believed to have been involved in active thrusting for atleast 10 million years before the movement shifted to Main Boundary Thrust in the south (Le Fort 1989). Seeber et al. (1981) have shown that earthquake foci along the MCT indicate the presence of a major, northward dipping detachment surface

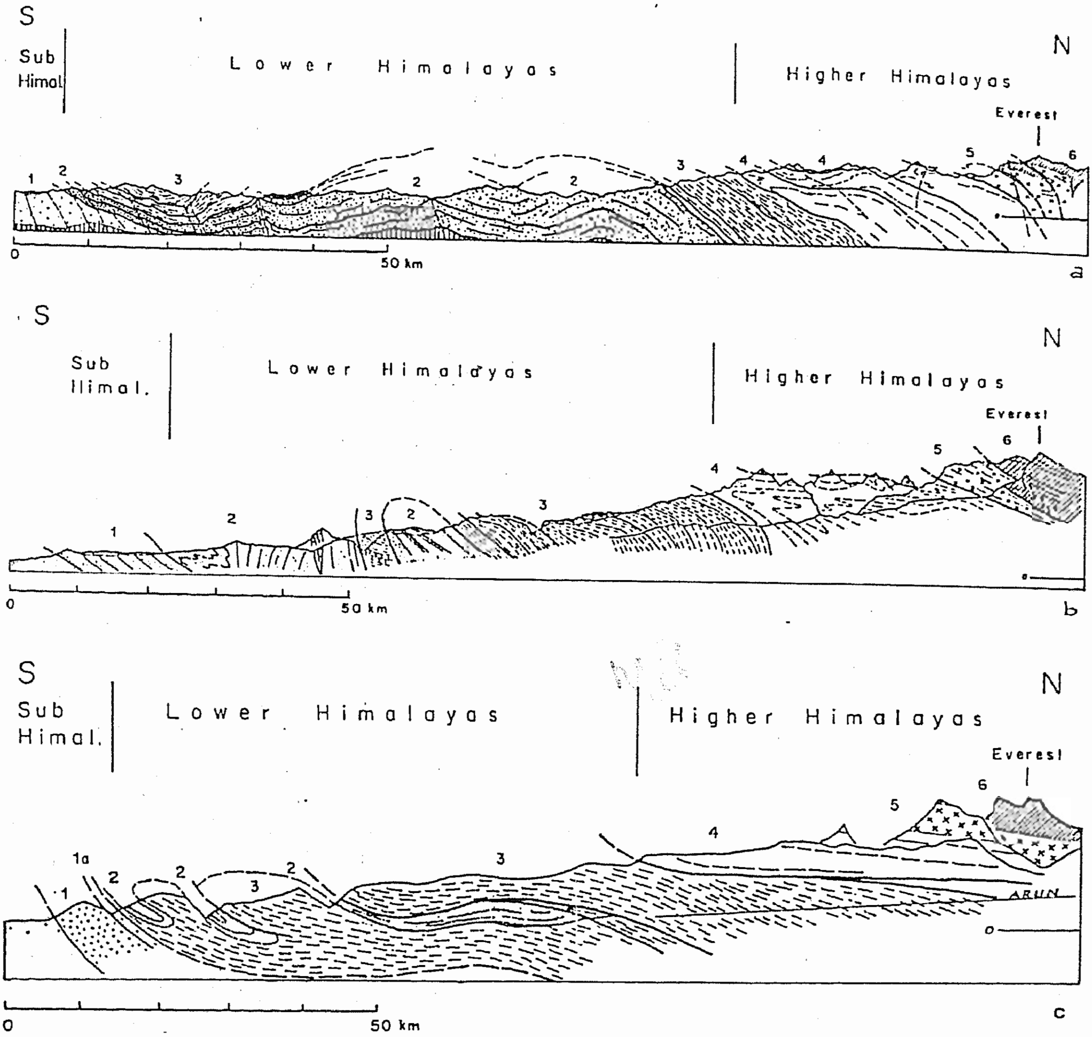


Fig. 3.8. Geological cross-sections from Everest to the Siwaliks, a comparison; a–from Hagan (1959), b–from Lombard (1958), c–from Bordet (1961); all sections reduced to same scale; sections b and c have slightly exaggerated heights. 1–Siwaliks, 1a–Sanguri (Bordet), 2–Nawakot nappes (serie de couverture, Bordet), 3–Kathmandu nappes (migmatites of Lower Himalaya, Bordet), 4–Khumbu nappes, 5–Makalu granite, 6–Tibetan zone, Everest. (From Gansser 1964).

which extends under the Tibetan block north of the Indus-Tsangpo Suture.

High Himalayas

The MCT forms the base of a huge 10–15 km thick slab of high-grade metamorphic rocks which overlie the Lesser Himalayan sequence. This infracrustal thrust sheet of Precambrian Central Crystallines forms the High Himalayas and the very core of the Himalayan Range (Figs. 3.7 and 3.8). The basal part of the section is comprised of Precambrian gneisses with whole rock ages of 1,500–1,800 m.y. though these ages, according to Bhanot et al. (1977), mask still older metamorphic events. Medium pressure metamorphism during southward thrusting has formed widespread pelitic schists, marbles, paragneisses, orthogneisses, amphibolites and migmatites (Burg et al. 1987). The Central Crystalline Complex has been intruded by two-mica leucogranite. According to Denial et al. (1987) one of these granites, the Monasulu Granite has yielded magmatic ages of 25 Ma and 18 Ma. It is commonly believed that these granites formed by anatexis of the lower crustal Precambrian paragneisses (Gansser 1981, Windley 1988).

Tethyan Himalayas

The High Himalayas and its Central Crystalline Complex is bounded on the north by a northward dipping normal fault, the Trans-Himadri Fault (Valdiya 1989) (Figs. 3.7 and 3.9). It is a Late Tertiary gravity collapse structure with several tens of kilometre movement (Burchfield et al. 1985). North of this fault and overlying the Central Crystalline Complex there is a thick sequence of Late Precambrian, Paleozoic and Mesozoic fossiliferous Tethyan sediments which were deposited on the northern passive continental margin of India (Thakur 1981).

The Late Precambrian sequence of turbidite, deltaic and shelf sediments grades upwards into similar Cambrian and Early Ordovician sediments. The latter are overlain by a relatively thin Middle Ordovician to Carboniferous sequence of carbonates and quartzites with subordinate pelites. In fact the Ordovician limestone forms the summit of the world's highest peak, the Everest (Fig. 3.8). The Upper Carboniferous sequence largely contains glacial and continental Gondwana deposits. This sequence is interrupted by the Permian Panjal Trap. Following extensive rifting and outpouring of the Panjal Trap, the northern continental margin of India subsided rapidly, resulting in the formation of a thick sequence of limestone in the west (Zanskar and Spiti area), shale and sandstone in the centre (Kumaon), and sandstone, shale and limestone in the east (Nepal-Bhutan). During the Cretaceous, flysch type sediments comprising calcareous and variegated sequence of sandstone, shale and glauconitic beds were deposited. Towards the Late Cretaceous there was complete regression of the sea. A brief and limited transgression occurred subsequently with deposition of fossiliferous Eocene limestones followed by deposition of Neogene molasse (Varadan 1976, Gaetani et al. 1991).

The Tethyan Himalayas contain lower Ordovician gneissic porphyritic granites (Rb Sr isochron age 485 Ma), which are similar to the gneisses in the Central Crystalline Zone of High Himalayas (Burg et al. 1984). The geological structure in the western part of the region is relatively simple with open Jura-type folds (Fig. 3.9). In the eastern part, complex folds and thrust zones have formed due to buckling of the Tethyan sequence before the southward bulging mass of the Indus-Tsangpo Suture Zone.

Indus – Tsangpo Suture Zone

This suture zone terminates the Tethyan Himalayas on the north and marks the boundary between the Indian crustal plate and the various crustal fragments that had earlier accreted to the Asian Plate in the north. In the western region (Swat, Nanga Parbat), the Indus Suture is comprised of a complex sequence of imbricated melanges which are composed of tectonic blocks of ophiolites, blueschists, greenschists, metavolcanics and metasediments in a matrix of sheared meta-sediments and/or serpentinite (Jan 1980, Kazmi et al. 1984,

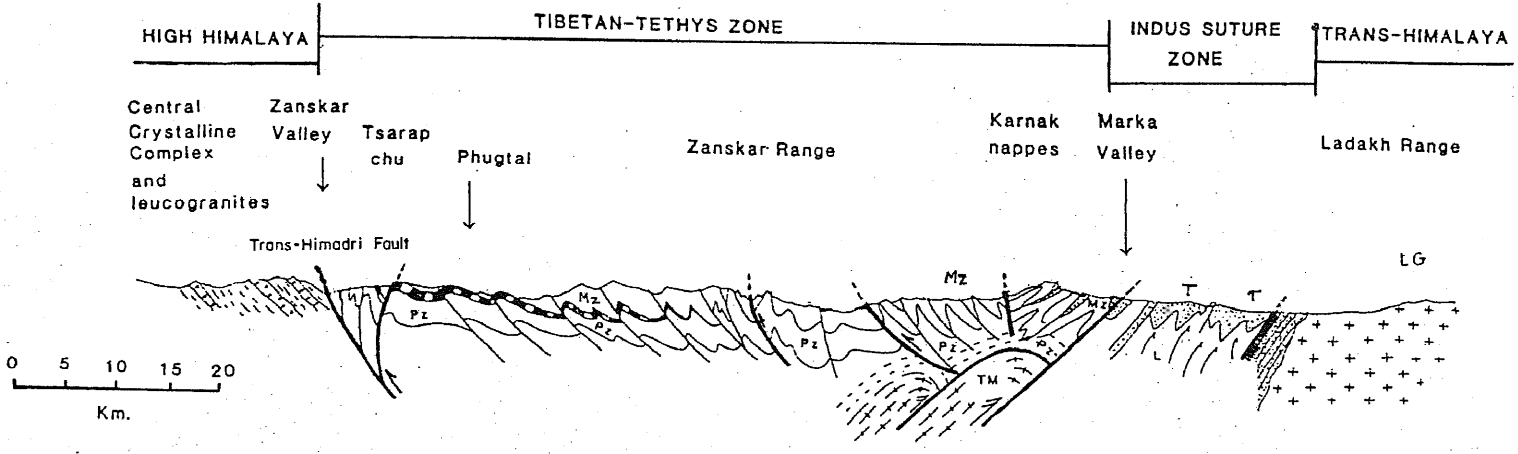


Fig. 3.9. Section across the Tibetan-Tethys zone of Zanskar and the Indus Suture Zone of Ladakh. Pz-Paleozoic, Mz-Mesozoic,T-Tertiary. The Ladakh granitoids (LG) occur along the northern edge of the Indus Suture Zone. Deep-level culminations of granite gneiss domes are extrapolated westward from the Tso Morari dome (TM). Upward-decreasing metamorphic isogrades above the Tso Morari granites are shown as dashed lines (from Searle 1986).

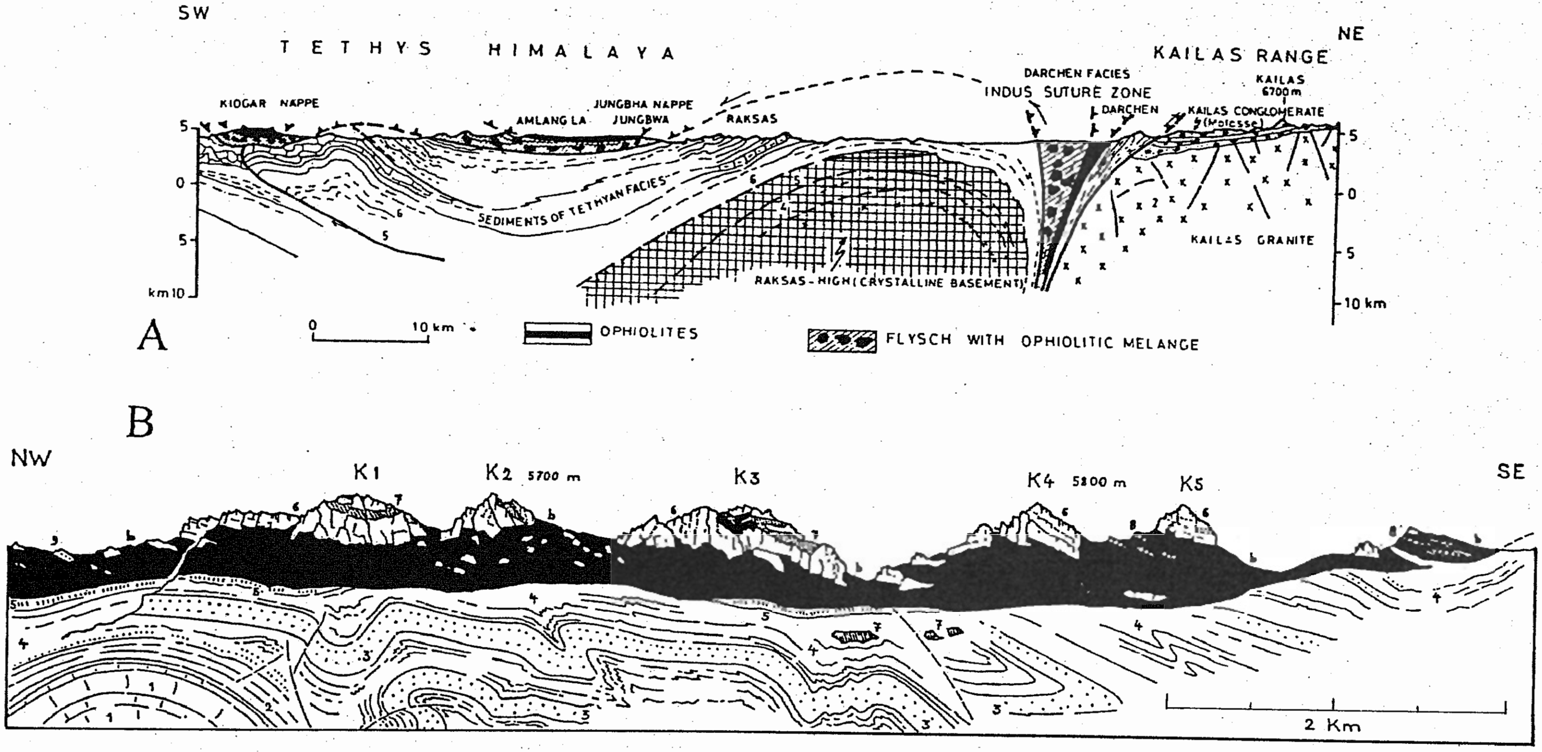


Fig. 3.10. Geological sections across the Indus-Tsangpo Suture Zone and the Tethyan Himalayas showing the general structure and the ophiolitic nappes (from Gansser 1964, Valdiya 1989). Secion B gives an enlarged view of Kiogar nappes seen at the southwestern end of section A.

Section A. Indus-Tsangpo Suture Zone and the Jungbha and Kiogar nappes. 1–ophiolites, 2–granites, granodiorites, 3–flysch with ophiolitic melange, 4–Precambrian metasediments, 5 and 6–Precambrian-Lower Paleozoic sediments, 9–Muth quartzite (Ordov.-Devonian), 10–Permo-Carboniferous sequence, 12, 13, 14–Mesozoic sedimentary rocks.

Section B. Section through the Kiogar Peaks, Tibetan Himalayas. 1–Kioto lst (U. Trias), 2–Spiti sh, 3–Giumal sst (U. Cret.), 4–sst with radiolarians (U. Cret.), 1–5 comprise Himalayan facies sequence. 6–Kiogar lst (Trias ?), 7–oolitic lst and Calc. schists with *Calpionella* (U. Jura.), 8–radiolarite, 9–green sst, b-basic and ultrabasic rocks with exotic blocks from Permian to Cret. age; 6–9 and b comprise Tibetan facies sequence.

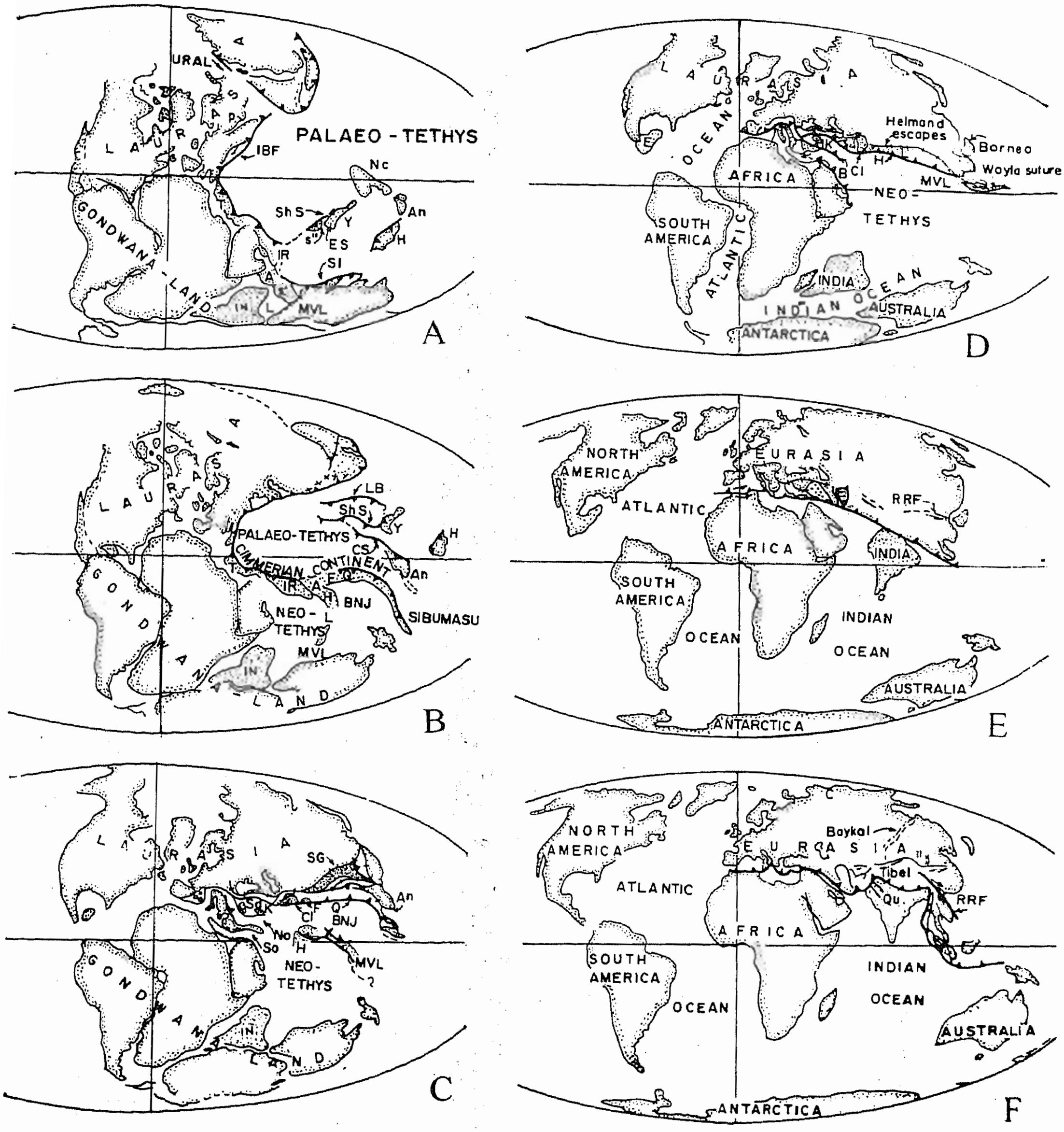


Fig. 3.11. Generalised reconstruction of the continents showing evolution of the Tethyan Domain (modified from Sengör et al. 1988). A, Late Permian; B, Early Triassic; C, Late Jurassic; D, Late Cretaceous; E, Middle Eocene; F, Late Miocene.

A–Afghan block, *An*–Annamia, *B*–Bitlis-Pötürge fragment, *BNJ*–Waser/Rushan-Pshart/Banggong Co-Nu Jiang/Mandalay Ocean, *CI*–Central Iranian microcontinent, *CS*–Chola Shan, *ES*–Emei Shan, *F*–Farah block, *H*–Helmand block (sensu Sengör 1984), *Hu*–Huanan block *s.l.*, *IBF*–Istanbul-Balkan fragment, *IR*–Iranian block (undisrupted: i.e., Sanandaj-Sirjan zone + north-west Iran and central Iranian microcontinent), *K*–Kirsehir block, *L*–Lhasa block, *LB*–Luochou Arc, *MVL*–Mount Victoria Land, *NC*–North China, *No*–northern branch of Neo-Tethys, *p*–Pachelma aulacogen, *Q*–Quangtang, *Qu*–Quetta-Sibi graben, *RRF*–Red River Fault, *s'.s''*.–Tarim, *S'*–Pamir-west Quangtang block, *S''*–

1986, Majid et al. 1985) and a tectonic wedge of high pressure garnet granulites (Jan and Howie 1981).

Eastward the suture zone contains allochthonous, pre-orogenic Triassic to Late Cretaceous basinal sediments comprising shales, turbidites, deep-water radiolarian cherts (Lamayuru Complex) and ophiolitic melanges (Thakur 1981, Frank et al. 1977). At places the granitic rocks of the Ladakh batholith intrude the melanges. Dras volcanics of Jurassic to Cretaceous age also occur in this zone (Radhakrishna et al. 1984). In Tibet the suture zone consists of flysch, conglomerates and ophiolites in several thrust slices (Burg et al. 1987). Ophiolitic nappes have been thrust on the Tethyan sequence to the south of the suture zone (Fig. 3.10). The ophiolite nappes include the Spontang nappe in Zanskar and the Jungbwa nappe in southwest Tibet (Gansser 1981).

TETHYAN DOMAIN

Encompassing India and the northern borders of Arabia and Africa, the Tethyan Domain stretches from the Pacific to the Mediterranean (Fig. 3.2). In shape it mimics the Tethys, the primogenic sea, being narrow in the west and extremely wide in the east. It is largely comprised of an enormous orogenic collage consisting of a number of continental blocks, stitched together, as it were, by a ramifying network of sutures.

Evolution of the Tethyan Domain.

This domain evolved through the unique process of rifting, successive fragmentation and "calving" of several continental blocks from Gondwanaland and their northerly migration across the vast oceanic spaces to unite with Laurasia. This process started in Middle to Late Paleozoic even while the assembly of Pangaea was in progress (Sengör et al. 1988). Thus the evolution of the Tethyan Domain, in essence, is linked with the opening and closure of several ocean spaces, through the repeated process of seafloor spreading, subduction and continental collision, involved in the drifting of each of the major crustal block (Fig. 3.11). With the assembly of Pangaea in Middle Carboniferous, the earliest of these ocean spaces between Laurasia and Gondwanaland has been universally known as the Paleo-Tethys. In Late Permian, rifting along the northern margin of Gondwanaland formed a Cimmerian microcontinent (Sengör et al. 1988) which has been referred to as 'Kreios' by Nakazawa and Dickins (1985).

This event was followed by (1) the opening of other smaller oceans that were back arc basins of Paleo-Tethys and (2) by the opening of the Neo-Tethys (Fig. 3.11). In the Late Permian, the Cimmerian continent began to disintegrate into independent blocks or island arcs in an ocean domain–the Waser Rushan Pshart / Banggong Co-Na Jiang / Mandalay Ocean (so named after the suture zones by Sengör et al. 1988). Shvolman (1978) and Belov (1981) refer to this ocean space as the Meso-Tethys and it conforms to Tethys-2 of Boulin (1981). By Late Triassic times most of these blocks had collided with Laurasia and by the Early Jurassic the Paleo-Tethys had closed. The Paleo-Tethyan sediments and suture-associated ophiolites and melanges were subsequently affected by Eocimmerian orogenic events and associated magmatism and metamorphism. Within the Tethyan Domain, this earlier collage of sutures and colliding blocks has been named Cimmerides by Sengör (1984).

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east Quangtang block, *Sa*–Sakarya Continent, *SG*–Songpan-Ganzi system, *ShS*–Shaluli Shan Arc, *Sl*–Sibumasu–China-Burma-Malaya-Sumatra portion of the Cimmerian continent, *So*–Southern branch of Neo-Tethys, *T*–Turkish blocks.

While the Paleo-Tethys closed in the north, during Jurassic the Neo-Tethys continued to grow and a new subduction system formed along the southern margin of the Cimmerides or within the Neo-Tethyan oceanic lithosphere. New fragments of Gondwanaland were rifted off and accreted to the southward growing margin of Eurasia along a younger series of Mesozoic to Cenozoic sutures. The Neo-Tethys finally closed with the collision of India (Indus-Tsangpo Suture) and Arabia with Eurasia (Zagros Suture). This younger Neo-Tethyan collage of sutures and continental blocks now forms a part of the Alpine-Himalayan orogenic belt and has been referred to as the 'Alpides' by Sengör (1984).

A number of models have been proposed by various workers to explain the evolution of the Tethyan Domain. One of these has been shown in Figure 3.11, whivch is self explanatory. The differences in the timing of Gondwanian rifting, positioning of various continental blocks, and the paleogeography of the ocean spaces proposed by various workers are mainly due to the meager paleomagnetic and geological data and the difference in the interpretative logic of the authors. The significant point, however, is that the Tethys was preceded by an earlier ocean between Laurasia and Gondwanaland, the Proto-Tethys of Dewey and Bird (1970), pre-Hercynian Ocean of Irving (1979) and Tethys I of Boulin (1981). Its position was somewhat different from the Paleo-Tethys, though it had similar internal configuration. It was characterised by a northern elongated basin, an east-west oriented middle ridge (the Protokreios microcontinent), and a southern trough in Europe (Tollman et al. 1985). The closure of this ocean space formed the Hercynian orogenic belt along the southern margin of Laurasia. This belt forms the northern limit of the Tethyan Domain. However, included within the Tethyan Domain there are many components with a pre-middle Carboniferous history of dispersal.

Geological Outline of the Tethyan Domain

The Tethyan Domain is neatly defined and charactcrised by several sutures along which the dismembered Gondwanic blocks have been accreted to Eurasia (Fig. 3.2). Each of these sutures also indicate the position of ancient oceanic realms (Songpan–Ganzi, Farah Rud, Makran–East Iran), island arcs (Shalali Shan, Hohxil Shan, Chola Shan, Qaidam, Ladakh, Kohistan), continental margin magmatic arcs (Central Pamir–West Qangtang), and back-arc/inter-arc basins and oceanic trenches (Kandhar Zone, Oruzgan Trench, Katawaz Basin). The sutures are characterised by extensive outcrops of ophiolites, melanges, oceanic sediments, subduction related as well as covergent margin magmatism, greenschist to amphibolite grade metamorphism and syntectonic granites. Collisional mountain belts (Zagros, Alburz, Himalayas, Hindu Kush–Karakoram) have formed along some of these sutures.

The component blocks of this Domain are composed of an Archean to Proterozoic crystalline basement which is unconformably overlain by a thick Phanerozoic sequence (Stöcklin 1977, Kravchenko 1979, Boulin 1990). The Paleozoic sequence is largely comprised of shallow-water platform-type clastics, carbonates, volcanics and evaporites. Extensive rifting, volcanism and deformation during Early Mesozoic (Triassic) resulted in a regional unconformity, bauxite formation, and molasse deposits (Mennesier 1977, Tapponier et al. 1981). Above this unconformity Mesozoic coarse clastics, shallow water carbonates, coal beds, evaporites, molasse and red beds have been deposited (Stöcklin 1968, Boulin 1988, Blaise et al. 1982). Subsidence and deposition has formed a thick clastic sequence in some regions (Mesozoic in Tabas and Lut; Cenozoic in Farah Rud).

Cretaceous to Neogene flysch deposits occur in East Iran, Farah Rud, Helmand, Katawaz and Makran (Schreiber et al. 1972, Blaise et al. 1982).

Extensive Cenozoic volcanism covers Alburz, Central Iran, Balochistan and Kandhar region (Davies et al. 1972, Jacob et al. 1972, Sengör et al. 1988). The Tethyan Domain has been affected by Cimmerian and Alpine orogenic events with synorogenic magmatism from Jurassic to Neogene. Several prominent faults form an intricate network of geofractures (Figs. 3.2 and 3.3). Some of these are extensional features while others are thrust or large continental scale faults which have formed along suture zones or crustal-block margins. Several of these faults are seismotectonically active.

4

Tectonic Framework of Pakistan

GEODYNAMIC SETTING

The Indian Ocean and the Himalayas, two of the most pronounced global features surrounding the Indo-Pakistan subcontinent, have a common origin. Both are the product of the geodynamic processes of sea-floor spreading, continental drift and collision tectonics. A plate of the earth's crust carrying the Indo-Pakistan landmass rifted away from the supercontinent Gondwanaland followed by extensive sea-floor spreading and opening up of the Indian Ocean. Propelled by geodynamic forces the Indian Plate travelled 5,000 km northward and eventually collided with Eurasia. The subduction of the northern margin of the Indian plate finally closed the Neotethys and the Indian Ocean assumed its present widespread expanse. This collision formed the Himalayas and the adjacent mountain ranges. The geodynamics of the Indo-Pakistan subcontinent is briefly reviewed in the following section.

Northward drift of India and opening of Indian Ocean

The Indo-Pakistan subcontinent separated from the Gondwana motherland about 130 million years ago (Johnson et al. 1976). Its precise location within Gondwanaland in relation to Africa, Antarctica and Australia is uncertain. Various authors have placed it in different positions (Fig. 4.1). This is because prior to the plate tectonic concepts, reconstructions of Gondwanaland were entirely dependent on land-based data and geological arguments. Paleomagnetic data from ocean floor has enabled more precise reconstructions of the ancient supercontinents (Fig. 3.1). Nevertheless, India's predrift location vis-a-vis Australia, Antarctica and Madagascar has remained problematic as these plates have moved and rotated at varying rates and in different directions since India's separation. Furthermore, the topography of the Indian Ocean and its sea-floor spreading pattern are complex and magnetic lineations on the older parts of the ocean between the continents are absent (McKenzie et al. 1976). Thus while fixing India's position, together with geological and bathymetric arguments, various assumptions have been made concerning relative positions of Madagascar, Chagos Trench and the Mauritius Fracture Zone, and rotation of various plates (McKenzie et al. 1976, Powell 1979).

Though older magnetic anomalies are known , their age and relation to the relatively younger anomaly 32 are not established. Anomalies 33 and 32 are thus the earliest ones on which the reconstruction of the Indian Ocean and the relative positions of the continents since 80 to 75 m.y. is based (Fig. 4.2). It is believed that Chagos "trench" and Mauritius Fracture Zone were linked and a transform fault extended along Ninety East Ridge (McKenzie et al. 1976). The ridge south of Sri Lanka (precursor of Mid-Indian Ridge) was spreading in such a way as to propel India northward. The Chagos-Mauritius and Ninety East transform faults on its either side facilitated its northward movement. New crust was continuously generated along the spreading ridges. Australia and Antarctica

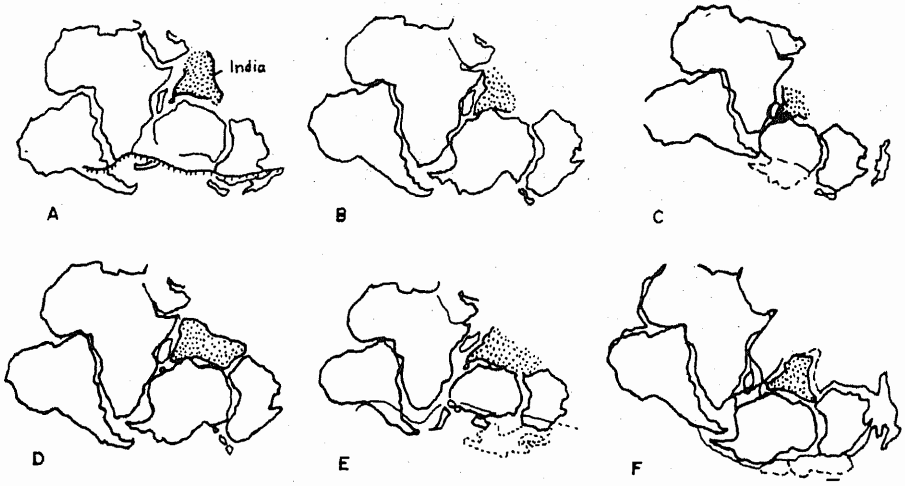


Fig. 4.1. Reconstructions of the southern continents. A–du Toit 1937, B–Smith and Hallam1970, C–Tarling 1972, D–Veevers et al. 1975, E–de Wit 1977, F–Barron et al 1977.

remained connected at this time as a single continent while Madagascar, Africa and South America had already separated.

It has been estimated that between 130 m.y. and 80 m.y. India moved northward at a rate of 3 to 5 cm/year (Johnson et al. 1976). Thereafter its movement accelerated considerably. The vast distance (5,000 km) between the matching set of anomalies 21 and 32 shows that from 80 m.y. ago India moved at an average rate of about 16 cm/year relative to Australia and Antarctica (Powell 1979). According to Patriat and Achache (1984), before anomaly 22 (50 m.y.) this rate of movement varied between 15 and 25 cm/yr. The movement was facilitated by transform faulting in the Proto-Owen fracture zone and extensive sea-floor spreading along Mid Indian Ocean Ridge. It is noteworthy that extensive extrusion of Deccan Trap Basalts occurred between 65–60 m.y. ago (Duncan et al. 1988), during the fast northward drift of India.

Formation of Kohistan-Ladakh Island Arc and its Collision with Eurasia

The Neotethys had begun to shrink by the time India began its northward drift around 130 m.y. ago. Intra-oceanic subduction generated a series of volcanic arcs (Kohistan-Ladakh, Nuristan, Kandhar) during the Cretaceous (Searle 1991, Treloar and Izatt 1993). One of these, the Kohistan-Ladakh arc, has been studied in considerable detail (see Chapter 6) and it appears that the mafic/ultramafic components of the arc-sequence formed during intraoceanic subduction stage. This arc was intruded by 102 Ma precollision granitoids (Petterson et al. 1985, Treloar et al. 1989b, 1993) followed by intra-arc rifting and magmatism (Khan et al. 1993). Arc magmatism covered a life-span of about 40 Ma after which the back-arc basin closed and Kohistan-Ladakh arc collided with Eurasia along the southern margin of the Karakoram plate. After accretion the arc formed an Andean-type continental margin (Petterson et al. 1985).

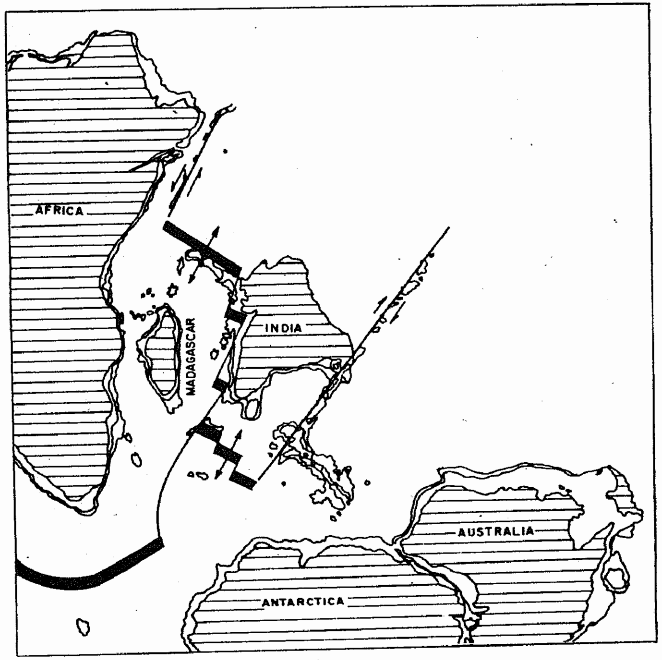


Fig. 4.2. Position of Indo-Pakistan plate 75 m.y. ago. After breaking away from Gondwanaland this plate moved rapidly northeastward. New crust was generated along the ridges marked by parallel lines and the transform faults east and west of the connected series of ridges enabled unimpeded movement of the Indo-Pakistan plate (modified from McKenzie and Sclater 1976).

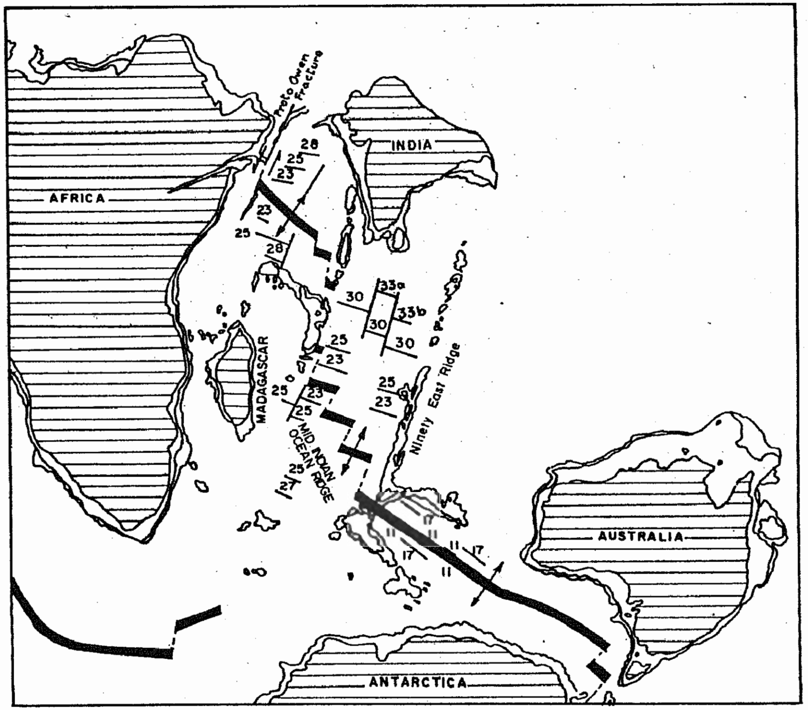


Fig. 4.3. Position of Indo-Pakistan plate 35 m.y. ago. By this time India had moved 4,000 km northward. At this stage direction of relative motion among the major Gondwanian plates changed drastically to produce the present-day sea-floor geometry (see Fig. 4.4). Modified from McKenzie and Sclater 1976.

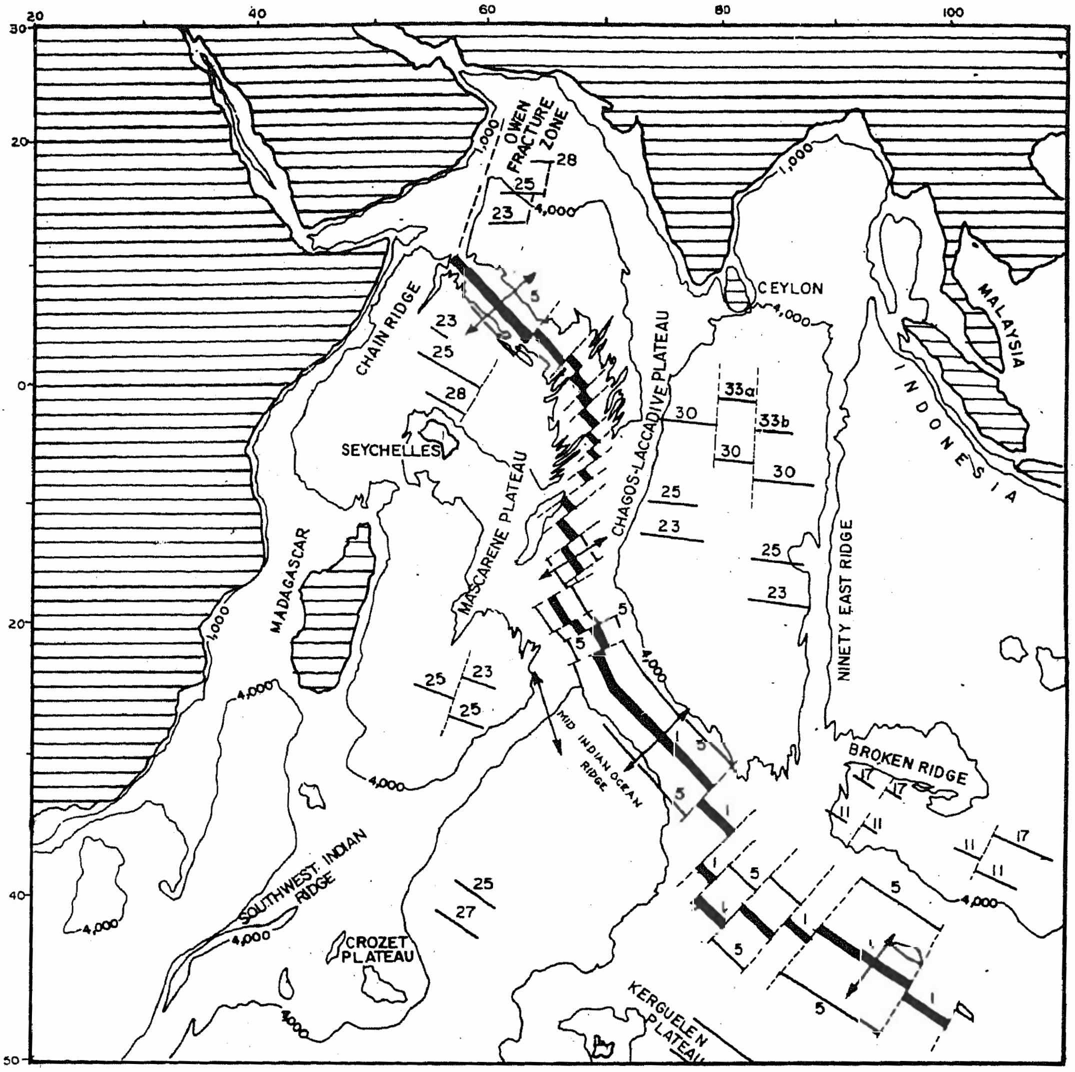


Fig. 4.4. Magnetic anomalies and the major sea-floor features of the Indian Ocean. Thick lines with numbers indicate some of the magnetic anomalies along present ridge axis. Dashed lines show transform faults. The more pronounced sea-floor features are the N-S Ninety East Ridge, Chagos–Laccadive Plateau and the Mascarene Plateau, and the E–W trending Broken Ridge and the Kerguelan Plateau. (Modified from McKenzie and Sclater 1976).

The time of Kohistan-Eurasia collision is constrained by the stratigraphy, structure and geochemistry of the Karakoram-Kohistan region and it appears that this event occurred during the Turonian period. The youngest marine sediments deposited in the back-arc basin, between Kohistan and Karakoram, are Early to Middle Cretaceous carbonates and volcanics (Chalt Formation). Radiometric data suggests that both Kohistan and Karakoram were intruded by early subduction related 102-95 Ma granites (Searle 1991). The suturing event is constrained by the 102 Ma Rb-Sr age of precollision

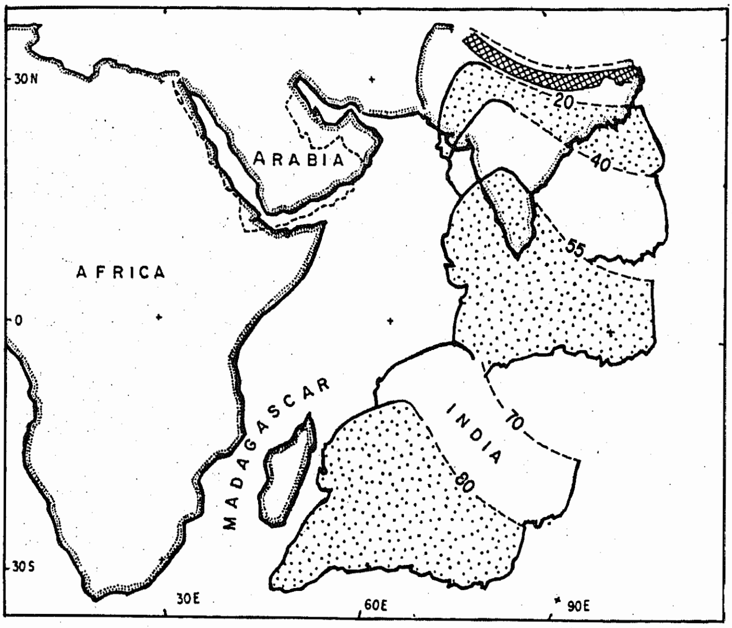


Fig. 4.5. Position of Greater India relative to Africa and Madagascar fixed in their present positions since 80 m.y. The stage positions are derived assuming the relationship between oceanic magnetic anomalies and the absolute time scale as proposed by La Brecque and others (1977). The cross-hatched strip is the present doubled-up Indian crust between the Himalayan front and the Indus Ophiolite belt. The dotted line indicates the position of Arabia before the Neogene opening of the Red Sea. (From Powell 1979).

Matum Das pluton and 75 Ma hornblende cooling ages of post collision dykes in the area. According to Treloar et al. (1989b,c) folding and cleavage developed in the suture zone around 100-95 Ma.

India-Eurasia Collision and Himalayan Upheaval

Continued subduction of the Neo-Tethys beneath Kohistan-Ladakh arc and Eurasia resulted in complete consumption of the leading oceanic edge of the Indian plate and its eventual collision with remnants of the Kohistan-Ladakh arc. The abrupt slowing down of India's northward movement between 55 and 50 m.y. ago is attributed to this collision (Powell 1979). According to Patriat and Achache (1984) the northward drift of India was dramatically slowed down at the time of anomaly 22 (50 m.y.) with the collision between India and Eurasia and the closure of Neotethys in Tibet. They consider that collision with the Ladakh island arc occurred earlier, near 54 m.y. though it did not affect the movement of the Indian plate. New paleomagnetic data from the Ninety East Ridge (Klootwijk et al. 1992) indicates that the abrupt slowing down of India from 18–19.5 cm/yr to 4.5 cm/yr occurred at 55 ± Ma which agrees closely with a reduction in spreading rate of the central Mid-Indian Ridge at magnetic anomaly 24. According to Klootwijk et al. (1992), the initial contact between northwestern Greater India and Asia was established at about 65 Ma when India's northwest margin had crossed the equator. It is interesting to note that slowing down of India's northward movement coincided with separation of Australia and Antarctica 45 m.y ago. A combined India-Australia plate started moving away from Antarctica (Fig. 4.3). Motion ceased along the former plate boundary (the Ninety East Ridge), and the Proto-Owen fracture no longer remained a transform fault, though it was reactivated later, about 20 m.y. ago. The direction of relative motion between major plates changed significantly, resulting in complicated pattern of ridge sediments, transform faults and other sea-floor features (Fig. 4.4).

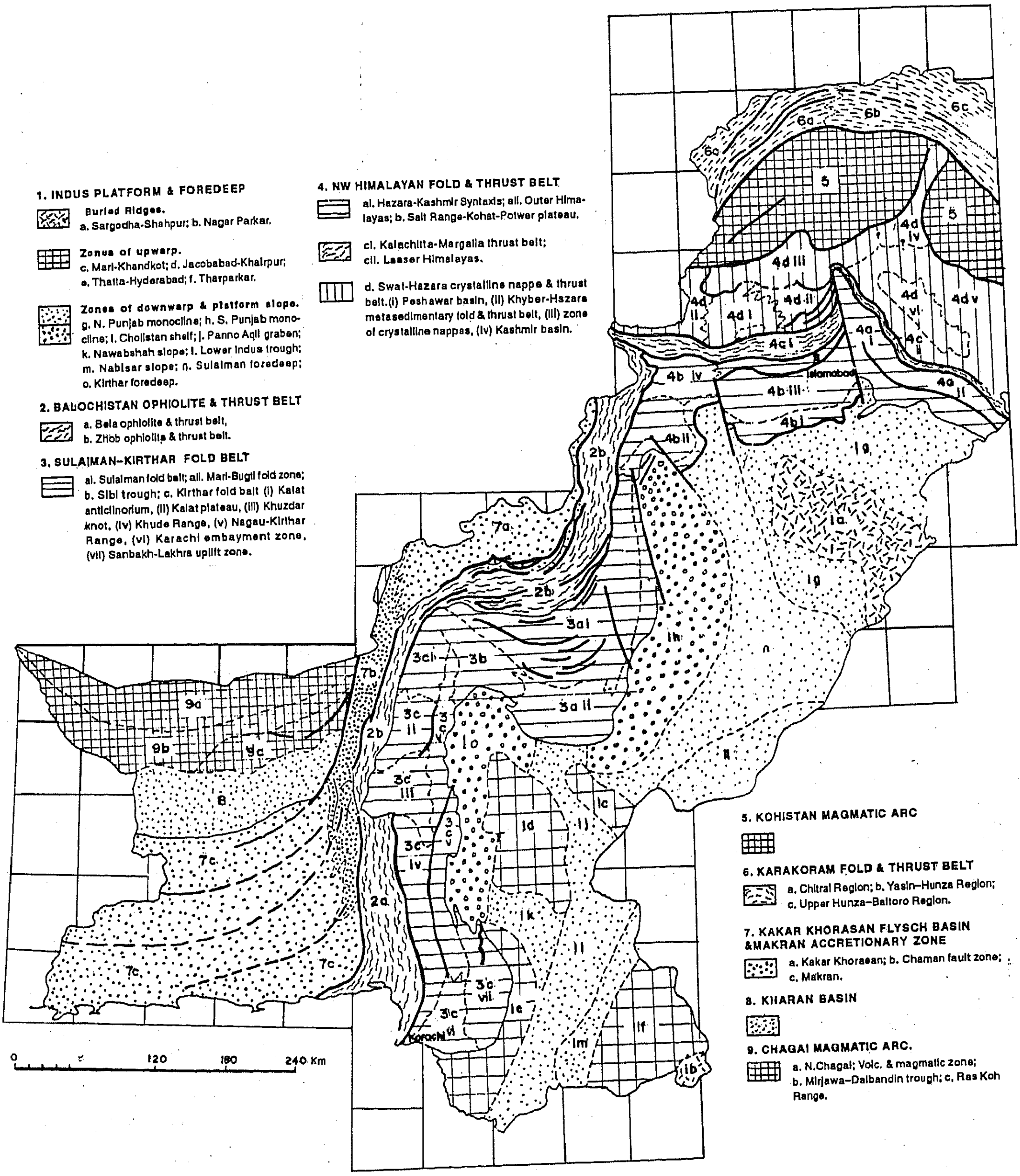
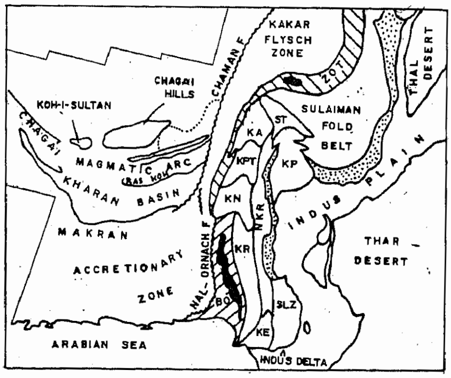


Fig. 4.6. Map showing tectonic zones of Pakistan.

Photo 16. Landsat composite photograph of southern part of Pakistan (Sind and Balochistan), showing physical and geological features (see Figs. 2.2 and 4.6). BO–Bela ophiolite and thrust belt, KA–Kalat anticlinorium, KE–Karachi embayment, KN–Karachi knot, KP–Kachhi Plain, KPT–Kalat Plateau, KR–Khude Range fold belt, NKR–Nagau-Kirthar Range fold belt, SLZ–Sanbakh-Lakhra uplift zone, ST–Sibi trough, ZOT–Zhob ophiolite and thrust belt. Ophiolites in solid black, Indus Piedmont zone shown dotted.



Since 55 m.y. ago, India has steadily rotated counterclockwise (Fig. 4.5). Coupled with Arabia's separation from Africa about 20 m.y. ago, this rotation caused convergence in Balochistan, closure of some of the smaller basins (Seistan, Katawaz), collision of various crustal blocks in Iran-Afghanistan region and formation of the Balochistan fold-and-thrust belt.

The India-Eurasia collision produced the spectacular Himalayas along uplifted and deformed 2,500 km long Indo-Pakistan plate margin. The timing of the collision is constrained by extensive emplacement of ophiolites along the Indus-Tsango Suture Zone, in Waziristan, Zhob Valley and Lasbela area. Along suture zones the youngest marine sediments are lower Eocene. At places Indus molasse overlies Early Eocene limestone and contains debris both of Eurasian and Indian terrains. Compressional tectonic began after deposition of Early Eocene Nummulitic limestone.

TECTONIC ZONES

The geological setting of the regions surrounding Pakistan has been reviewed in the previous chapter and the two broad geological divisions of this region, the Gondwanian and Tethyan Domains have been discussed. In this scenario Pakistan is unique inasmuch as it is located at the junction of these two diverse domains. The southeastern part of Pakistan belongs to Gondwanian Domain and is sustained by the Indo-Pakistan crustal plate. The northern most and western regions of Pakistan fall in Tethyan Domain and present a complicated geology and complex crustal structure. On the basis of plate tectonic features, geological structure, orogenic history (age and nature of deformation, magmatism and metamorphism) and lithofacies, Pakistan may be subdivided into the following broad tectonic zones (Fig. 4.6 and Photo. 16):–

–Indus platform and foredeep.
–East Balochistan fold-and-thrust belt.
–Northwest Himalayan fold-and-thrust belt.
–Kohistan-Ladakh magmatic arc.
–Karakoram block.
–Kakar Khorasan flysch basin and Makran accretionary zone.
–Chagai magmatic arc.
–Pakistan offshore.

Within these broad tectonic zones there are subtle differences in tectonics and changes in structure style to merit further subdivision into smaller tectonic units as shown in Figure 4.6 and also discussed in the following pages.

INDUS PLATFORM AND FOREDEEP

This zone extends over an area exceeding 250,000 km² in southeastern Pakistan and includes the Indus Plain and Thar-Cholistan Deserts. It hosts more than 80% of Pakistan's population, extensive coal deposits, valuable oil and gas fields, potential for geothermal energy and vast groundwater reservoirs.

Gravity and seismic surveys, supported by limited bore hole data, indicate that in the eastern part Precambrian rocks form a gentle westward dipping monocline covered by a veneer of Mesozoic to Cenozoic marine to deltaic sediments. However, there are broad zones of upwarp and downwarp which are well defined by gravity surveys (Fig. 4.7). The sedimentary cover is relatively thin in the upwarp zones. The downwarps contain a thick sedimentary pile, particularly the foredeeps at the western edge of the platform slope where the sedimentary cover is up to 10,000 m thick (Figs. 3.5 and 4.8).

Structural Zones

The Indus platform and foredeep comprise the following main structural zones (Kazmi and Rana 1982, Balakrishnan 1977):

Buried ridges	Sargodha-Shahpur ridge Nagar Parkar ridge
Zones of upwarp	Mari-Khandkot high Jacobabad-Khairpur high Thatta-Hyderabad high Tharparkar high
Zones of downwarp and platform slope	Northern Panjab monocline Southern Panjab monocline Cholistan shelf Panno Aqil graben Nawabshah slope Lower Indus trough Nabisar slope
Foredeeps	Sulaiman foredeep Kirthar foredeep

2.1

These structural zones are shown in Figure 4.6. It may be noted that the Jacobabad–Khairpur upwarp divides the Indus platform into two segments. The lower segment is comprised of the Lower Indus trough. It is bounded by Nawabshah and Nabisar slopes which are in turn flanked by Thatta–Hyderabad and Tharparkar highs. The upper segment, in Panjab, is traversed by Sargodha-Shahpur ridge, splitting it into northern Panjab (Khushab-Gujranwala) monocline and southern Panjab (Mianwali-Bahawalpur) monocline and Cholistan shelf. Westward the Indus platform sharply steepens to form the Sulaiman and Kirthar foredeeps.

Basement Rocks and Structures

Precambrian basement rocks (Kirana Group), crop out in the form of small monadnocks in the Sargodha-Shahkot region. These are the exposed summits of the buried Sargodha-Shahkot ridge (Fig. 4.9) and are largely comprised of metasediments (phyllites, quartzites) and metavolcanics. Some of these rocks have given an isochron age of 870 ± 40 Ma (Davies and Crawford 1971).

Precambrian igneous rocks (Nagar Parkar Igneous Complex) crop out in the form of small scattered hillocks in Nagar Parkar area of Thar Desert. In the eastern part these rocks lie at shallow depths and gently slope westward (Figs. 4.8 and 4.9). The Nagar Parkar Igneous Complex and Kirana Group have been described in Chapters 5, 6, and 7.

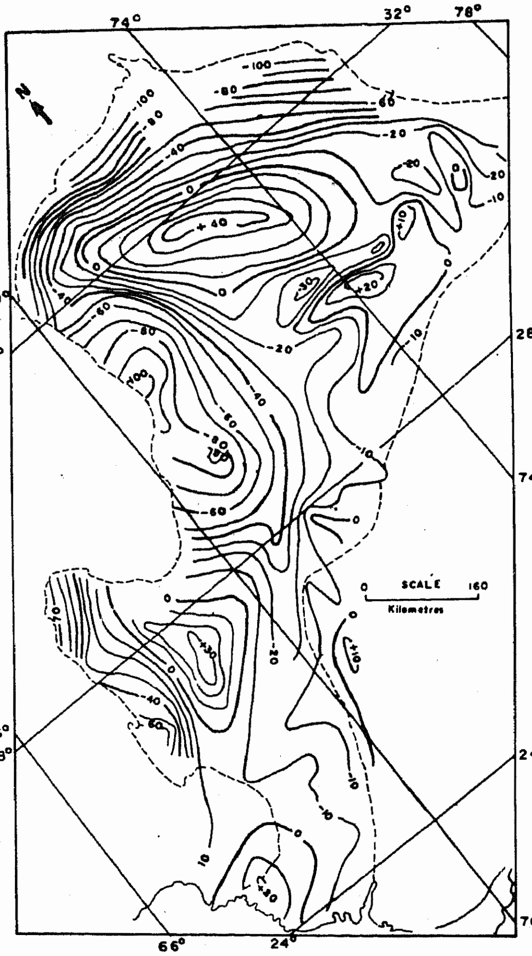


Fig. 4.7. Gravity map of the Indus Plain, showing modified topo anomalies in milligals (from Glennie 1955).

Geophysical surveys (Balakrishnan 1977, Farah et al. 1977, Seeber et al. 1980, Malik et al. 1988) and remote sensing studies (Kazmi 1979a, Kazmi and Rana 1982) indicate that the basement is extensively traversed by NNE to NE, NNW and E-W trending faults (Fig. 4.10). The NE oriented faults conform to the Aravalli–Delhi trend whereas the NNW faults are parallel to Jhelum, Kalabagh, Chaudhwan, Kingri and Mach strike slip faults (Kazmi and Rana 1982). Some of the basement faults straddle major structures such as the Sargodha-Shahpur Ridge (Farah et al. 1977) and Khandkot–Mari and Jacobabad–Khairpur horsts (Ahmed et al. 1991a). Other documented and prominent Faults are the Cutch and Talhar Faults (Kazmi 1979a, Kazmi and Rana 1982), Thar Fault (Fasset et al. 1994), and the faults in Punjab Seismic Zone (Seeber et al. 1980). Most of these basement faults are apparently Late Cretaceous and affect Mesozoic and earlier rocks. In Sindh they are largely concealed beneath the Deccan Trap basalts and

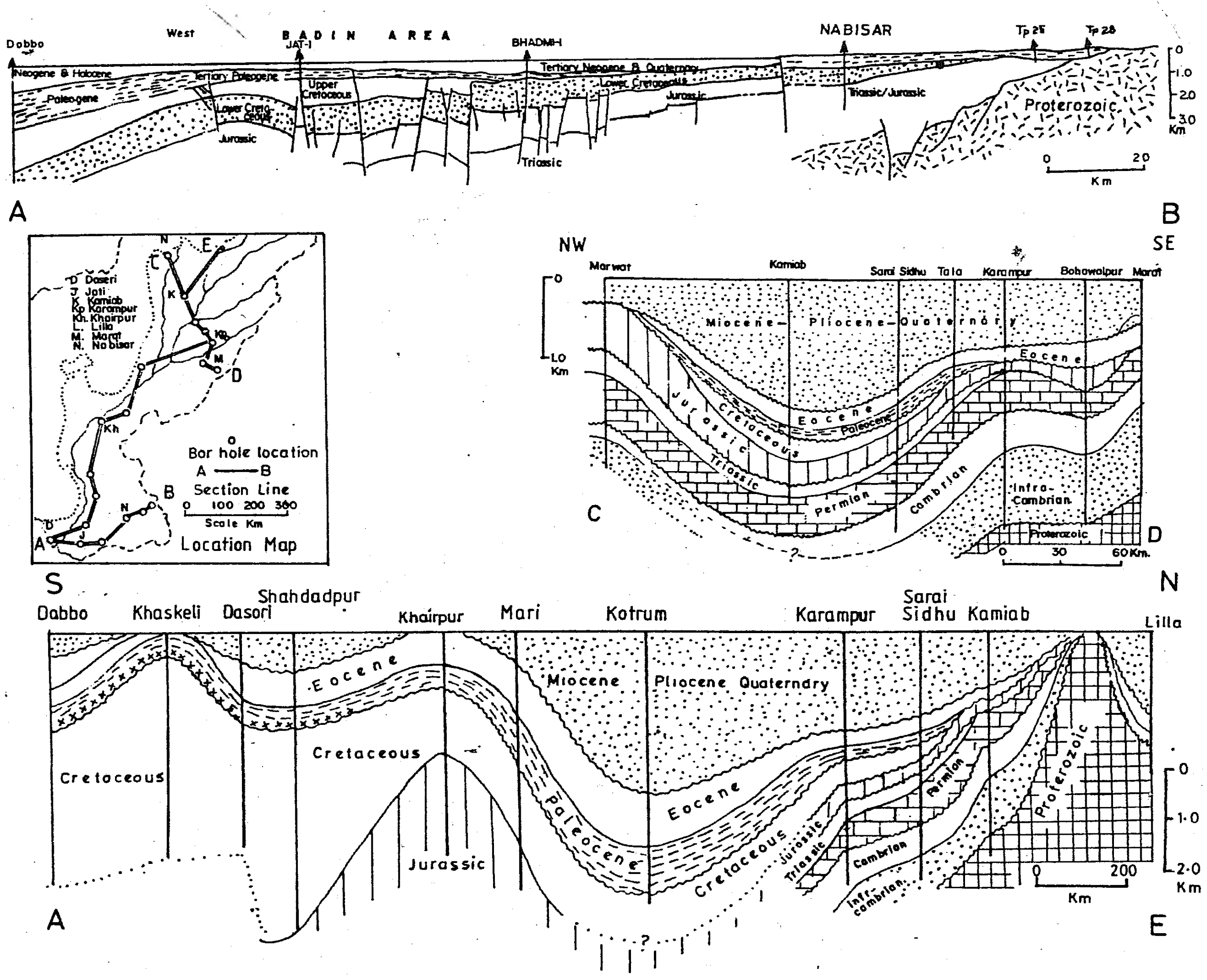


Fig. 4.8. Sections across the Indus Platform showing broader structures and stratigraphy. Based on information from Bakr et al. 1964, Quadri et al. 1986, Malik et al. 1988, Raza et al. 1989a, 1989b, Jadoon 1991, Fasset et al. 1994.

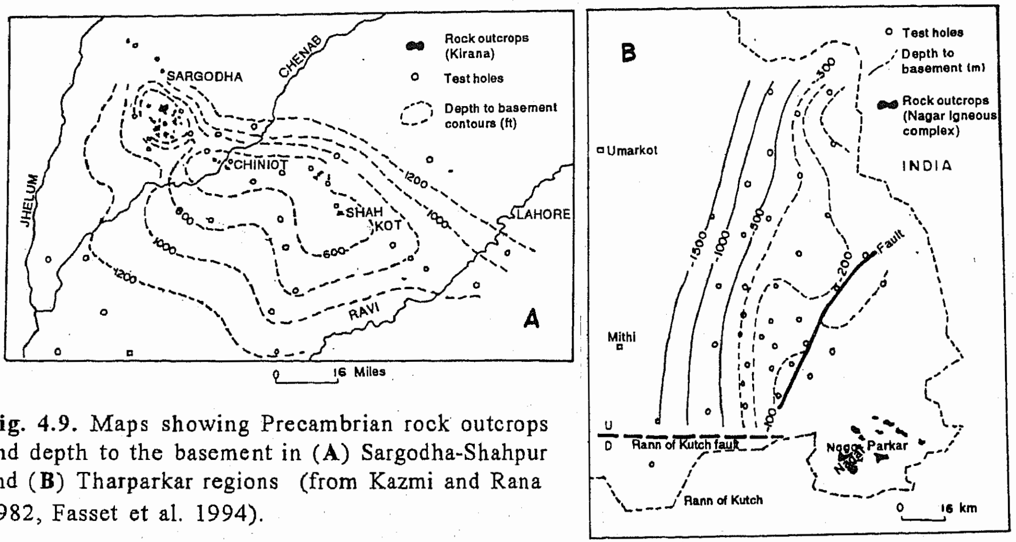


Fig. 4.9. Maps showing Precambrian rock outcrops and depth to the basement in (**A**) Sargodha-Shahpur and (**B**) Tharparkar regions (from Kazmi and Rana 1982, Fasset et al. 1994).

younger sedimentary rocks (Fig. 4.8). Subsurface seismic profiles indicate that in the foredeep zone and in Jacobabad–Khandkot zone some of the faults are likely to be post Eocene (Oligocene ?) and post Miocene (Raza et al. 1990a, Ahmed et al. 1992). Satellite photographs reveal an extensive network of lineaments (Fig. 4.10) and some of these are likely to be traces of buried basement faults. Such lineaments are a common feature in other major shield areas of the world (Wise 1968, Withington 1973).

Most of the Late Cretaceous basement faults and rift structures are likely to be associated with detachment of the Indian plate from Gondwanaland, whereas the Tertiary faults may have resulted from bending of the crustal plate due to collision and rebound relief tension or compression release. Seismic studies and fault plane solutions indicate that some of these faults are extensional features, while others are strike slip faults (Seeber et al. 1979). The Cutch Fault is an active fault that caused the devastating 1819 earthquake. The Panjab Seismic Zone is also comprised of buried active faults. Along the fold belt region, the western margin of the foredeep is seismically active and contains a number of active faults (Kazmi 1979a, 1979c).

Sedimentary Cover

The Indus platform and foredeeps are covered by unconsolidated Quaternary deposits with a maximum thickness of up to about 500 m. They constitute a vast groundwater reservoir and have been described in Chapter 5. They are underlain by Siwalik molasse in northern and western part of the platform and in the foredeep region. South of Sargodha-Shahpur ridge and extending up to Khandkot–Mari high, in a roughly triangular area, the Quaternary deposits are underlain by post-Eocene, largely fluviatile deposits (Nari and Murree Formations). In southern part of the platform, east of Indus, the Quaternary deposits are underlain by Paleogene marine and deltaic sedimentary rocks (Table 4.1). This shallow Cenozoic sequence is likely to contain substantial secondary groundwater aquifers in non-marine facies and vast deposits of coal have been already proved in Lower Paleogene sequence of Sindh (Chapter 9).

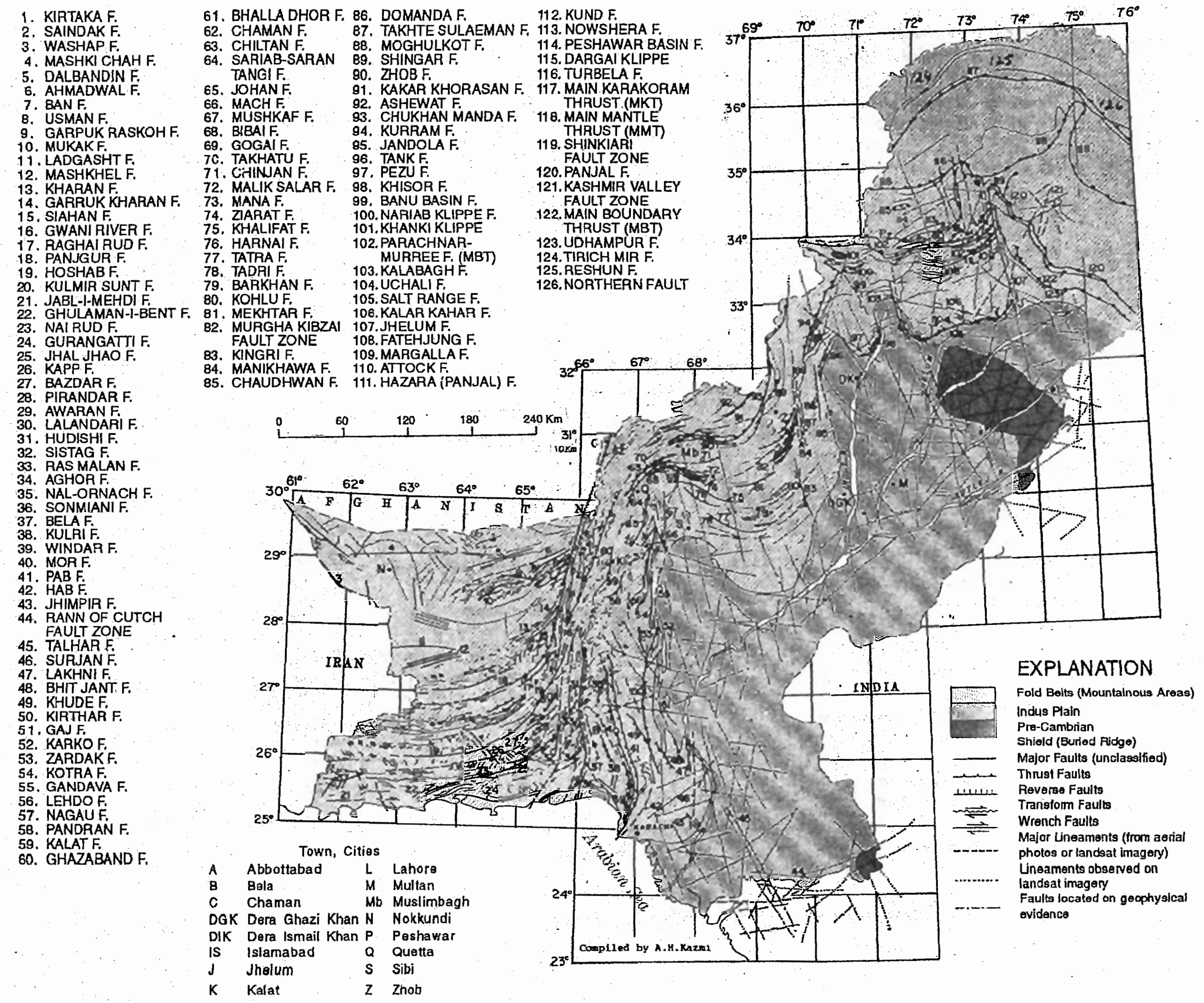


Fig. 4.10. Map showing major faults and lineaments in Pakistan (from Kazmi and Rana 1982).

Table 4.1. Stratigraphic sequence in the Indus Platform and Foredeep Zone. (Modified from Quadri et al. 1986, Raza et al. 1989b, Ahmed et al. 1991 and Shah 1977).

ERA	PERIOD	PUNJAB PLATFORM SLOPE AND SULAIMAN FOREDEEP		JACOBABAD-KHANDKOT HIGH	LOWER INDUS TROUGH AND INDUS DELTA	
		Platform	Foredeep		Foredee[p	Delta
CENOZOIC	Neogene	Siwalik Gr.	Siwalik Gr. Gaj Fm.			Siwalik Gr. Gaj Fm.
	Oligocene		Nari Fm.			Nari Fm.
	Eocene	Sakesar Lst. Nammal Fm.	Kirthar Fm. Ghazij Fm.	Kirthar Fm. Ghazij Fm.	Laki Fm.	Kirthar Fm. Laki Fm.
	Paleocene	Patala Fm.	Dungan Fm. Khadro Fm.	Dungan Fm.	Ranikot Gr.	Ranikot Gr.
MESOZOIC	Cretaceous		Pab sandstone Mughal Kot Fm. Parh Fm. Guru Fm. Sembar Fm.	Pab sandstone Mughal Kot Fm. Parh Fm. Guru Fm. Sembar Fm.	Parh Fm. Guru Fm. Sembar Fm.	Pab sandstone Mughal Kot Fm. Parh Fm. Guru Fm. Sembar Fm.
	Jurassic	Samana Suk Fm. Shinwari Fm. Datta Fm.	Chiltan Fm. Shirinab Fm.		Chiltan Fm. Shirinab Fm.	
	Triassic	Kingriali Fm. Tredian Fm.	Wulgai Fm.		Wulgai Fm.	
Paleozoic	Permian	Zaluch Gr. Nilawahan Gr.				
	Cambrian	Jhelum Gr. Salt Range Fm.				
pC	Precambrian	Kirana Fm.				

The Precambrian and Paleozoic sequence is restricted to Panjab part of the platform and the Sulaiman foredeep. It thins out in Jacobabad-Khairpur region and has not been encountered in deep wells in Sindh area. The general stratigraphy of Indus platform and foredeep is shown in Figure 4.8 and Table 4.1. The cover is thinnest in the region of Sargodha-Shahpur Ridge (0–600 m), Jacobabad–Khairpur high (up to 4,000 m), Tharparkar high (10–200 m) and eastern margin of the platform. In other parts (platform slopes and downwarp zones) the sedimentary cover is fairly thick and in foredeep zones it attains a thickness of 10,000 m to 15,000 m (Raza et al. 1989c, 1990a).

Structurally the sedimentary cover of Indus platform and foredeep is comprised of several large, gently dipping anticlinal flexures and fault blocks. Oil and gas reserves have been found in some of these structures. In the Panjab part of the platform gas reserves have been found in Mesozoic sandstones at Nandapur and Panjpir. In Khandkot–Mari horst gas occurs in Eocene carbonate reservoirs. The Jacobabad–Khairpur horst contains gas in Eocene carbonates and in Jurassic and Triassic sandstones, and in Badin region of Sindh oil and gas occur in sandstones of Cretaceous Goru Formation in structural traps (Malik et al. 1988, Raza et al. 1989b).

North and west of Indus platform and the foredeeps, the sedimentary cover has been deformed intensely by collision of the Indo-Pakistan plate with Eurasia. It forms a broad and extensive fold-and-thrust belt characterised by fold festoons with arcuate or sinuous fold axes and intervening sharp structural flextures– the "syntaxes" (discussed in Chapter 2). The northern and western margins of this zone are characterised by sutures and obducted masses of ophiolite, whereas the southern and eastern part of the zone, on the other hand, is comprised of contorted fold and thrust belts (Fig. 4.6). 2-1

The marginal fold-and-thrust belt is divided in two main parts, the Himalayan fold-and-thrust-belt in the north, and the East Balochistan fold-and-thrust belt in the south. These two zones differ considerably in their basement characteristics, structure and stratigraphy, as may be seen from the following descriptions.

Table 4.2. Generalised stratigraphic sequence in Sulaiman-Kirthar Fold Belt.

	KIRTHAR RANGE	QUETTA SYNTAXIS	SULAIMAN RANGE
Neogene	Siwalik Group	Siwalik Group	Siwalik Group
	xxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxx	xxxxxxxxxxxxxxxxxxxxxxx	xxxxxxxxxxxxxxxxxxxxxxxxx
	Gaj Formation	Gaj Formation	
Paleogene	Nari Formation	xxxxxxxxxxxxxxxxxxxxxxx	Chitarwata Formation
	Kirthar Formation	Kirthar Formation	Kirthar Formation
	Marap Conglm. Ghazij Group Laki Formation Meting Limestone	Ghazij Group	Ghazij Group
	xxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxx	xxxxxxxxxxxxxxxxxxxxxxx	xxxxxxxxxxxxxxxxxxxx
	Lakhra Fm. Bara Fm. Khadro Fm. Ranikot Fm. Dungan Fm	Dungan Fm.	Dungan Formation Khadro Formation
	xxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxx	xxxxxxxxxxxxxxxxxxxxxxx	xxxxxxxxxxxxxxxxxxxx
Cretaceous	Pab Sandstone		Pab Sandstone
	Ft. Monro Limestone	Ft. Monro Limestone	Bibai Volc. Ft. Monro Limestone
	Mughal Kot Formation	Bibai Volcanics	Mughal Kot Fm.
	Parh Limestone	xxxxxxxxxxxxxxxxxxxxxxx	
	Goru Formation	Parh Limestone	Parh Limestone
	Sembar Formation	Goru Formation	Goru Formation
			Sembar Formation
Jura.	xxxxxxxxxxxxxxxxxxxxx	xxxxxxxxxxxxxxxxxxxxxxx	xxxxxxxxxxxxxxxxxxxx
		Chiltan Formation	Chiltan Fm. Mazar Drik Fm.
	Shirinab Formation	Shirinab Formation	Shirinab Formation
Trias.		Wulgai Formation	Wulgai Formation

(xxxx lines indicate unconformity).

EAST BALOCHISTAN FOLD-AND-THRUST BELT

This zone of folds and thrusts is 60 to 150 km wide, with a strike length of about 1,250 km. It extends southward from Waziristan, through Loralai–Bugti area, around the Quetta syntaxis, down south to Karachi and the Indus Delta (Fig. 4.6, Photos. 12 and 13). The Sulaiman–Kirthar foredeeps lie on its eastern flank, whereas on the west it is truncated by Ornach–Nal Transform Fault and Ghazaband Fault. Its north-south structures are in fault contact with the east-west trending Himalayan fold-and-thrust belt (Fig. 4.6).

The East Balochistan fold-and-thrust belt is the product of transpression and oblique collision of India-Pakistan plate with the Afghan block. This belt is reportedly underlain by relatively thinner transitional or oceanic crust at least in its northern part (Jadoon et al. 1989, 1996, Khurshid et al. 1992). The basement is covered by a 10 km thick Phanerozoic sedimentary wedge (Raza et al. 1990a). Towards the hinterland this zone consists of a thrust belt characterised by schuppen structures, nappes and klippes of ophiolites and melanges. Inward there is a broad folded zone with narrow anticlinal hills and wide synclinal valleys. Magmatism is largely confined to basic dykes and sills in the Mesozoic sequence. Metamorphism is restricted to ophiolite bearing sections of the thrust belt (see Chapters 6 and 7).

The East Balochistan fold-and-thrust belt contains more than 10 km thick sequence of Permian to Mesozoic pericratonic shelf carbonates, neritic shales and volcanics, Paleogene shallow-water interlayered marine and continental deposits and Neogene molasse. The general stratigraphy of this belt has been shown in Table 4.2 and the various lithostratigraphic units have been discussed in Chapter 5. There is an intimate relationship between the structure and stratigraphy of this belt. In fact, since the India-Eurasia collision, the deposition centres and the deformation have gradually migrated south and eastward towards the foreland. The tectonostratigraphic zones shown in Figure 4.6 stand out sharply not only on the geological and tectonic map of Pakistan (Kazmi and Rana 1982, Qureshi et al. 1993) but are also clearly discernible on satellite photographs.

Towards the west, the outer part of the East Balochistan fold-and-thrust belt is comprised of an over 550 km long and 20 to 40 km wide imbricate zone of thrusts and nappes, with melanges (Fig. 4.6). This belt is in two parts, the Zhob ophiolite-and-thrust belt in the north and the Bela ophiolite-and-thrust belt in the south (Kazmi and Rana 1982). These zones were earlier included in their Central and Bela geanticlines by HSC (1960).

Bela Ophiolite-and-Thrust Belt

This belt extends northward for about 320 km from the sea coast and abruptly terminates against the Khuzdar knot near Khuzdar. Bounded to the east and west by Pab and Ornach–Nal Faults respectively, this belt is largely comprised of a thick, broken and faulted sequence of Early Mesozoic to Neogene rocks as shown in Figure 4.11. As may be seen in this figure, above the Sembar/Parh Formation there is a significant stratigraphic break indicated by erosion, lava flow and volcanic boulder conglomerates overlying Bad Kachu, Rattaro and Thar Formations (HSC 1960). The Kanar Melange (Sarwar et al. 1984, Sarwar 1992) and Bela Ophiolites (DeJong and Subhani 1979) also tectonically overlie Sembar and Parh Formations. Unlike other rocks, the Shirinab Formation has been intensely deformed and is characterised by tight and steep isoclinal folds derived from pre-Himalayan deformation. The Kanar Melange has been thrust over

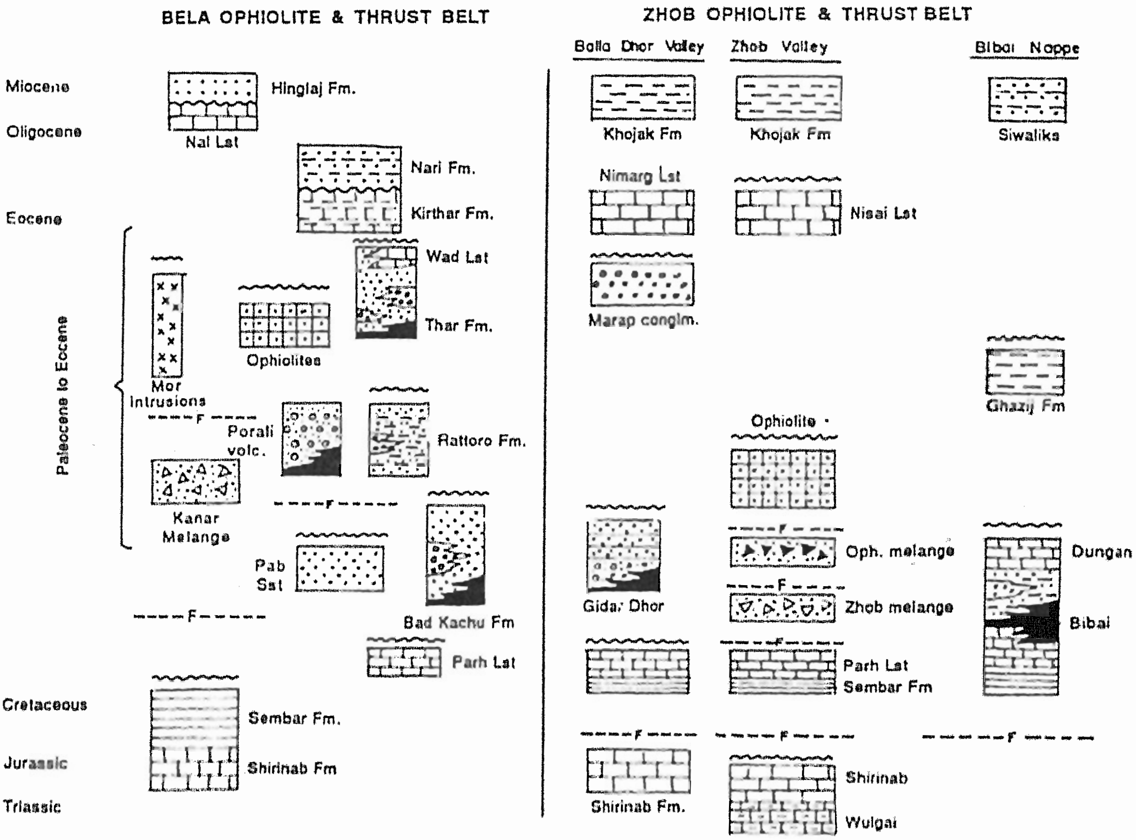


Fig. 4.11. Tectonostratigraphic sequence in the Bela-Zhob Ophiolite-and-Thrust Belt. (Based on data from HSC 1960, Kazmi 1979, Sarwar et al. 1984).

the Sembar Formation (Fig. 6.11) and is composed of clasts derived both from the oceanic crust as well as from the continental rocks (see Chapter 6). It contains huge blocks of ophiolites, Shirinab and Sembar which are hundreds of metres in dimension. The Shirinab blocks show tight, disharmonic folds. The ophiolite suite in the melange includes scattered fragments and blocks of serpentinite, diabase, gabbro and massive or pillowed basalt. Exotic clasts of Porali Volcanic Conglomerate, biogenic limestone, schist, brachiopod-bearing limestone, grey and pink granite (similar to the basement related pink granites of Nagar Parkar) also occur in Kanar Melange (Sarwar et al. 1984, Sarwar 1992). According to DeJong and Subhani (1979) the whole Bela ophiolite belt may be considered as a 'megamelange' because large slabs of this belt are interspersed with Kanar and Kunno Melanges, and other melange types also occur within this belt.

The Porali Volcanic Conglomerate (Sarwar et al. 1984) was previously referred to as Porali agglomerate (DeJong and Subhani 1979). It is comprised of clasts of alkali basalt, alkaline ultramafics and sedimentary rock. It occurs as a tectonic wedge. It is thought to be part of a volcanic island arc sequence on continental crust. The Mor intrusives form sills and dykes in the Shirinab and Sembar Formations and their petrology and geochemistry is similar to the Porali Volcanic clasts (Sarwar et al. 1984). The Porali Volcanics and Bad Kachu, Rattaro and Thar Formations are coeval and similar in many respects and probably formed around scattered volcanic centres.

The Bela Ophiolites (DeJong and Subhani 1979, Sarwar 1992) are largely comprised of a 3 to 5 km thick sequence of pillow basalt, inter-flow cherts, argillites, limestones, and diabase-gabbro sills (see Chapter 6). Small bodies of granites and oceanic plagiogranites occur as intrusive bodies within the ophiolites near Khuzdar (Ahmed 1993). The age of the ophiolites, as indicated by microfauna is Aptian-Early Maestrichtian (Sarwar 1992). Because the Eocene limestone unconformably overlies the ophiolites and the latter do not contain any sedimentary lenses or blocks younger than Late Maestrichtian, Alleman (1979) considers that the ophiolites were emplaced between the Paleocene and Lower Eocene. U-Pb data on zircons from granites and plagiogranites indicates 65 ± 1 Ma as the crystallisation age for these rocks (Ahmed 1993).

Geological map and sections in Figures 4.12 and 6.11 show the general structure of Bela ophiolite-and-thrust belt. Apart from nappes and thrust slices of ophiolites and melanges, this zone has been intensely faulted. Reverse, normal and tear faults are common and the region is characterised by short, tight, steeply-dipping or isoclinal, complex folds. Adjacent fold axes have been commonly offset or truncated by faults. In the eastern part Eocene Kirthar limestone and shale and Oligocene Nari sandstone and shale unconformably overlie Mesozoic to Paleocene rocks. Westward the latter rocks are overlain by Oligocene Nal limestone and Miocene Hinglaj Sandstone.

The Bela ophiolite-and-thrust belt has been apparently, a region of intermittent deformation, magmatism, uplift, erosion and sedimentation since the Early Mesozoic (Kazmi and Rana 1982). Continental collision and Himalayan Orogeny further accentuated the tectonostratigraphic features that had a pre-orogenic beginning. Presence of two different types of basalts in the Bela belt (tholeiitic and continental basalts of the Porali Volcanics) indicate that the ophiolite emplacement was preceded by volcanism on continental margin which may have been caused by hot spot activity, prolonged effects of Gondwana rifting or subduction related plate bending stresses (Sarwar 1992). According to Sarwar and DeJong (1984), the Bela Ophiolites originated along a leaky transform fault or fracture zone, which was destroyed as the transform movement gave way to oblique convergence, resulting in ophiolite emplacement (Fig. 4.13).

Zhob Ophiolite-and-Thrust Belt

The southwestern portion of the Zhob ophiolite-and-thrust belt lies only about 20 to 40 km north of the northern edge of the Bela ophiolite-and-thrust belt (Fig. 4.6). Neogene to Recent deposits cover the connection between the two belts. It is noteworthy that blocks of ophiolites and volcanics occur west of the Ornach–Nal Fault. This fault has been traced up to Nal. About 20 km northwest of Nal, there is a prominent NNW -trending fault, the Hund River Fault that links up with southern extension of the Zhob belt (Fig. 4.10). A NNE- trending fault, the Gidar Dhor Fault, joins Hund River Fault near its southern end. The Gidar Dhor Fault runs along the western margin of the Khuzdar knot and probably links up with Anjira–Gazan Fault (Sarwar et al. 1979), which separates the Kalat Plateau from Khuzdar knot. It is likely that the Ornach–Nal Fault splays out near Hazarganji and its offshoots link up with Hund River and Gidar Dhor Faults. In the absence of concrete geological and geophysical evidence, however, the abrupt northward termination of Ornach–Nal Fault remains an enigma.

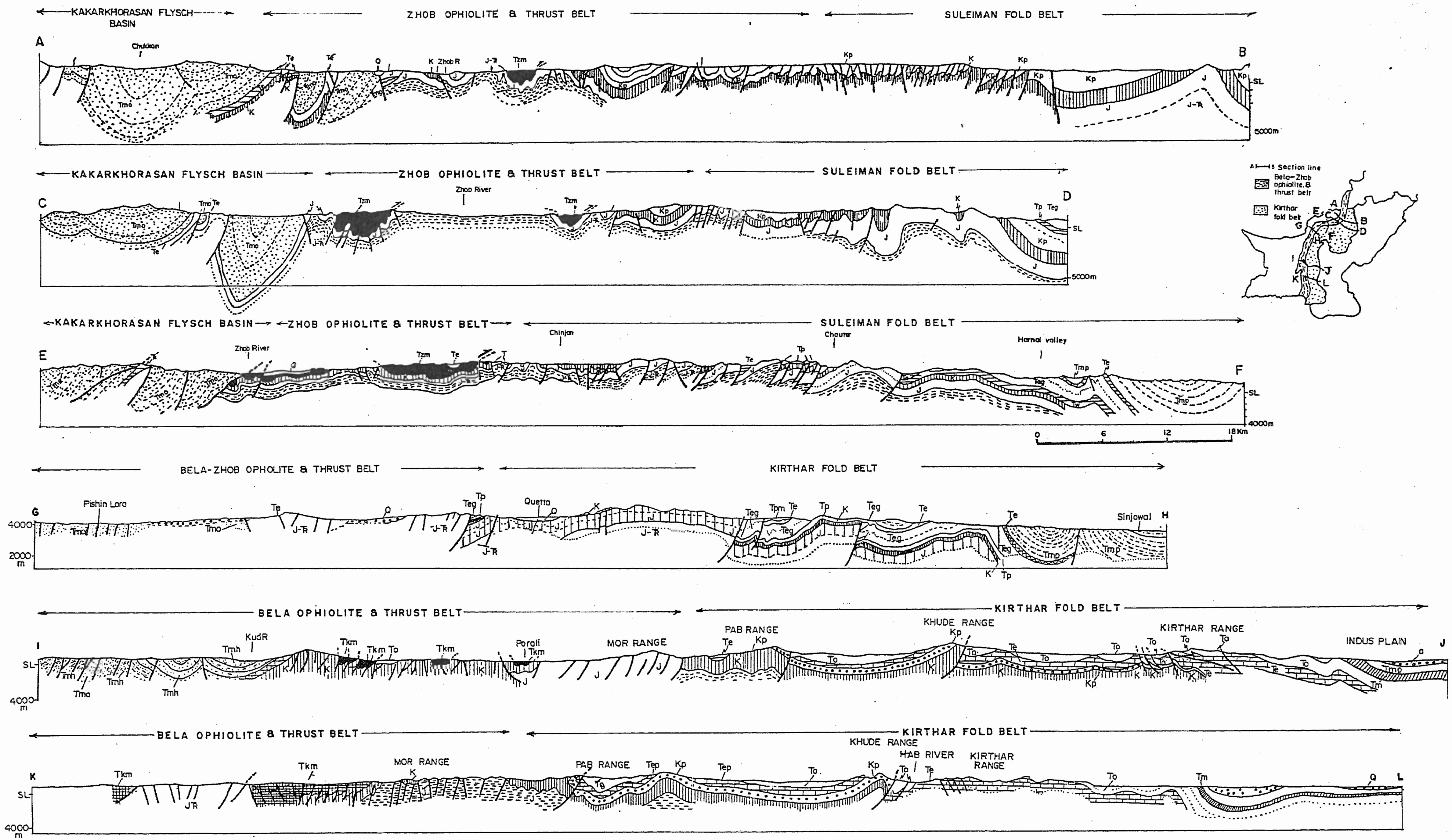


Fig. 4.12. Geological sections across Bela-Zhob Ophiolite-and-Thrust Belt and Sulaiman-Kirthar Fold Belts. J-Ѣ-Wulgai Shirinab Fms; J–Shirinab, Chiltan Mazar Drik Fms; K–Sembar, Goru, Parh Fms; Kp–Pab sst, Ft Monro Lst, Mughalkot Fm; Tzm–Zhob Melange and Ophiolites; Tp–Dungan, Khadro, Ranikot Fms; Tep–Paleocene to Eocene sequence; Teg–Ghazij Fm; Tkm–Kanar Melange and Ophiolites; Te–Kirthar Fm, Nisai Fm; To–Nari Fm; Tm–Gaj Fm; Tmh–Hinglaj Fm; Tmo–Khojak Fm; Tmp–Siwalik Gr; Q–Quaternary deposits. Modified from HSC 1960.

The Zhob and Bhalla Dhor Faults (Fig. 4.6 and 4.10) define the northern and western margins of the Zhob ophiolite-and-thrust belt, whereas a series of thrust faults, namely the Kurram Fault, Murgha Kibzai Fault zone, the Ziarat Fault, Saran Tangi–Sariab–Chiltan Fault and Kalat Fault, demarcate its eastern and southern limits. This belt is largely comprised of small nappes, thrust sheets and fault blocks containing outcrops of Mesozoic to Neogene sediments and obducted masses of melanges and ophiolites (Figs. 4.14, 4.15 and 4.16). From west of Nal, this belt extends northwestwards through Quetta Syntaxis to Zhob and Kurram Agency where it terminates against the Sarobi Fault and the MBT (Fig. 4.6). The general stratigraphic sequence in this belt is shown in Figure 4.11.

The southwestern portion of the Zhob ophiolite-and-thrust belt between Mashkai and Quetta largely consists of thrust blocks and slices of highly deformed Shirinab Formation, faulted against Sembar and Parh Formations. The Shirinab Formation contains tight and complicated folds as in the Bela belt. The Parh limestone is also characterised by tight irregular disharmonic folds, which are commonly overturned or form small patternless domes and basins (HSC 1960). Eastward the Mesozoic sequence is topped by a sharp angular unconformity, overlain by Late Cretaceous to Early Paleocene Gidar Dhor Group, Eocene Marap Conglomerate and Nimarg Limestone (HSC 1960). Westward the Mesozoic rocks are faulted against, or unconformably underlie Oligocene to Miocene Khojak Formation (Cheema et al. 1977). It is noteworthy that the lower part of Gidar Dhor Group contains basaltic lava flows and thick lenses of boulder conglomerate with subrounded to angular fragments of Jurassic limestone, pink granite, lavas, diorite and jasper. They are thus similar to the Bad Kachu, Rattaro and Thar Formations of the Bela belt. The Marap Conglomerate is poorly sorted and contains pebbles and boulders of Shirinab and Parh Formations. It overlies the latter formations with a sharp angular unconformity and passes eastward into the Ghazij Formation (HSC 1960).

Between Quetta and Zhob this belt finds its best and widest expression (Photo. 12) with thick imbricates of Mesozoic rocks, enormous slabs of obducted melanges and ophiolites and small nappes of Cretaceous to Paleogene sedimentary rocks. These tectonites are unconformably overlain by Eocene Ghazij Formation, Eocene Nisai Limestone and Neogene molasse. North and westward the Neogene sequence of the Kakar Khorasan flysch basin is separated from this belt by the Zhob Fault that apparently underlies the Zhob Valley (Figs. 4.6 and 4.10).

In the Zhob belt the Mesozoic sequence is comprised of Triassic Wulgai Formation, Jurassic Shirinab Formation, Cretaceous Sembar, Goru and Parh Formations (Fatmi 1977) and the Bibai Volcanics (Kazmi 1955, 1979b). In this region the Wulgai and Shirinab were earlier mapped as Alozai Group and Loralai Limestone by Hunting Survey Corporation. The Mesozoic sequence has been affected by pre-Himalayan deformation and exhibits tight, regular but complex folds which are often isoclinal and have steep dips (Fig. 4.14).

According to Williams (1959), HSC (1960), Sokolov et al. (1965) and Fatmi (1977), the fossiliferous Triassic sequence in the Zhob area is comprised of about 985 m to 1,180 m thick interbedded shales, marls and limestones with sandstones and conglomerates as minor constituents. Permian fossils have been reported from this sequence by Vredenburg (1904) and Crookshank (quoted in HSC 1960), though none of the recent workers have found Permian fossils in this sequence. Okimura et al. (1988) have, however,

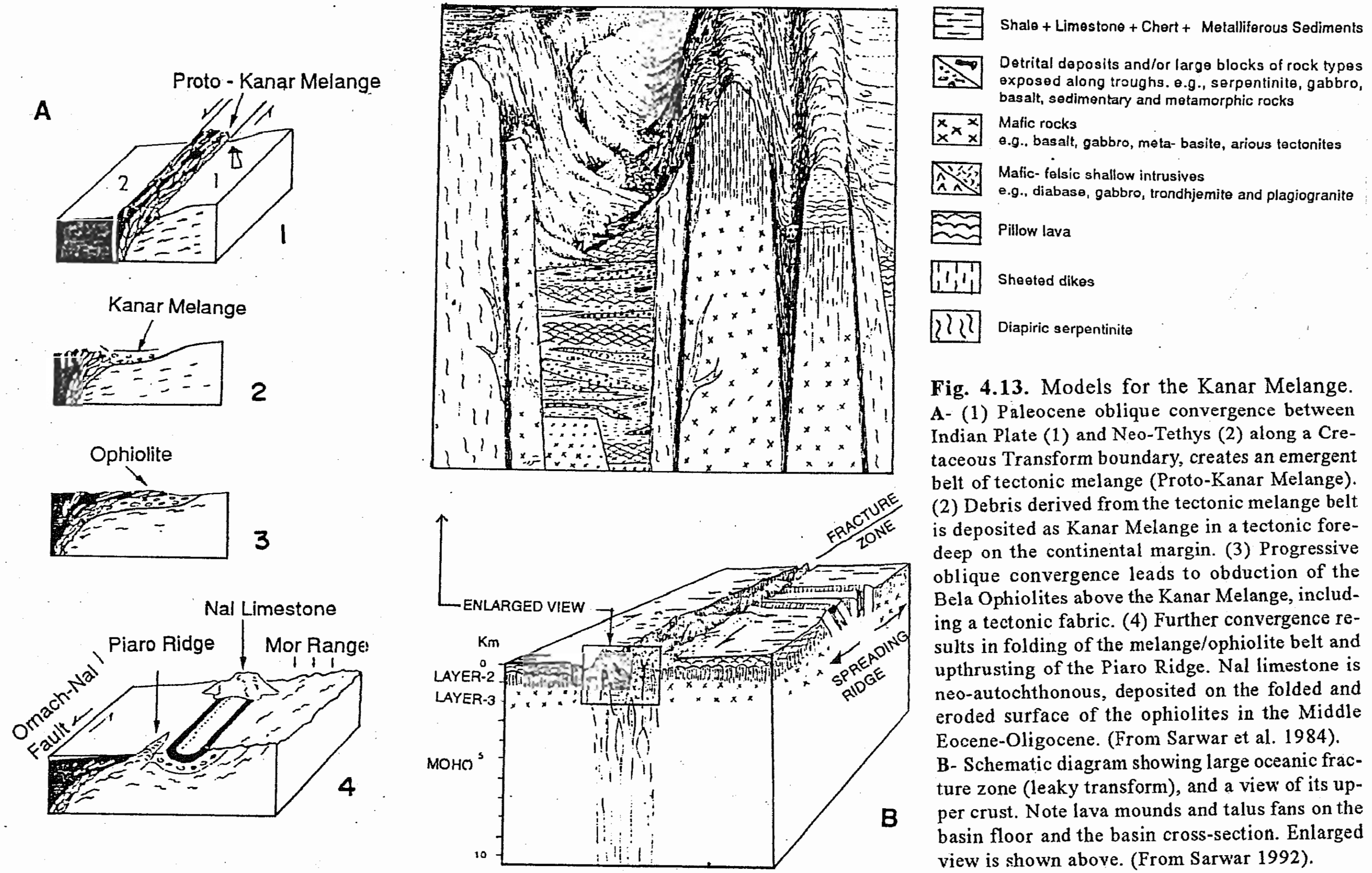


Fig. 4.13. Models for the Kanar Melange. **A**- (1) Paleocene oblique convergence between Indian Plate (1) and Neo-Tethys (2) along a Cretaceous Transform boundary, creates an emergent belt of tectonic melange (Proto-Kanar Melange). (2) Debris derived from the tectonic melange belt is deposited as Kanar Melange in a tectonic foredeep on the continental margin. (3) Progressive oblique convergence leads to obduction of the Bela Ophiolites above the Kanar Melange, including a tectonic fabric. (4) Further convergence results in folding of the melange/ophiolite belt and upthrusting of the Piaro Ridge. Nal limestone is neo-autochthonous, deposited on the folded and eroded surface of the ophiolites in the Middle Eocene-Oligocene. (From Sarwar et al. 1984). **B**- Schematic diagram showing large oceanic fracture zone (leaky transform), and a view of its upper crust. Note lava mounds and talus fans on the basin floor and the basin cross-section. Enlarged view is shown above. (From Sarwar 1992).

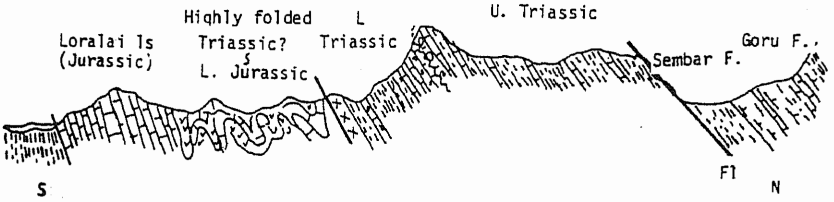
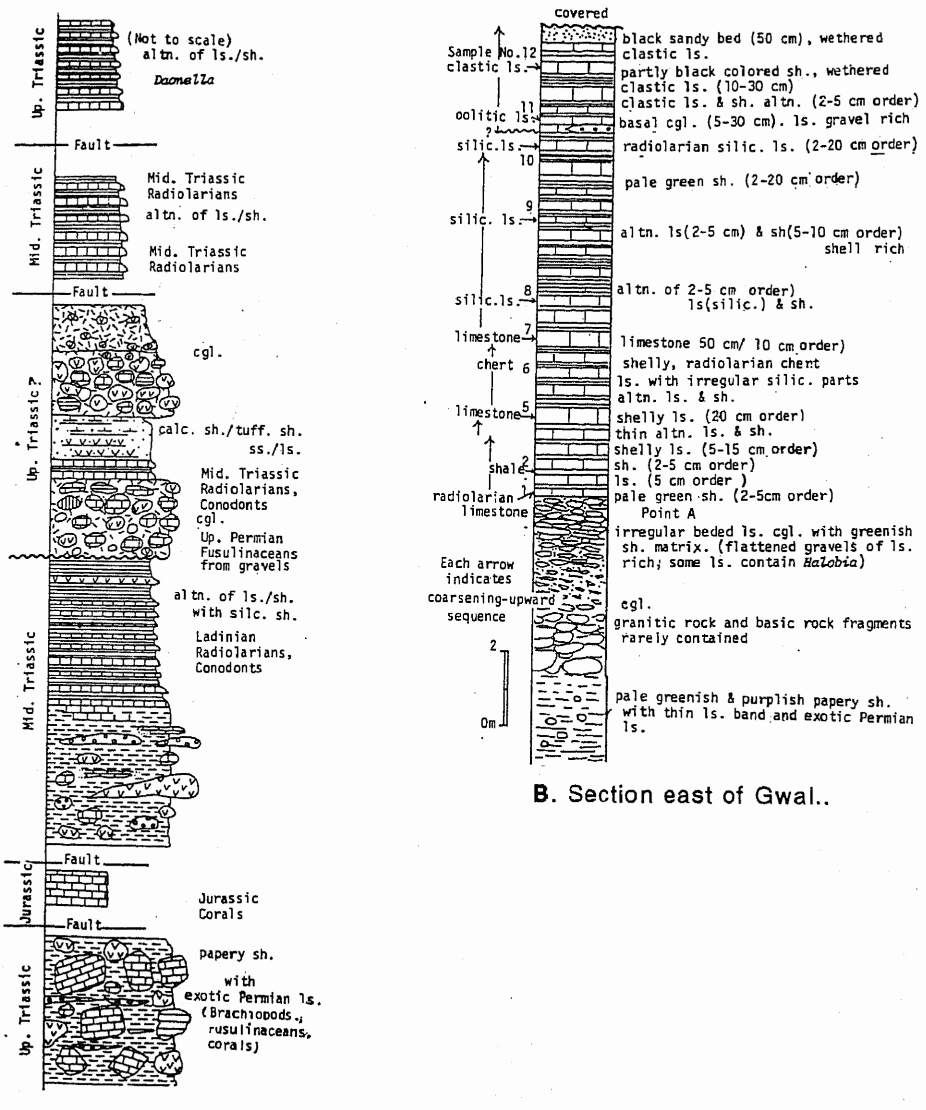


Fig. 4.14. Geological section at Gwal, Zhob Valley. (From Okimura and Fatmi 1989).



B. Section east of Gwal..

A. Wulgai Nala section.

Fig. 4.15. Stratigraphic section showing Triassic-Jurassic tectonostratigraphic sequence near Wulgai and Gwal in the Zhob Valley. (From Okimura and Fatmi).

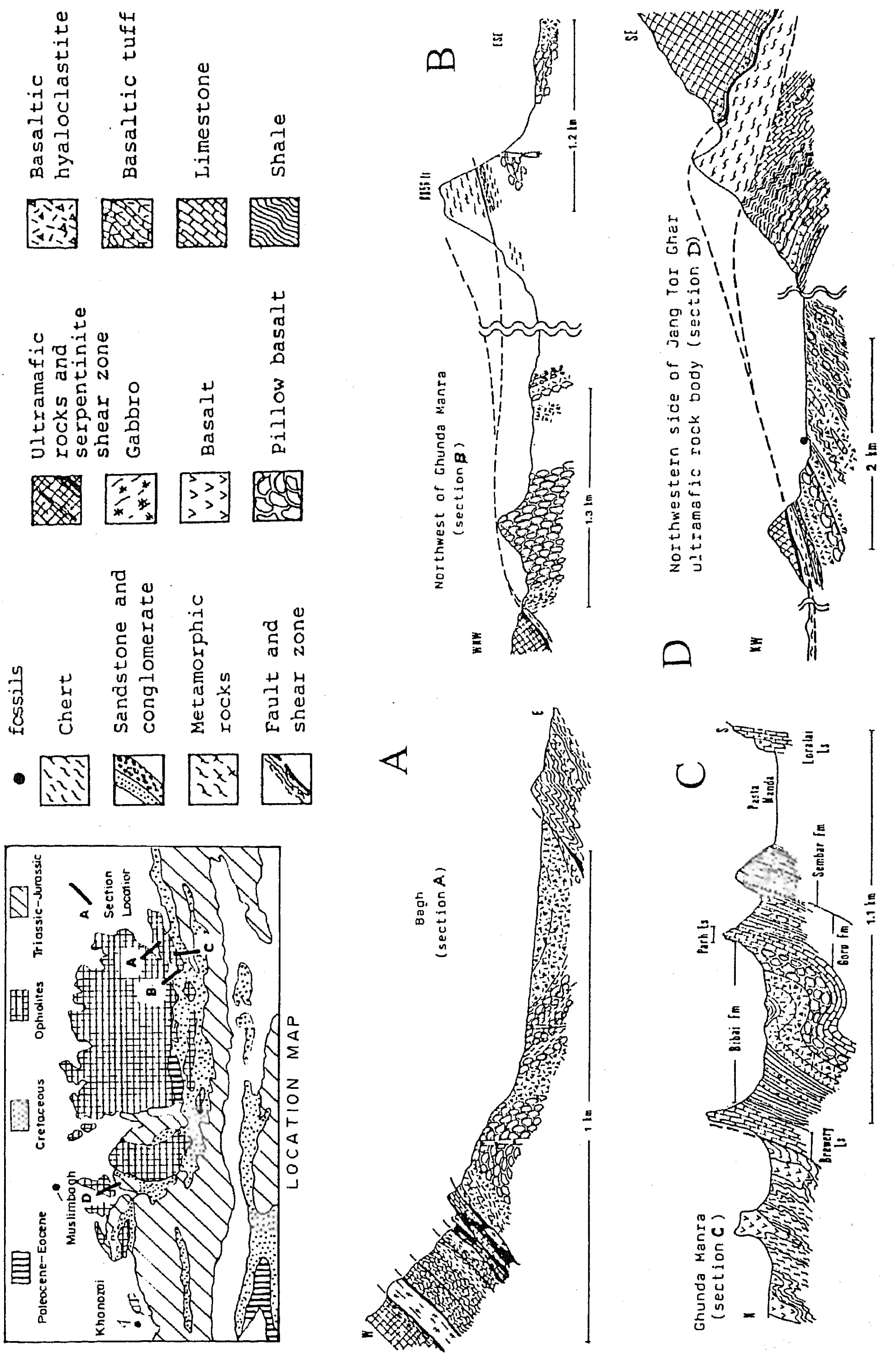


Fig. 4.16. Geological cross-sections through Zhob Ophiolites south of Muslimbagh. (Modified from Otsuki et al. 1988).

obtained Permian brachiopods and fusulinids from limestone olistoliths and limestone clasts in conglomerates embedded in the Triassic sequence of the Gwal–Wulgai area of Zhob. According to these workers, the Triassic rocks are comprised of greenish or purple coloured tuffaceous and siliceous shale, interbedded with thin microlitic or radiolarian limestone. At places they contain beds of boulder conglomerate and olistostromes in the lower part (Fig. 4.15). The olistoliths dominantly comprise marine limestone, quartzose sandstone, and ophiolite. The conglomerates contain occasional granitic and basic rock fragments. Gansser (1979) has reported blocks of radiolarites, microgranite and granodiorite in the ophiolite melange in the Khanozai area.

The Zhob Ophiolites occur as tectonic klippe and thrust sheets overlying the Mesozoic sequence or as thrust slices within the imbricated Mesozoic blocks. The melange contains clasts with Maestrichtian fossils and is overlain by Eocene limestone. It is inferred that they were obducted during Paleocene and Early Eocene (Alleman 1979, Okimura et al. 1988). Small scattered slivers of ultrabasic rocks, gabbros, diorites and volcanics occur in the Mesozoic sequence between Mashkai and Surab (HSC 1960). However large bodies of ophiolites occur only in the Zhob–Waziristan region and these have been discussed in Chapter 6. In the Muslimbagh area, thrust sheets of ophiolites overlie the melanges which in turn have been thrust over the Mesozoic sequence (Figs. 4.11). Immediately below the ophiolites, slivers of garnet-amphibolite schists and chlorite schists occur at many localities (see Chapter 7). The latter may be tectonic slices of older basement rocks (Gansser 1979) or formed due to contact metamorphism (Abbas and Ahmad 1979).

The Zhob Melange sequence is comprised of an upper thrust sheet of ophiolitic melange with blocks of serpentinites, gabbro, basalt, radiolarite and marble. It is underlain by another and older melange which is comprised of a shaly matrix with large blocks of tuffs, pillow lavas, serpentinites, pelagic carbonates, and large olistoliths of various types (Abbas and Ahmad 1979). This melange has been referred to as the "Parh Group" and it is characterised by a sequence of pillow basalt and basaltic hyaloclastites, olistoliths of ophiolites, gabbro, basalt and limestone and greenish, brown to black papery shale intercalated with thin microlitic limestone (Otsuki et al. 1988). The pillow basalt olistoliths are topped by reefal limestone and corals. Triassic fossils have been reported by Otsuki et al. (1988) from their "Parh Group" which, we suspect may have come from clasts in the melange. On the other hand, it is noteworthy that Alleman (1979) has reported Lower Cretaceous to Senonian-Maestrichtian foraminifera from the sedimentary sequence in the Gwal section southeast of Muslimbagh. This sequence is comprised of thin shales, pelagic or microlitic limestone, radiolarite, conglomerate breccias and volcanics (diabase flows and tuffs). Okimura et al. (1988) have referred this sequence as Triassic "Parh Group". The HSC (1960) has mapped this sequence as part of their Parh Group, whereas we believe that all of it is a part of the Zhob Melange sequence which crops out around the ophiolitic masses.

The confusion in identifying the Zhob Melange has been due to the presence of clasts of Permian and Triassic fossiliferous limestone. It misled earlier workers into classifying them as Permo-Triassic sedimentary sequences. In this melange the olistostromes and conglomerates are more conspicuous at some places than others. Furthermore, the melange has been imbricated with the Parh Formation and the thrusts have made the structure more complicated. The Gwal section is comprised of several tectonic slices

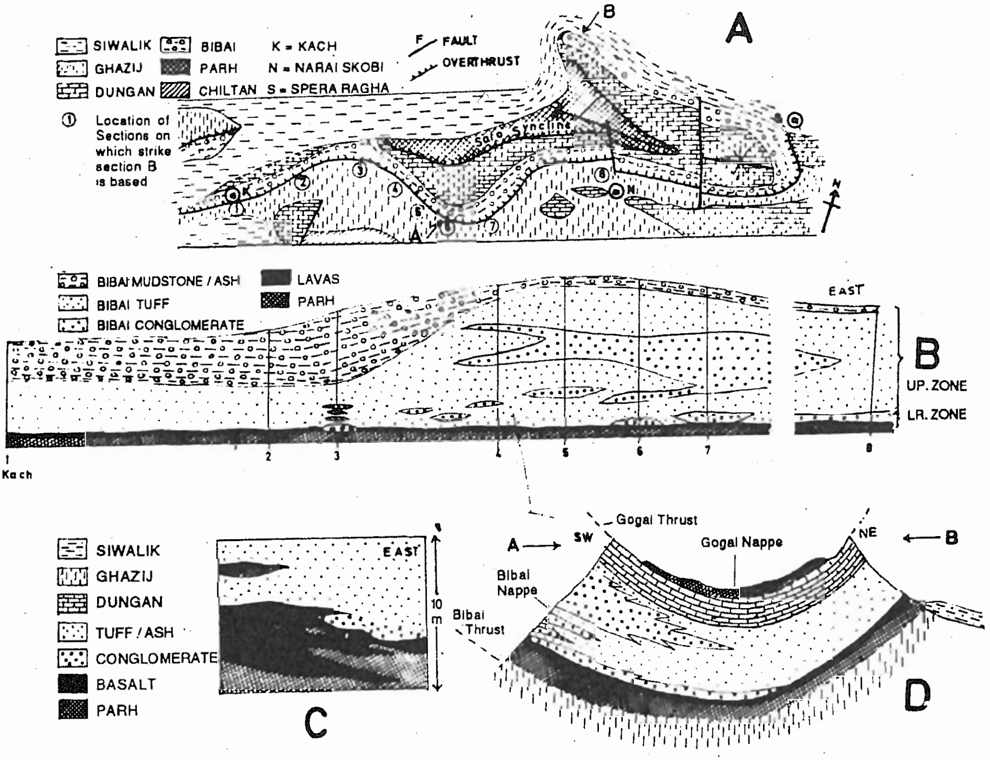


Fig. 4.17. Map and diagrammatic sections showing the structure and stratigraphy of the Bibai Formation, (a) schematic geological map showing the location of the sections on which figure (b) is based; (b) schematic section of the Bibai Formation along the strike of the Bibai nappe; (c) Lower Zone and lower part of the Upper Zone of the Bibai Formation; (d) cross-section of the Bibai and Gogai nappes. (From Kazmi 1979b).

which pinch out or swell within a short distance. The whole stratigraphic section is structurally overturned with the youngest beds (Maestrichtian-Cenomanian) at the topographic base and the oldest (Lower Cretaceous shale and Jurassic limestone) towards the top of the mountain (Alleman 1979).

The southern part of the Zhob thrust belt contains scattered and intermittent outcrops of the Cretaceous Bibai Volcanics (Kazmi 1955, 1979b, Siddiqui et al. 1994). At the base the Bibai lava flows are interlayered with Parh Formation (Fig. 4.17) and the upper part of Bibai is comprised of pyroclastics and volcanic conglomerates (see Chapters 5 and 6). This Formation is similar to the Bad Kachu and Thar Formations in the Bela area. The Triassic to Paleogene sequence of the Zhob belt has been thrust southward over the autochthonous sequence of the Sulaiman fold belt (Fig. 4.12).

In the Waziristan–Kurram region, the Zhob ophiolite-and-thrust belt largely consists of (a) the ultramafic to acidic Waziristan Igneous Complex which, according to Beck et al. (1995), is the remnant of a trans-Himalayan island arc (c.f., Kohistan–Ladakh arc); (b) Kahi Melange, which represents the Neotethyan accretionary prism and trench complex, and comprises intensely deformed glaucophane-bearing tuff which encloses olistoliths of dunite, peridotite, pillow basalt, manganese nodules, Mesozoic belemnite-bearing limestone and kilometre scale blocks of fossiliferous upper Cretaceous limestone (Parh?) and (c) deep to shallow marine Triassic-Cretaceous carbonates, shales and sandstones.

Two distinct episodes of collisional thrusting occurred in this region. Beck et al. (1995) have shown that the first one, associated with the collision of India and the Trans-Himalayan volcanic arc, occurred between 66 and 55.5 m.y., when the Kahi Melange was obducted on the Late Cretaceous to Early Paleocene sequence of the Indian Craton. The unconformity below Lower Eocene marks the time of second major thrusting episode when the Kahi Melange along with Mesozoic to Paleocene Indian slope sediments was thrust farther on to the Indian shelf.

Otsuki et al. (1988) have proposed an interesting tectonic model for the formation of the Zhob ophiolite-and-thrust belt. They envision that with the disintegration of Gondwanaland and northward flight of its several slices, such as Tibet and Karakorams, a mid-oceanic ridge had formed at a short distance from the Indian continental margin during the Triassic. Between this ridge and the continent a relatively shallow and narrow sea existed. The lower part of the Zhob Melange (their "Parh Group") formed due to volcanism on the mid-oceanic ridge. The "Alozai Group" with olistoliths of fusulinid limestone was laid down on the continental side of the shelf. Constrained by the existence of turbidites and olistoliths, they assume that the basin between the oceanic ridge and the continent was about 200 km wide. They attribute scarcity of olistoliths in upper part of the Zhob Melange ("Parh Group" and "Alozai Group") to the widening and shallowing of the basin and deposition of shelf carbonates on continental margin during Lower Triassic to Jurassic, followed by subduction in Cretaceous and eventual obduction of the ophiolites and melanges on the continental margin during Paleocene-Early Eocene.

Sulaiman-Kirthar Fold Belt

This 1,250 km long and 75 to 180 km wide structurally complex zone has a faulted contact with thrust belts to the north and west, whereas south- and eastward its folds gradually lose their amplitude and merge with the foredeep zone (Fig. 4.12). It is comprised of an up to 10 km thick sequence of Jurassic to Recent sedimentary rocks (Table 4.2) which form relatively broad upright or asymmetrical folds. Based on differences in tectonic style and manifestation of surface structures (Photo. 16), the Sulaiman-Kirthar fold belt may be divided into a number of tectonostratigraphic zones (Fig. 4.6) which are described in the following pages.

Sulaiman Arc

This belt is characterised by east-west trending arcuate, convex-to-the south folds (Photo 12). Jurassic to Recent sedimentary rocks are exposed in this belt as shown in Table 4.2. There is a widespread unconformity at the base of the Paleocene Dungan Limestone with thick laterite outcrops between Kach and Loralai. The protolith of this laterite is the basalt of the Bibai Volcanics which suggests that the northwestern part of this belt was covered by Bibai basalts. Ten kilometres southeast of Loralai, near Tor Ghar, an unmetamorphosed alkaline intermediate type intrusion in Cretaceous Sembar Formation has been reported (Jadoon et al. 1991). Bannert et al. (1992) have reported a volcanoe in Paleogene limestone about 75 km southward near Kohlu.

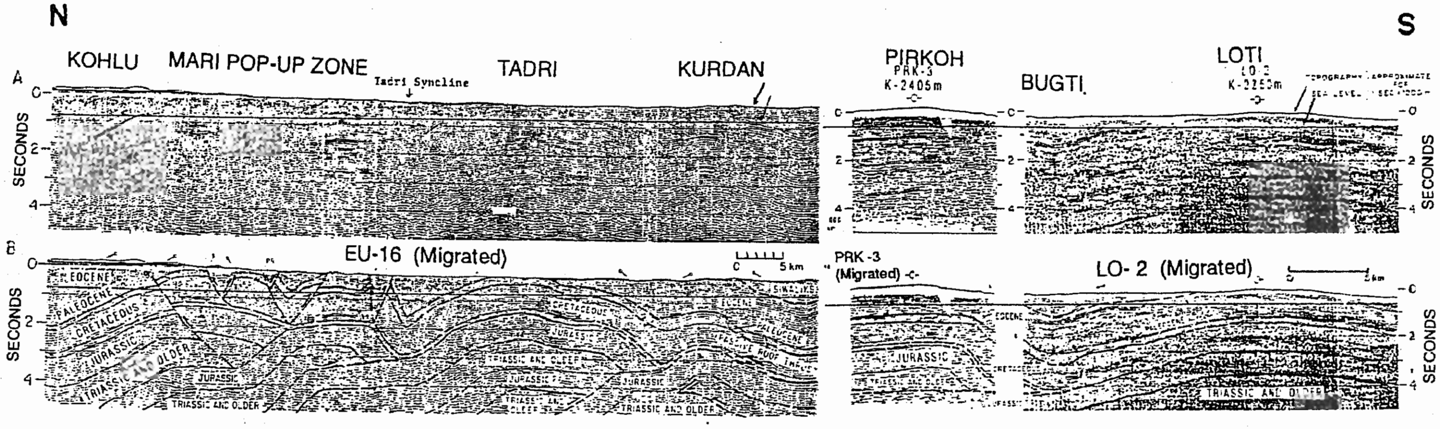


Fig. 4.18. Composite uninterpreted (A) and interpreted (B) seismic line from Sulaiman fold belt showing a passive-roof structure bounded by a roof thrust in Cretaceous shales and a floor thrust in Paleozoic section at the base of the wedge. The simple folds in the passive roof sequence south of Tadri syncline mimic the shapes of duplex horses below them. In contrast, north of Tadri pop-ups are present. These structures are interpreted as concentric, buckle folds formed primarily due to ductility of material at the detachment horizon. (From Jadoon et al. 1993).

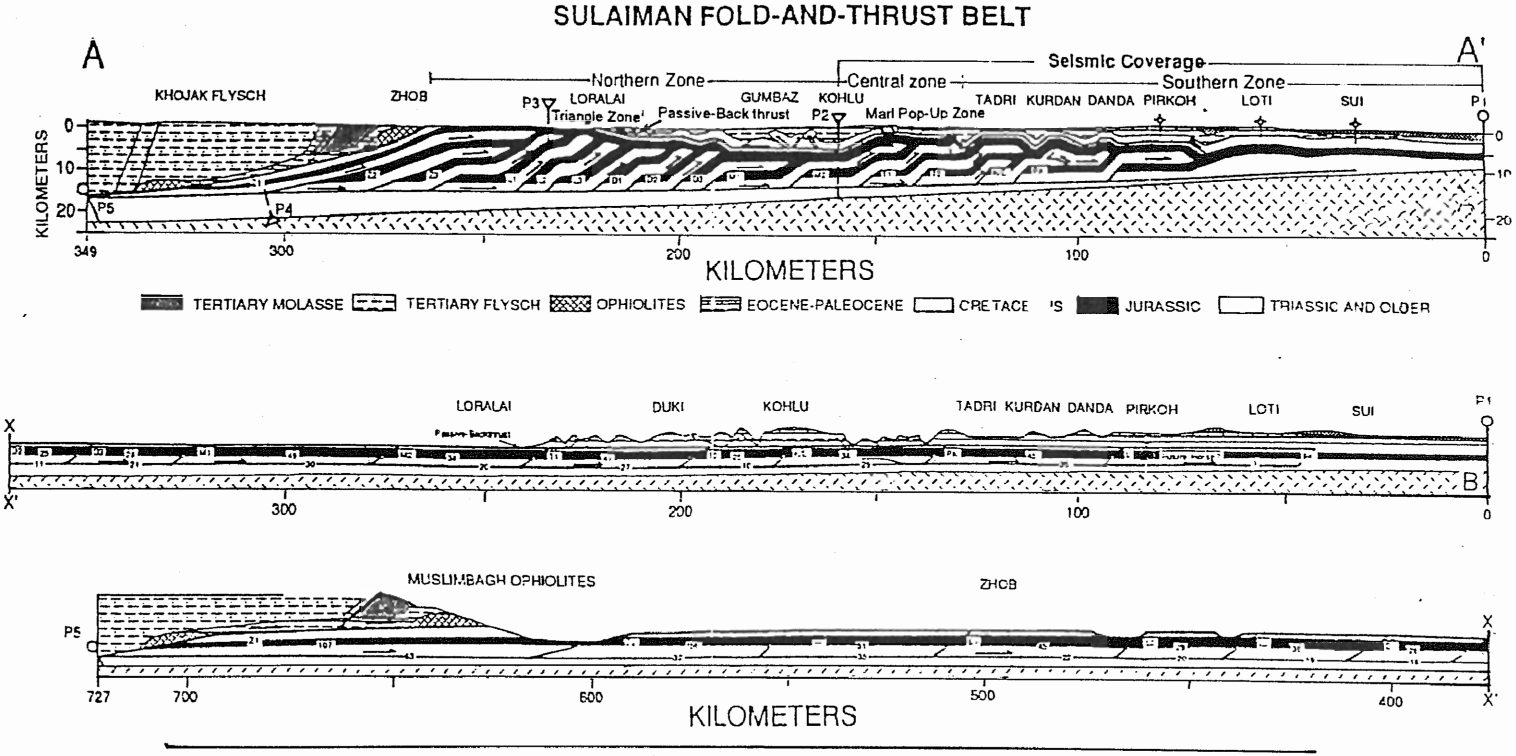


Fig. 4.19. (A) Balanced and (B) retrodeformed structural cross-sections from the Sulaiman lobe in Pakistan. 349 km long deformed section (A) restores to 727 km (B), indicating 378 km of shortening related to Himalayan collision. Horses in the duplex sequence numbered as follows:- Z1-Z3=Zhob Valley, L1-L3=Loralai Valley, D1-D3=Duki Valley, M1-M2=Mari, T1-T2=Tadri, K=Kurdan, D=Danda, P=Pirkoh, L=Loti. (From Jadoon et al. 1993).

In the Sulaiman belt the folds are commonly open, upright, parallel or subparallel, and en echelon. Northward their topographic relief is accentuated, they become tighter, asymmetrical and progressively older strata are exposed in the core of anticlines. The folds are commonly affected by small normal faults and at places anticlines have been cut by tear faults or their flanks have been truncated by reverse faults or high angle thrusts (HSC 1960, Kazmi 1955, 1979a, Jadoon et al. 1992, 1993). In the northern and central part of the fold belt there are extensive high angle thrust faults such as the Ziarat, Khalifat, Mekhtar, Kohlu and Barkan Faults (Kazmi 1979a, Kazmi and Rana 1982, Bannert et al. 1992). The fold belt contains several large NW- and NNW- trending strike slip faults along its eastern and western margins respectively (Fig. 4.10). More significant amongst these are the right-lateral Harnai and Tatra Faults along the western margin and the left lateral Kingri, Manikhawa, Chaudhwan, Domanda, Sulaiman and Moghalkot Faults along the eastern margin (Kazmi 1979a,c, Kazmi and Rana 1982). These strike slip faults are important components of the syntaxis. According to some workers, the lobate geometry of the Sulaiman fold belt has resulted from southward translation of the deformation front along a weak decollement bound by tear faults (Kazmi 1979a, Sarwar et al. 1979, Seeber et al. 1981, Bannert et al. 1992).

Seismic reflection data in eastern and southern part of the Sulaiman fold belt has been recently interpreted to demonstrate the presence of duplex (Fig. 4.18) bounded by a passive-roof thrust in Cretaceous shales and a floor thrust above the crystalline basement (Banks et al. 1986, Humayun et al. 1991, Jadoon 1991, Jadoon et al. 1993). The thickness of Phanerozoic sedimentary wedge in Sulaiman fold belt is more than 10 km. Seismic reflection profiles show that Eocambrian evaporite sequence may not be present in this region. Nevertheless Lillie et al. (1987), Humayun et al. (1991) and Jadoon et al. (1992) believe that there is a basal decollement in pelitic rocks or fine carbonates above the crystalline basement at a depth of more than 14 km. At this depth, with the average geothermal gradient of 25° C/km (Khan and Raza 1986), a temperature of about 350° C may be expected which is sufficient to make the sediments ductile.

A north-south balanced cross-section across the Sulaiman fold belt is shown in Figure 4.19 (Jadoon et al. 1993, 1994). The southern half of this section is well-constrained with seismic coverage but the northern part has been largely extrapolated and filled largely by duplexes of Jurassic and older rocks, analogous to those from the south. This section shows that the forward propagating duplexes of massive Jurassic limestones and older rocks developed in piggy-back fashion. The basement dip is about 2.5° to the north and the sedimentary wedge is exceptionally thick (10 km) at the deformation front. Due to structural duplication the thickness increases northward to about 20 km. Major thrust faults that are responsible for this thickening of the sedimentary wedge do not crop out south of the Loralai Valley (Jadoon et al. 1994a) in the autocthonous zone.

According to Jadoon et al. (1992) the sequential development of structures in this section is as follows:–

i) Growth of broad, concentric fault-tip folds (Sui and Loti anticlines) in the foreland.

ii) Propagation of the basal decollement and uplift of the passive-roof sequence above the duplex structures.

iii) Propagation of the duplexes as critical taper is achieved in the overthickened wedge.

This sequence suggests that folding has preceded thrusting in the frontal fold belt. The roof-sequence is characterised by fault-related folds of variable amplitude, symmetry and extent due to its extended length in the foreland.

On the basis of surface and subsurface data Jadoon et al (1993) identified three structural zones (southern, central and northern) in the Sulaiman fold belt (Fig. 4.19). The southern zone which has been also referred to as Mari-Bugti zone (Kazmi and Rana 1982, Shuaib et al. 1993), extends from Sulaiman foredeep to Tadri and Siah Koh anticlines. It is comprised of Cretaceous to Neogene sedimentary rocks which form east-west oriented, doubly plunging, broad surface folds. Along the foredeep margin there are simple, broad, dome shaped anticlines (Sui and Loti) which are replaced northward by ramp-fault and duplex structures (Pirkoh, Danda, Kurdan, Tadri). Northward the character of folding changes from fault-bend folds (e.g. Pir Koh), to leading edge, ramp overlap anticlines (e.g., Danda anticline), and intraplate folds (e.g. Kurdan anticline) to overlapping ramp anticlines (e.g. Tadri).

The central zone lies between Tadri and Kohlu synclines and is characterised by north and south verging faults which are interpreted as reverse faults and are largely confined to the roof sequence. The roof sequence contains tight anticlines between paired faults which are interpreted as pop ups (Fig. 4. 19). These faults are secondary structures (out-of-sequence thrust) of considerable lateral extent. Some of these faults are likely to be active (Kazmi 1979a). The region is characterised by high but shallow (about 5 km) seismicity (Quittmeyer et al. 1979, 1984), possibly related to the out-of-sequence deformation.

The northern zone lies between Kohlu and Muslimbagh and includes the Zhob ophiolite-and-thrust belt which has been described earlier. The tight folds in this region may be detachment folds, above an upper decollement in thick Cretaceous shale that crops out extensively in the Loralai Valley and has been interpreted as emergent passive-roof sequence along the passive Loralai back-thrust. This Cretaceous sequence forms a conspicuous and enigmatic zone in Loralai area and has been referred to as the Loralai triangle zone*. This zone broadly overlaps the Sanjawi arc (HSC 1960). It may have originated in the hinterland due to structural uplift but subsequently migrated towards the foreland with increasing shortening (Jadoon et al. 1993). South of the Loralai triangle zone lies the Gumbaz structural depression which is largely composed of Paleogene rocks.

Sibi Trough

The Sibi trough (Sarwar et al. 1979, Kazmi and Rana 1982, Banks et al. 1986, Shuaib et al. 1993) has been variously referred to as Urak trough (HSC 1960), Sibi re-entrant (Movshovitch et al. 1965) and Sibi depression (Raza et al. 1989a,b). It is bounded by the Sulaiman arc and Kirthar fold belt on the east and west, and the Zhob thrust-belt and Kirthar foredeep to the north and south respectively (Fig. 4.6 and Photo. 16). In this trough a 15 km thick sequence of Triassic to Recent sedimentary rocks lies above the basement (Fig. 4.20), the upper 7,000 m of which is comprised of Siwalik molasse. The Paleogene sequence is exposed along its eastern, northern and western margins only. In the northern part of the trough, the Siwalik molasse is topped by about 1,600 m thick Urak Boulder Conglomerate (Soan Formation). This Plio-Pleistocene conglomerate caps

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*According to Gordy et al. (1977) a triangle zone is a region between extensive faults of opposing vergence.

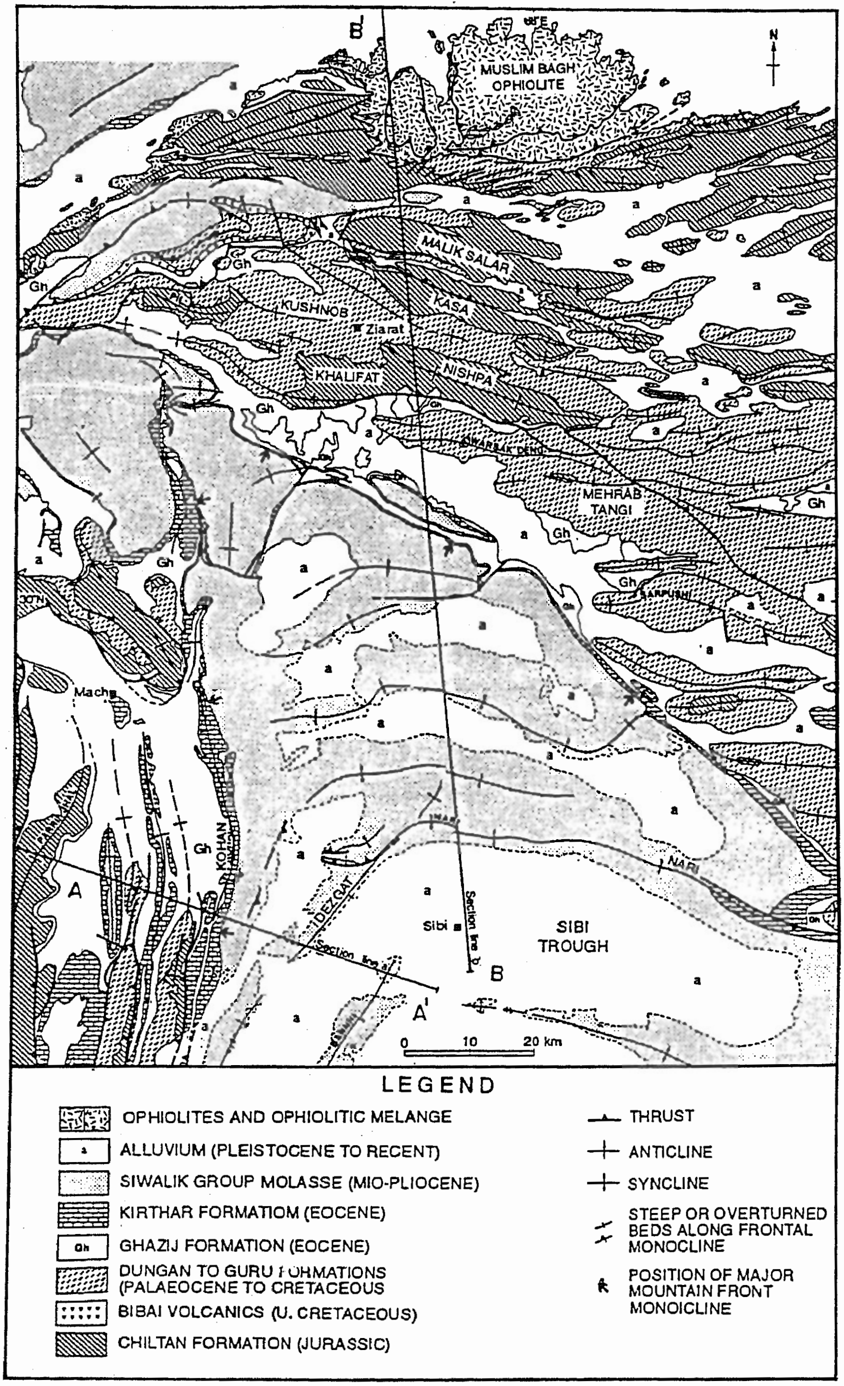
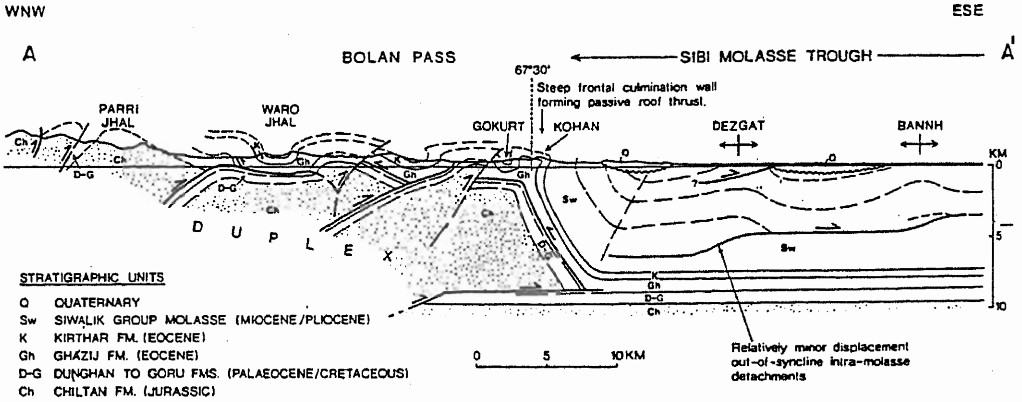
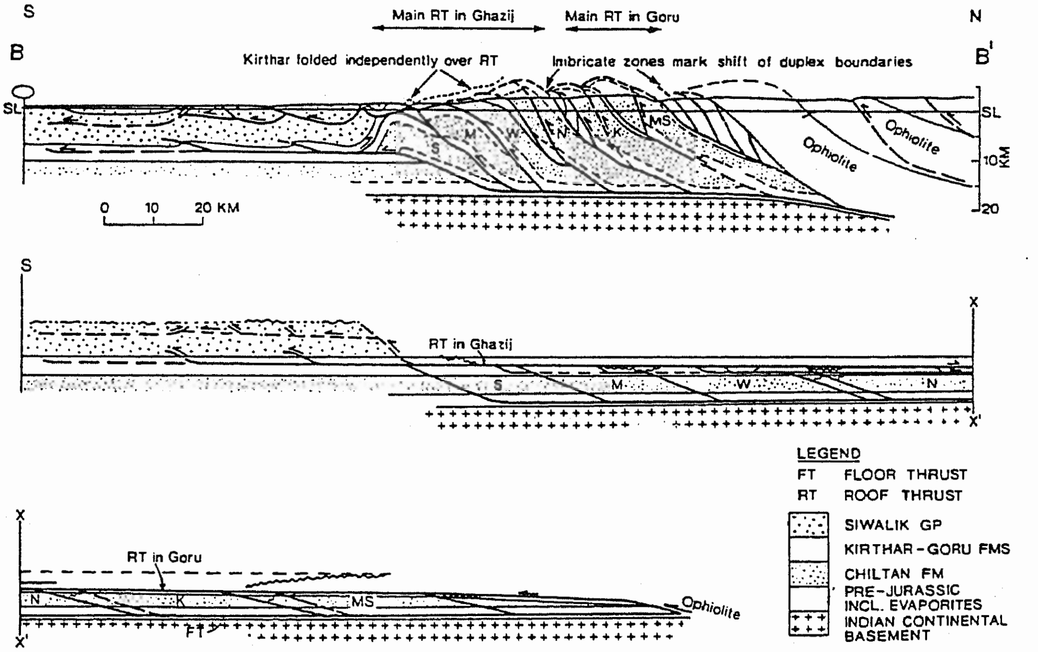


Fig. 4.20. Geological map and sections of a part of the Sulaiman-Kirthar Fold Belt. (Modified from Banks et al. 1986).



Geological cross-section across the Kirthar fold-and-thrust belt (section line AA').



Actual and restored cross-section across the Sulaiman fold-and-thrust belt (section line BB').

the Zarghun Range, the loftiest mountain (3,562 m) in Balochistan, higher than the surrounding proximal ranges comprised of Jurassic limestones. The basal part of the Urak Conglomerate is largely comprised of Eocene limestone boulders, whereas in its upper part clasts of Jurassic limestone dominate. This feature reflects the uplift and the unroofing and denudation history of the region (Kazmi 1955, 1961).

Fanfold-type anticlinal flexures traverse the Sibi trough in a NE to EW direction. Some of the folds are sliced along their crests by extensive reverse faults. These structures have divided the trough into smaller synclinal zones. From north to south, the structure gradually changes, particularly in the orientation of fold axes which have a northerly trend in the north, but southward it changes to east-west (Photo. 12). Folds vary greatly in size and at some places their axes are sinuous. The smaller ones are commonly normal to one another, particularly in the north where they are largely disharmonic. According to Kazmi (1955, 1961), the Sibi trough lies over a decollement in the thick incompetent Ghazij shale. More recently Banks et al. (1986) have also endorsed this view with interpretation of a roof thrust in the Ghazij shale (Fig. 4.20). They have highlighted the presence of a broad zone of steep foreland dip and occasional overturning of molasse sediments of the Sibi trough at the mountain front. They interpret this steep zone as a major frontal culmination wall in the roof sequence of a duplex, developed in thick limestone units beneath the Ghazij shales.

It appears that the Sibi trough is wedged in between duplex sequences of the Sulaiman arc and Kirthar fold belt (Fig. 4.20). This view is, to some extent, constrained by limited seismic data. Bannert et al. (1992) consider the north lying Zarghun basin to constitute a passive roof duplex between the Sulaiman and Kirthar belts. According to them the Sulaiman and Kirthar belts are large fault blocks bound by reactivated basement faults.

Kirthar Fold Belt

This fold belt forms a 50 to 70 km wide and about 380 km long, north-south trending folded zone between Quetta and Karachi. It is bounded by the Bela–Zhob ophiolite-and-thrust belt to the north and west, and by the Sibi trough, Kirthar foredeep and Indus platform to the east (Fig. 4.6). Stratigraphic sequence in this belt consists of Jurassic to Recent sedimentary rocks (Table 4.2). Based on differences in tectonic style and variations in stratigraphy, this belt may be divided into a number of smaller structural units (Fig. 4.6 and Photo. 13) which are briefly described below.

Kalat anticlinorium: It forms the northern-most part of this belt and is largely comprised of thick to massive Jurassic limestones. It consists of 20 to 50 km long NNE-trending, parallel or en echelon, doubly plunging anticlinal hill ranges, which include some of the tallest peaks (above 3,000 m) in Balochistan. The anticlines are broad, with relatively steep limbs and flatter crests. Several small normal and reverse faults cut these folds. The anticlinal hills are separated by equally wide synclinal valleys covered by alluvium and underlain by Cretaceous and Paleogene shales and limestone. Westward the Sariab–Saran Tangi Thrust Fault (Kazmi and Rana 1982) separates this zone from the Zhob thrust belt and eastward it is faulted against the Zarghun basin of the Sibi trough and the fold belt of Nagau Range.

Kalat Plateau: This plateau lies to the south of Kalat anticlinorium and has a similar thrust contact with Zhob thrust belt and the folded zone of the Nagau Range to the west and east respectively. It is apparently a subsided block or depression filled with Eocene Kirthar Limestone (Spintangi Limestone of HSC 1960) which forms a gently undulating synclinorium with small, gentle NE-trending folds. In rare instances the folds have steep or vertical dips. Reverse and tear faults are common. Along the northern and eastern margins, Ghazij shale and Marap Conglomerate underlie the Kirthar Limestone (HSC 1960). Westward this limestone is in direct fault contact with the Triassic to Cretaceous rocks. Eastward it overlaps Lower Eocene Ghazij and Paleocene Dungan formations. Northward the anticlinal hill ranges are composed of massive Jurassic Chiltan Limestone and they abruptly plunge southward beneath the Paleogene sequence of Kalat Plateau. There is no accord between the structures north and south of Kalat Plateau and we believe that the latter's Paleogene sequence sits on a decollement based in the thick plastic Ghazij shale. Southward the left lateral Anjira–Gazan Fault (Sarwar et al. 1979) apparently separates mildly deformed Paleogene Kalat Plateau sequence from the intensely deformed Mesozoic sequence of the Khuzdar knot.

Khuzdar knot: It is an enigmatic and irregularly shaped, structural feature, about 50 km wide and 70 km long. It is bound by faults to the west, north and east, though its southern boundary is uncertain. It is an intensely deformed zone. However, the intensity of deformation decreases southeastward. Thick to massive Jurassic limestones form scattered, irregularly oriented, tightly folded anticlinal hills separated by narrow, irregularly shaped small valleys and basins. The latter are filled with tightly folded Cretaceous shale and thin limestone, unconformably overlain by less deformed Eocene Kirthar Limestone. The anticlinal axes trend in almost every direction from NE, NS to NW. They are invariably curved, wiggly or even form "U" shaped bends (Photo. 13).

The synclines are narrow, branching structures, squeezed into Y-shaped or triangular valleys. Eastward the folds are larger, broader, and at places dome-shaped with a northwest to northeast strike. In the west the fold axes, despite their wiggly nature, have a general broad northeasterly curvature (Photo. 13). This feature is attributed to the left lateral strike slip movement along the Ornach-Nal and Anjira–Gazan Faults.

It is noteworthy that the general structural trend surrounding the Khuzdar knot is N-S, NNE or NNW though within the knot it is chaotic and the folding is disharmonic. We, therefore, suspect that the thick, highly disturbed, and distorted Mesozoic and Cenozoic sequence of the Khuzdar knot overlies a decollement possibly in an evaporite sequence above the basement. This feature, coupled with transpressional and compressional tectonics, is probably responsible for the chaotic structure of the Khuzdar knot.

Khude Range fold belt: This belt is about 10 to 30 km wide, 200 km long, and is situated south of the Khuzdar knot. Along its western and eastern margins, the Pab and Kirthar Thrust Faults, separate it from the Bela ophiolite-and-thrust belt and Nagau–Kirthar fold belt respectively. The thrust faults dip away from the Khude belt. Its western part is covered by Paleogene sedimentary rocks which fill a narrow elongated synclinal trough with a few anticlinal kinks (Fig. 4.12). The eastern part comprises a number of parallel or en echelon anticlinal flexures with Jurassic to Cretaceous rocks exposed in the core. In the Khude belt the folds are simple, linear features with a NNW trend. Southward near Karachi the folds swing to the southwest.

Nagau–Kirthar fold belt: It forms the inner and eastern part of the Kirthar fold belt and covers the Nagau and Kirthar Ranges. This belt is 10–30 km wide and consists of shallow marine to deltaic and estuarine Paleogene sequence which forms the high hill-ranges in western part of the belt, and a thick conformable sequence of Neogene molasse that forms relatively low hills on the eastern side. In this belt the outer folds are tight, mostly parallel to subparallel, en echelon, doubly plunging with a northerly trend, though several small folds have an oblique northeasterly direction. Reverse and tear faults are common. The western margin of this fold belt has been truncated by reverse or thrust faults, namely the Nagau, Lehdo, Zardak, Kirthar and Hub Faults (Kazmi and Rana 1982). Its eastern margin is comprised of several small pop-up structures and it is characterised by steeply dipping anticlinal folds of Eocene Kirthar Limestone. This limestone forms the steep frontal culmination wall of the passive roof thrust of a duplex (Fig. 4.20).

According to Ahmed et al. (1991b) the northern part of this belt is comprised of a wrench-driven thrust and fold belt; the central part is characterised by a left lateral convergent regime followed by foreland dipping Kirthar Thrust, and the southern part exhibits a left lateral divergent trend.

Karachi embayment zone: Earlier referred to as the Lyari embayment zone by Kazmi and Rana (1982) and as Karachi synclinorium by HSC (1960), this zone is the southward continuation of the Kirthar fold belt. It is largely covered by Miocene to Recent sediments which form well defined broad SSW trending folds.

Sanbakh–Lakhra uplift zone: Previously referred to as Hyderabad anticlinorium or Hyderabad arch (HSC 1960), it lies to the east of Karachi embayment zone and forms the southeastern part of the Kirthar fold belt. It is covered by Paleogene and Neogene sedimentary rocks with many unconformities. This sequence forms broad doubly plunging, low dipping, north-south trending folds, intersected by numerous normal and reverse faults. It is characterised by slightly higher regional gravity (Fig. 4.7). The unusually frequent stratigraphic unconformities suggest that it is a structural high. It is, nevertheless, a zone with a 6–8 km thick pile of sediments in which the Mesozoic alone is 3 km thick (HSC 1960, Raza et al. 1990a).

Kakar Khorasan flysch basin: This basin lies to the north of Zhob ophiolite-and-thrust belt. Mentioned as the North Zhob District by HSC (1960), it was included in their Arenaceous Zone. Ahmed (1991) has used the name "Pishin Basin", Raza et al. (1989a) call it "Pishin Rear Depression", whereas Stöcklin (1977), and Treloar et al (1993) consider it as a part of the larger Katawaz Basin which extends southward from near Kabul to Zhob and links up with Makran flysch basin. We have therefore discussed it with the Makran accretionary zone.

NORTHWEST HIMALAYAN FOLD–AND–THRUST BELT

The Northwest Himalayan fold-and-thrust belt occupies a 250 km wide and about 560 km long, irregularly shaped mountainous region stretching from the Afghan border near Parachinar, up to the Kashmir Basin (Photo. 11). The Hazara–Kashmir and Nanga Parbat Syntaxes form its eastern margin. It covers all the terrain between the Main Mantle Thrust (MMT) in the north and the Salt Range Thrust in the south. This region is comprised of the mountain ranges of Nanga Parbat, Hazara, southern Kohistan, Swat, Margalla, Kalachitta, Kohat, Sufaid Koh, Salt Range and its western extension (Figs. 2.2 and 4.6.).

A major thrust fault, the Panjal–Khairabad Fault divides the NW Himalayan sequence into a deformed southern zone, often referred to as the external or foreland zone, and a deformed and metamorphosed northern zone, also known as the hinterland zone (DiPietro et al. 1996, Pivnik et al. 1996). The foreland zone is comprised of the Hazara–Kashmir Syntaxis, Salt Range and Kohat–Potwar fold belt and the Kurram–Cherat–Margalla thrust belt, whereas the hinterland zone comprises the Himalayan crystalline nappe-and-thrust belt (Fig. 4.6).

The Himalayan crystalline belt is comprised of the exposed northern margin of Indo-Pakistan crustal plate and it is characterised by intensely deformed, tightly folded and imbricated Precambrian to Early Mesozoic metamorphic and igneous rocks. Southward these crystalline rocks have been thrust over the Kurram–Cherat–Margalla thrust belt which is comprised of several thrust sheets of largely unmetamorphosed Precambrian to Paleogene rocks. Farther to the south, the latter sequence has been emplaced over the Kohat–Potwar fold belt which terminates southward in the Salt Range and Trans-Indus Ranges. Precambrian to Neogene rocks are exposed in these ranges, though larger part of the fold belt is covered with Neogene sediments forming large broad folds. The Kohat–Potwar region is characterised by Paleogene and Precambrian evaporites which effectively decouple the sediments from the basement and control the tectonics of this belt through formation of decollement zones and thrusts having up to 100 km southward translation (Lilie et al. 1987).

The eastern margin of the NW Himalayan fold-and-thrust belt in Pakistan comprises the north-south trending Hazara–Kashmir Syntaxis and farther to the northeast, the Nanga Parbat–Haramosh Massif. These syntaxes separate the above fold belt from the Central Himalayan fold-and-thrust belt of India. They apparently comprise a major tectonic divide, inasmuch as the structures on their either side have been abruptly truncated.

The four-fold structural division of the Himalayas in India (see Chapter 3) is not so clearly defined across Hazara–Kashmir Syntaxis. However there are similarities between some of the structural zones of the Himalayan fold-and thrust belt in Pakistan and those of the Indian Himalayas. According to Yeats et al. (1984) there is a broad equivalence between Sub-Himalayas and Kohat–Potwar fold belt; the Lesser Himalayas, and Kurram–Cherat–Margalla fold-and-thrust belt, and between the High Himalayas and Nanga Parbat–Haramosh Massif. The southern part of Swat–Hazara belt with its well developed Paleozoic sequence may be correlated with the Tethyan Himalayas, though the structural sequence is not quite the same. This comparison becomes evident from the following account of the geology and tectonics of NW Himalayan fold-and-thrust belt in Pakistan.

Hazara–Kashmir Syntaxis

In the northern corner of Pakistan, between Mirpur and Muzaffarabad and farther to the north and northeast, the geological formations and broader geological structures of the Himalayas make an abrupt hairpin bend as if "they were bent round a pivotal point obstructing them" (Wadia 1931). This spectacular structural feature of the Himalayas (Fig. 4.6) was first discussed in detail by Wadia (1931) who referred to it as the 'North-west Himalayan Syntaxis'. Since then several authors have proposed different names e.g., Punjab Orocline" (Carey 1958), "Abbottabad Syntaxis" (HSC 1960), "Western Himalayan Syntaxis" (Gansser 1964), "Punjab Re-entrant" (Johnson et al. 1976), "Kashmir Syntaxis" (Crawford 1974b), "Hazara–Kashmir Syntaxis" (Calkins et al. 1975, Desio

1976) and "Jhelum Re-entrant" (Visser et al. 1978). In recent years most workers have referred to it as the Hazara–Kashmir Syntaxis (Sarwar et al. 1979, Bossart et al. 1984, 1989, Ghazanfar et al. 1986, Ottiger 1986, Wells et al. 1987, Greco et al. 1993) and we propose that this name should be formally adopted for this syntaxis.

Tectonostratigraphic Setting

The Hazara–Kashmir Syntaxis is a complex tectonic zone (Photo. 6). It is difficult to define an outer limit for it, though its axial zone is well-defined by a stack of thrust faults which form a loop around its axis. Precambrian to Neogene sedimentary, volcanic and metamorphic rocks and Cambrian or earlier granitic rocks are exposed in the syntaxial zone and its vicinity. The general geology and tectonostratigraphic sequence of the region has been shown in Fig. 4.21. Various lithostratigraphic units exposed in the syntaxis have been described in Chapter 5.

The axial zone of the syntaxis has a NNW orientation and is largely covered by Murree Formation (Oligocene ? to Miocene). This Formation comprises reddish siltstone and shale and is at least 1,700 m thick. Near Muzaffarabad, Precambrian to Cambrian and Paleocene sedimentary rocks are exposed in an anticline (Fig. 4.21) which is crossfolded, overturned and thrust southwestward along the Muzaffarabad Fault. This fault dips 20°–25° E and has moved the older rocks westward over Paleocene limestone and Murree rocks (Calkins et al. 1975). Bossart et al (1988) suggest that the core of the syntaxis is comprised of a deformed dome which developed by 'layer compression' sub-perpendicular to a southwestward over-thrust direction (Fig. 7.14). Near Muzaffarabad the sediments have been metamorphosed to prehnite-pumpellyite grade (Greco 1991).

The axial zone of the syntaxis, with its Miocene molasse cover (Murree Formation), is in continuation of the Sub-Himalayan zone to the southeast. The Himalayan Frontal Thrust (see Chapter 3) continues from the Sub-Himalayan zone northwestward into the core of the syntaxis and is terminated against the MBT (Kazmi and Rana 1982, Baig and Lawrence 1987).

North of Balakot, the axis of Hazara–Kashmir Syntaxis bends northeastward and continues into Kaghan and beyond, into the Nanga Parbat–Haramosh region, where some workers call it "the Nanga Parbat Syntaxis" (Coward 1985). This eccentric position of the Nanga Parbat Massif is attributed by Desio (1979) to a possible clockwise rotation. On western side of Hazara–Kashmir Syntaxis there are several arcuate, south verging thrust faults which terminate in the Jhelum Fault at an acute angle between Balakot and Kohala (Fig. 4.21).

MBT (Murree Thrust)

A hairpin-shaped system of faults truncates the Murree Formation on the east, north and west. It abuts the Mesozoic and earlier rocks against the Murree Formation (Fig. 4.21). West and north of this fault zone, within the short distance of 1 to 5 km, there is a parallel thrust fault along which Precambrian sequence has been pushed over the Paleozoic and Mesozoic rocks. These two faults were named as Murree and Panjal Thrusts respectively by Wadia (1931). Some recent workers call the Murree Thrust on both limbs of the syntaxis as the Main Boundary Thrust (Treloar et al. 1989a, 1990, Greco 1991). This fault continues northwestward, turns westward near the apex of the syntaxis and then bends southward towards Balakot. According to Calkins et al. (1975),

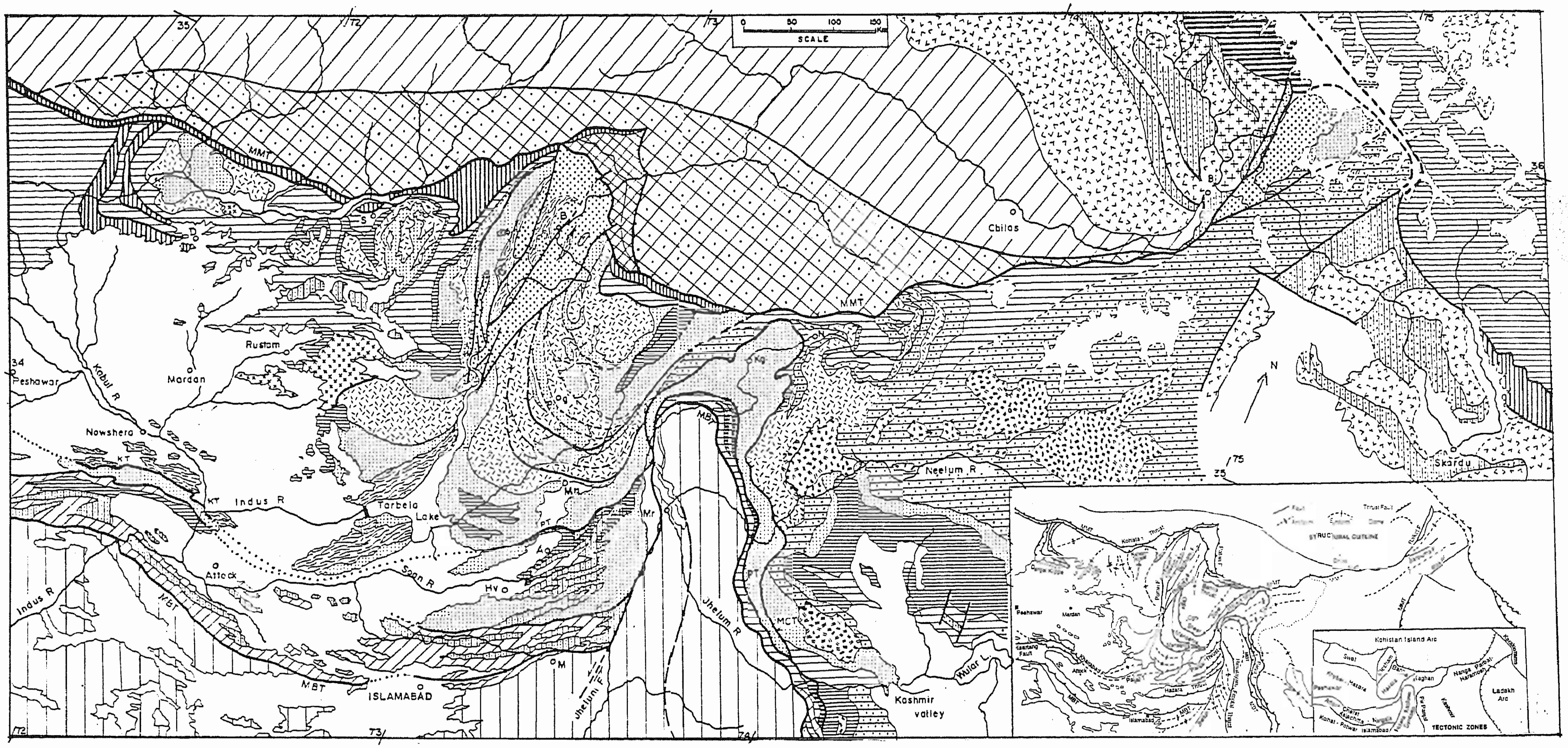


Fig. 4.21. Geological map of the Swat-Hazara crystalline nappes Kalachita-Margalla fold-and-thrust belt & adjacent regions A-Abbotabad B-Besham, Bj_Bunji, D-Dargai, Ka-Kagahan, M-Muree, Mn-Mansehra, Mr-Muzaffarabad, N-Naran, Og-Oghi, S-Swat, Kt-Karirabad Thrust, MBT_Main boundar Thrust, MCT-Main Central Thrust, MMT-Main Mantle Thrust. (based on data from Qureshi et al. 1993, Di Pietro et al. 1993, 1995, Greco et al. 1993, Pogue et al. 1992, Treloar 1989, 1991, Searle 1991, Madin et al. 1989, Lawrence et al. 1989, Yeats et al. 1987, Chaudhry et al. 1986, Kazmi et al. 1982, Calkins et al. 1975, Lateef 1970, Gansser 1964).

	KOHAT-POTWAR	HAZARA-KASHMIR SYNTAXIS	ATTOCK-CHERAT KALACHITTA-MARGALLA	KHYBER-HAZARA METASEDIMENTARY BELT, PESHAWAR BASIN	MALAKAND-SWAT	BESHAM	HAZARA	KAGHAN	NANGA PARBAT HARAMOSH	KASHMIR-PIR PANJAL	KOHISTAN LADAKH	KARAKORAM
Cenozoic ▥	Siwalik Gr. Rawalpindi Gr.	Murree Fm.	Murree Fm.						Leucogranites, foliated granites, pegmatites			-
Paleocene-Eocene ▨		Patala Fm. Lockhart Lst.	Kuldana, Chor Gali, Margalla Hill, Patala, Lockhart & Hangu Fms.		Indus suture melange & ophiolites Kot Prangha melange & ophiolites		Indus suture melange			Subathu Fm.	Granite, Diorite, Gabbro, Shyok melange Kohistan Batholith, Chalt Volcs. Metasediments, Chilas Complex, Kamila Amphibolite, Jijal Complex	Karakoram Batholith Complex
Mesozoic ▦		Kingriali Fm.	Kawagarh, Lumshiwal, Chichali, Samana Suk, Shinwari, Datta, Kingriali, Tredian & Mianwali Fms.	Nikanai Ghar Fm.		Sodic Granites				Megalodon Lst.		
Late Paleozoic to E. Mesozoic ☰		Panjal Fm. Panjal Trap		Kashala Fm. Karapa Green Schist	Alpurai Gr.		Banna Gr.	Burawai Fm.		Panjal Fm. Panjal Trap		Karakoram Metamorphic Complex
Paleozoic (Undifferentiated) ☰			Inzari Lst. Hisartang Fm. Darwaza Fm.	Jafar Kandao, Panjpir, Nowshera, Misri Banda, Ambar Formations, Ambela Granite	Morghazar Fm., Malakand Granite Jobra Fm., Swat Granite Gneiss	Sodic Granites		Naran Fm.		Dogra Slate, gneiss & migmatites		
Cambrian ▨		Abbottabad Gr.	Abbottabad Gr.				Abbottabad Gr., Mansehra Granite	Sharda Granite Gneiss.	Shingus Gneiss			
Late Proterozoic ░		Hazara Fm. Tanawal Fm. Kaghan Fm.	Dakhner Fm., Hazara Fm.	Shekhai, Utch Khattak, Shahkot, & Tanawal Fms	Manglaur Fm.	Tanawal Fm.	Tanawal Fm.	Kaghan Gr.	Haramosh Schist	Salkhala Fm.		
Mid to Late Proterozoic ▒				Manki Fm.		Karora Gr.						
Early Proterozoic ░						Besham Gr.			Iskere Gneiss			

IGNEOUS ROCKS
- ＋ Tertiary granites & Gabbros
- ▨ Cretaceous to Paleogene granites
- ✕ Late Paleozoic to Mesozoic granites
- ▨ Cambrian granites
- ▨ Granites (Undifferentiated)
- ▥ Ophiolites & melanges
- ▨ Gabbronorites (Chilas Complex)
- ▩ Amphibolites (Kamila)
- ▩ Ultramafics (Jijal, Pattan)
- ☰ Volcanics (Panjal)

it dips 50° to 70° E northwest of Muzaffarabad. Southward it steepens to 70° E and locally it is vertical. Vertical stratigraphic displacement is about 3,300 m (Calkins et al. 1975) and strike-slip movement is not evident.

The Murree Fault runs in an E-W direction south of the Margalla Hills. Westward, apparently it links up with the Parachinar Fault. Most of the present workers now refer to the Murree–Parachinar Faults as the MBT.

Jhelum Fault

Kazmi (1977b) pointed out that the fault along the western margin of the axial zone of the syntaxis was a left-lateral strike-slip fault and he named it the Jhelum Fault. Baig and Lawrence (1987) described the Jhelum Fault as a left-lateral strike-slip fault and reported that along this fault Murree, Abbottabad and Hazara Formations are highly deformed between Balakot and Muzaffarabad. Blocks of Panjal Volcanics and Triassic limestone have been dragged several kilometres southward, the rocks are brittlely deformed and a left-lateral offset of about 31 km is indicated on the western limb of the syntaxis. The Jhelum Fault apparently dislocates the MBT and terminates the eastward continuation of some of the structures of NW Himalayan fold-and-thrust belt which shows that it is the youngest major tectonic feature in the syntaxial zone. A number of east-west trending faults join the Jhelum Fault at an acute angle pointed northward, indicating a relative left-lateral strike slip movement.

Panjal Thrust (Khairabad Fault)

The Panjal Thrust runs parallel to MBT on the eastern limb of the syntaxis. The two faults curve around the apex of the syntaxis then bend southward (Fig. 4.21). According to Wadia (1931), Calkins et al. (1975), Bossart et al. (1984) and Greco (1991), the two faults join about 5 km north of Balakot. However, according to Baig and Lawrence (1987) a separate left-lateral strike-slip fault truncates the Panjal Thrust and MBT north of Balakot. The Panjal Thrust probably separates from MBT about 6 km south of Balakot and continues beneath Kunhar Valley alluvium up to Garhi Habibullah. It then swings southwestward and is clearly exposed. From the point of bifurcation, the MBT takes a southeastward course up to Muzaffarabad and then southward to Kohala (Fig. 4.21).

Greco (1991) has used the name 'Mansehra Thrust' for Panjal Thrust on western side of the syntaxis. A change in name would be very appropriate because this thrust has a different tectonic and stratigraphic setting from the Panjal Thrust on the eastern limb of the syntaxis. There the Precambrian Salkhala Formation overlies a thick sequence of Permian Panjal Volcanics along the Panjal Thrust, whereas on western side of the syntaxis Precambrian Tanawal Formation has been thrust over Precambrian to Cambrian Abbottabad Group and Jurassic to Cretaceous rocks. On the eastern side the MBT runs parallel to Panjal Thrust but on the western limb MBT follows an oblique course and southward the gap between the two widens considerably. In fact on the western limb of the syntaxis the N-S trending fault may not be the MBT, but a sinistral strike slip fault as pointed out by Kazmi and Rana (1982) and Baig and Lawrence (1987). The name 'Mansehra Thrust' has been previously used for another thrust fault near Mansehra, north of the Panjal Thrust (Baig et al. 1987, Treloar et al. 1989a, 1990). To avoid this confusion in names, Baig et al (1989) have referred to the Panjal Thrust in Hazara as "Abbottabad Thrust". Westward, the Panjal Thrust apparently links up with the Khairabad Thrust (Yeats and Hussain 1989). Some recent authors refer to it as Khairabad–Panjal or as Khairabad Fault (Searle and Khan 1996, Di Pietro et al. 1996).

Main Central Thrust (MCT)

The Main Central Thrust (MCT) is an intracontinental thrust that separates the Higher and Lesser Himalayas in India and Nepal. It was first described by Heim and Gansser (1939) as a tectonic contact between the Himalayan autochthonous sedimentary sequence and the overlying crystalline complex. It has been traced from Nepal up to southern Kashmir (Gansser 1964). Its westward extension across the Hazara–Kashmir Syntaxis into the Pakistan part of the Himalayas, has been in doubt largely because the region northwest of the last confirmed location of MCT (i.e., the eastern limb of the Hazara–Kashmir Syntaxis) has been only scantily mapped. Some of the earlier workers (Gansser 1964, Yeats and Lawrence 1984) extended MCT into the Panjal Thrust. Since then several suggestions have been made for the location of the MCT in Pakistan (Fig. 4.22).

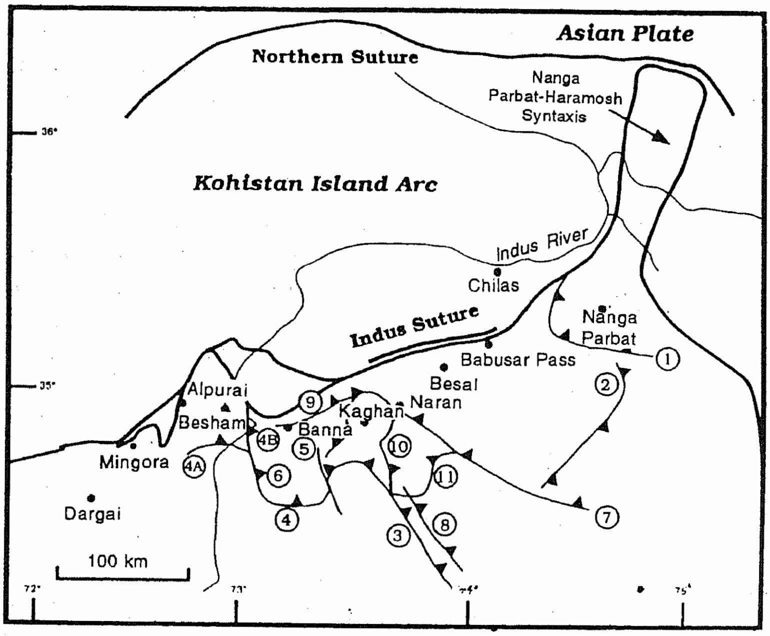
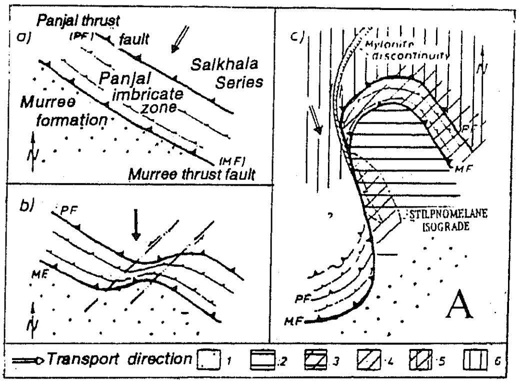


Fig. 4.22. Sketch map of NW Himalaya showing suggested locations of the Main Central Thrust (MCT) or thrusts considered as analogues of MCT. 1. MCT-Coward et al. (986) and Madin et al. (1989). 2. Shontargali Thrust-Tahirkheli (1987, 1992). 3 Panjal Thrust-Gansser (1964), Lawrence et al. (1988). 4. Mansehra and Panjal Thrust (coeval to MCT), Coward et al. (1988). 5. Balakot Shear Zone-Bossart et al. (1988). 6. Oghi Thrust-Treloar et al. (1989), Greco et al. (1989). 7. Luat Fault-Chaudhry and Ghazanfar (1986). 8. MCT-Greco (1989). 9. MCT-Chaudhry and Ghazanfar (1992). 10. MCT-Papritz (1989), Rey (1989), Greco et al. (1989). 11. MCT-Fontan and Schouppe (1994).

Coward et al. (1986) and Madin et al. (1989) consider the Raikot Fault at the western edge of the Nanga Parbat Syntaxis as a terminal tear fault of the MCT, though they have not mapped its connection with the known MCT in Kashmir. Chaudhry et al. (1986) have shown that in Kaghan the granites and gneisses of the Sharda Group have been thrust southwards over the metasedimentary sequence of the Kaghan Group. According to them the Luat–Batal Fault that separates these two groups is the MCT. There is a sudden jump in metamorphic grade across this fault. Chaudhry and Ghazanfar (1992) extend this fault westward up to the MMT. It reappears east of the Indus and continues across the southern part of the Besham block to Malakand. Coward et al. (1988b) suggested that the Mansehra and Panjal Thrusts were the equivalent of the MCT as they separated the internal and external units of the Indian Plate.

Greco (1991) has shown that the MCT does not link up with the Panjal Thrust. It runs parallel to this thrust in a NW direction. According to him the MCT joins a prominent granite-derived mylonite zone, the Balakot Shear Zone, east of Mahandri. Bossart et al. (1984) first observed this zone near Balakot. It is a 1-5 km wide zone of ductile deformation comprising of mylonites with alternations of dark mica-rich layers, and mylonites with feldspar clasts. South of Balakot it is covered by alluvium. Greco et al. (1989) mention that the Balakot Shear Zone links up with the Oghi Shear Zone in Hazara. North of Balakot, in Kaghan this mylonite zone has been mapped and it forms a long loop around the apex of the syntaxis up to Naran, then curves southeastward and links up with MCT near Mahandri. It abuts the low metamorphic grade Precambrian Salkhala Formation against amphibolite-facies Paleozoic (?) carbonaceous schists.



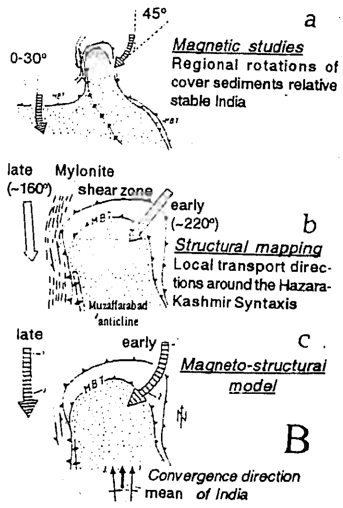


Fig. 4.23. Models for the formation of Hazara-Kashmir Syntaxis. (From Bossart et al. 1984, 1988, 1990). **Model (A).** Arrows indicate the overthrust direction. (a) Initial overthrust geometry of the three tectonic elements. (b) Rotation of overthrust direction and related formation of the shear zone. (c) Further rotation of overthrust direction leading to the present-day geometry of the syntaxis. Note narrowing of the shear zone and formation of mylonite discontinuity. Distribution of metamorphic facies (1) very low grade, (2) prehnite-pumpellyite, (3) transition prehnite-pumpellyite to greenschist, (4) greenschist, (5) synkinematic greenschist overprinted on amphibolite, (6) amphibolite.
Model (B). Syntaxis derived from a combination of (a) regional model and (b) local model. An early transport initially opposite to the motion of India was successively rotated to its present position (220°). The late transport direction in the west rotated little and was always approximately opposite to the movement of India.

Tahirkheli (1987, 1989, 1992) has shown that in the southern part of the Nanga Parbat Massif, the Precambrian Salkhala Formation (greenschist facies) has been thrust over the 1,800–2,700 Ma Nanga Parbat Gneisses (amphibolite facies) and he suggests that this thrust– the Shontargali Thrust is an analogue of the MCT. According to Papritz (1989), Rey (1989) and Greco et al (1989) the MCT is folded to form the Kaghan Syntaxis in the Naran Area. Fontan and Schouppe (1994) have recently described a 100 m thick shear zone in the Musa Gali area of Kaghan which they consider as MCT (Fig. 4.22).

Formation of the Syntaxis

Most of the early geologists (Wadia 1931, HSC 1960, Calkins et al. 1975) viewed the syntaxis as a product of southerly compression, folding and wrapping of the younger sediments around a northward 'tongue-like' projecting promontory of the Indian Peninsular Shield. The rigid basement buttressed the southward advance of the fold-and-thrust front of Tethyan sediments and forced the resulting structures to bend around, thus creating a hairpin bend in the strike of the structures.

Crawford (1974b) proposed that the Salt Range, Potwar and Hazara once lay in line with the Himalaya and that a 70° counter-clockwise rotation of this region occurred around a pole near the eastern end of Salt Range. The main movement followed a plane in or below the Eocambrian evaporites. Constrained by paleomagnetic research which indicates regional rotation of cover sediments, and detailed structural mapping which suggests changes in local transport direction, Bossart et al (1984, 1988, 1990) have proposed a model for the formation of this syntaxis (Fig 4.23). They have shown 22°–43° clockwise rotation of the sedimentary cover along the eastern limb and up to 39° anticlockwise rotation along the western limb of the syntaxis. Their model envisions an original SW direction for thrusting. The transport direction then shifted north-south forming a left lateral shear zone and further shifting of thrust direction NNW-SSE resulted in the present configuration of the syntaxis.

The formation of the Hazara–Kashmir Syntaxis began in Early Pliocene (Burbank 1983). Treloar et al. (1991d) have shown that the syntaxis is the result of Early Pliocene thrusts. The thrust systems to the east of the syntaxis have different geometries from those to the west. On the eastern side MBT was the main structure, whereas on the western margin, the simultaneously propagating structures, such as the Salt Range Thrust and related breakback structures, were in the footwall of the MBT (Fig. 4.23). There is paleomagnetic evidence of a clockwise rotation of 45° within the core of the syntaxis (Bossart et al. 1990). Treloar et al. (1991d) point out that this contradicts the anticlockwise rotation model of Bossart et al. (1990).

Read it

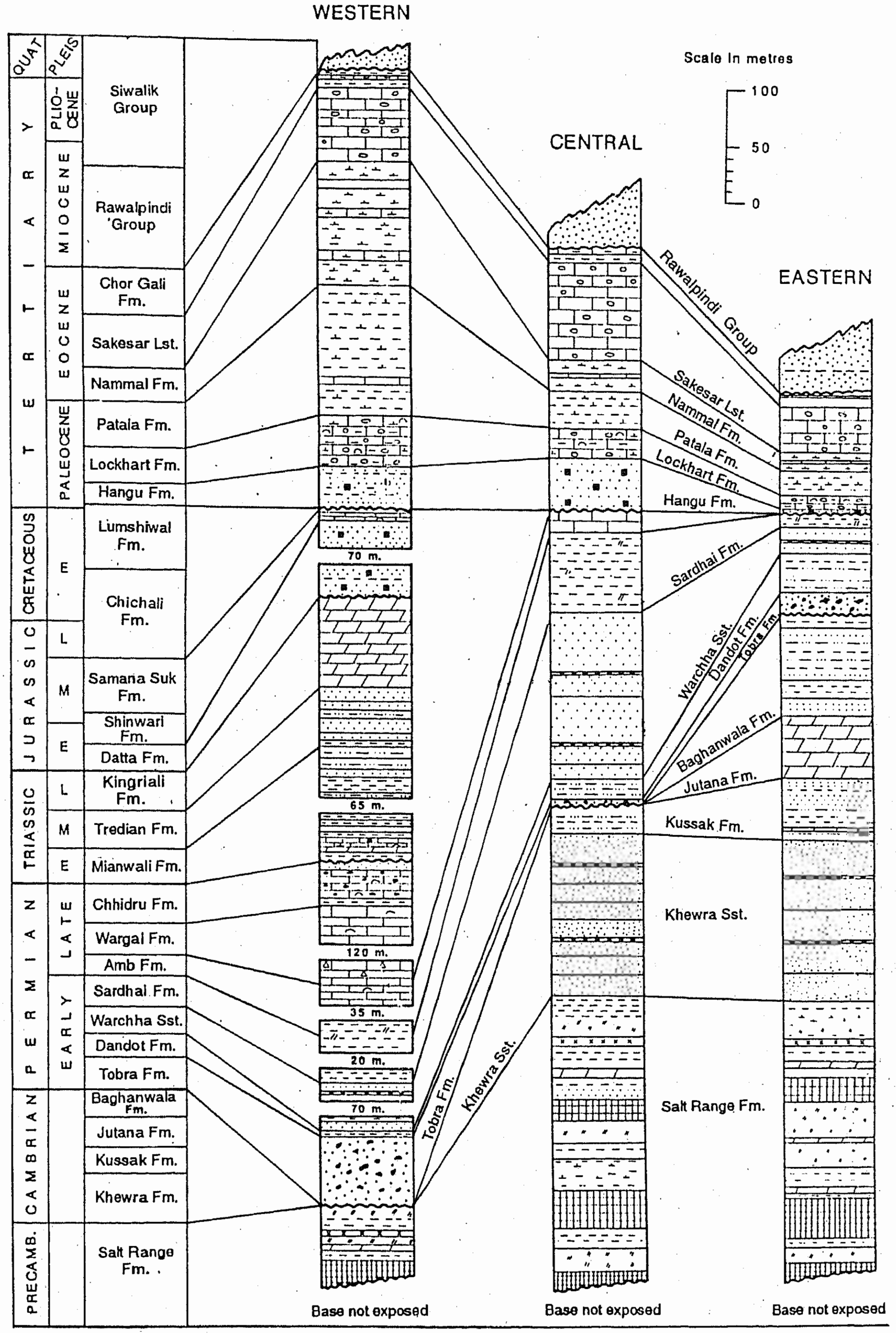
Salt Range and Kohat-Potwar Fold Belt

This east-west trending fold belt comprises the low rolling hills and valleys of the uplifted Kohat-Potwar Plateau, the Salt Range and its westward extensions (Fig. 4.6 and Photo. 11). It is about 85 km wide and extends for about 200 km. It is a discrete structural zone bounded in the north by the north-dipping Main Boundary Thrust (Sarwar et al. 1979, Yeats et al. 1984, Coward et al. 1985). Southward the Salt Range Thrust, Kalabagh Fault and the Surghar Thrust form its southern boundary. West and eastward it is terminated by the N-S oriented Kurram Thrust and Jhelum Fault respectively (Kazmi and Rana 1982).

Stratigraphic Setting

A sedimentary sequence ranging from Eocambrian to Recent is exposed in the Salt Range and Kohat-Potwar Plateau (Fig. 4.24, Table 4.3). Exposures of Mesozoic and earlier rocks are largely confined to the southern margin of Salt Range, Surghar Range and Khisor Range. These comprise Eocambrian evaporites (Salt Range Formation) and shallow marine to non-marine Lower to Middle Cambrian sequence of dolomites, shales and sandstones (Jhelum Group) which are unconformably overlain by a thick Permian clastic and carbonate succession (Nilawahan and Zaluch Groups). The angular unconformity at the base of Permian is characterised by the widespread Talchir boulder bed

Table 4.3. Columnar sections of the sedimentary sequences in the Salt Range. (From Fatmi et al. 1984).



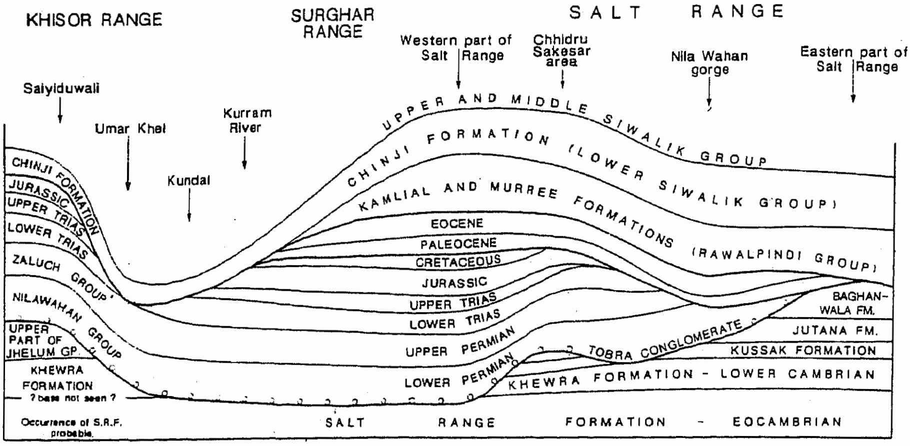


Fig. 4.24. Section showing the stratigraphic sequence of the Salt Range and Trans-Indus Ranges. (From Gee 1989).

(Tobra Formation). The exposed Eocambrian rocks are thickest (about 550 m) in eastern Salt Range and thin out westward due to pre-Talchir erosion (Fig. 4.24). The Permian sequence is thickest (700 m) in western Salt Range but it becomes thinner eastward.

A paraconformity separates the Mesozoic shelf (Tredian and Datta Formations) and shallow marine deposits (Mianwali, Kingriali, Shinwari and Samana Suk Formations) from the Paleozoic sequence. The exposed Mesozoic sequence is thickest (1,000 m +) in western Salt Range, but it has been greatly attenuated eastward due to erosion and overlap by the Cenozoic sequence (Fig. 4.24). There are a number of minor unconformities indicating disruption in sedimentation and instability in the foreland sedimentary basin during the Mesozoic.

The Cenozoic sedimentary rocks consist of a 125 to 400 m thick Paleogene sequence, deposited in widely varying paleoenvironments, ranging from shallow marine (Lockhart, Nammal, Sakesar, Chor Gali, Kohat Formations), marine to continental (Hangu, Kuldana Formations), marine to lagoonal (Patala Formation), isolated desiccated basins (Panoba Shale, Bahadurkhel Salt, Shekhan Limestone, Jatta Gypsum), fluvial and deltaic (Mami Khel Formation), to continental molasse type (Rawalpindi and Siwalik Groups). The Cenozoic sequence is about 8,000 m thick (Shah 1980). In the western Salt Range, a significant angular unconformity above the Cretaceous (Fig. 4.24) cuts down-section with progressive eastward overlap of Paleogene sequence over successively older rocks until the Paleogene (Lockhart Formation) directly overlies the Permian (Sardhai Formation). Another important angular unconformity separates the Neogene Siwalik molasse from the underlying older formations. In the eastern part of Kohat-Potwar fold belt the Siwaliks overlie Cambrian rocks, whereas in the western part (Surghar and Khisor Ranges) they overlie Triassic/Jurassic rocks.

The stratigraphic sequence in Kohat-Potwar fold belt is highly fossiliferous (see Chapter 5). The Salt Range contains a perfect Permian section and is considered a type locality for the Permian system on a world-wide basis. It contains a wide-ranging rich fauna. The lower Permian contains a cold-water flora and fauna, whereas the upper Permian is characterised by a warm-water flora which indicates shifting of Gondwanaland to lower latitudinal areas due to rotation (Nakazawa 1985). The Lower Eocene

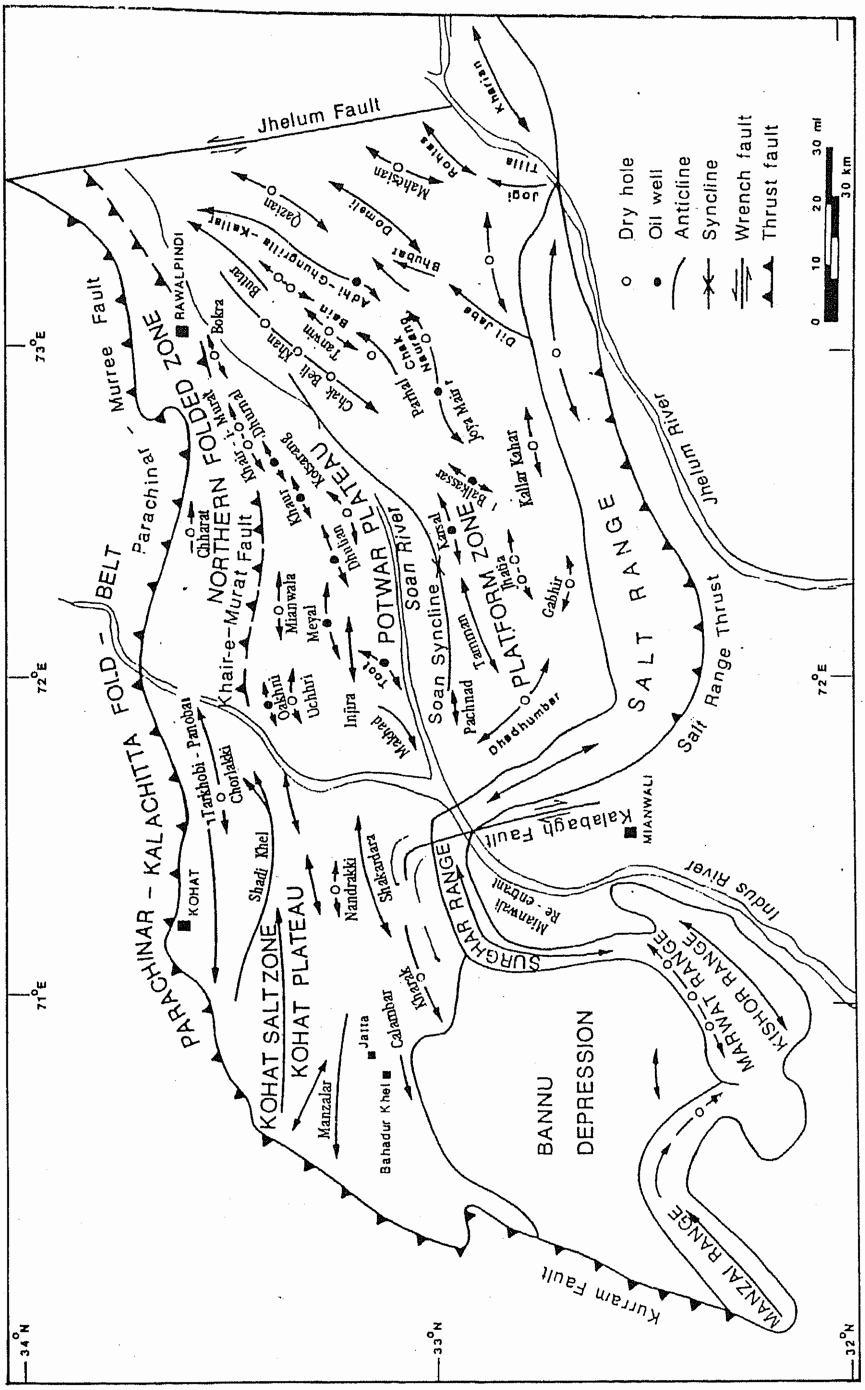


Fig. 4.25. Structural map of the Kohat-Potwar Plateau. (From Kazmi and Rana 1982, Khan et al. 1986).

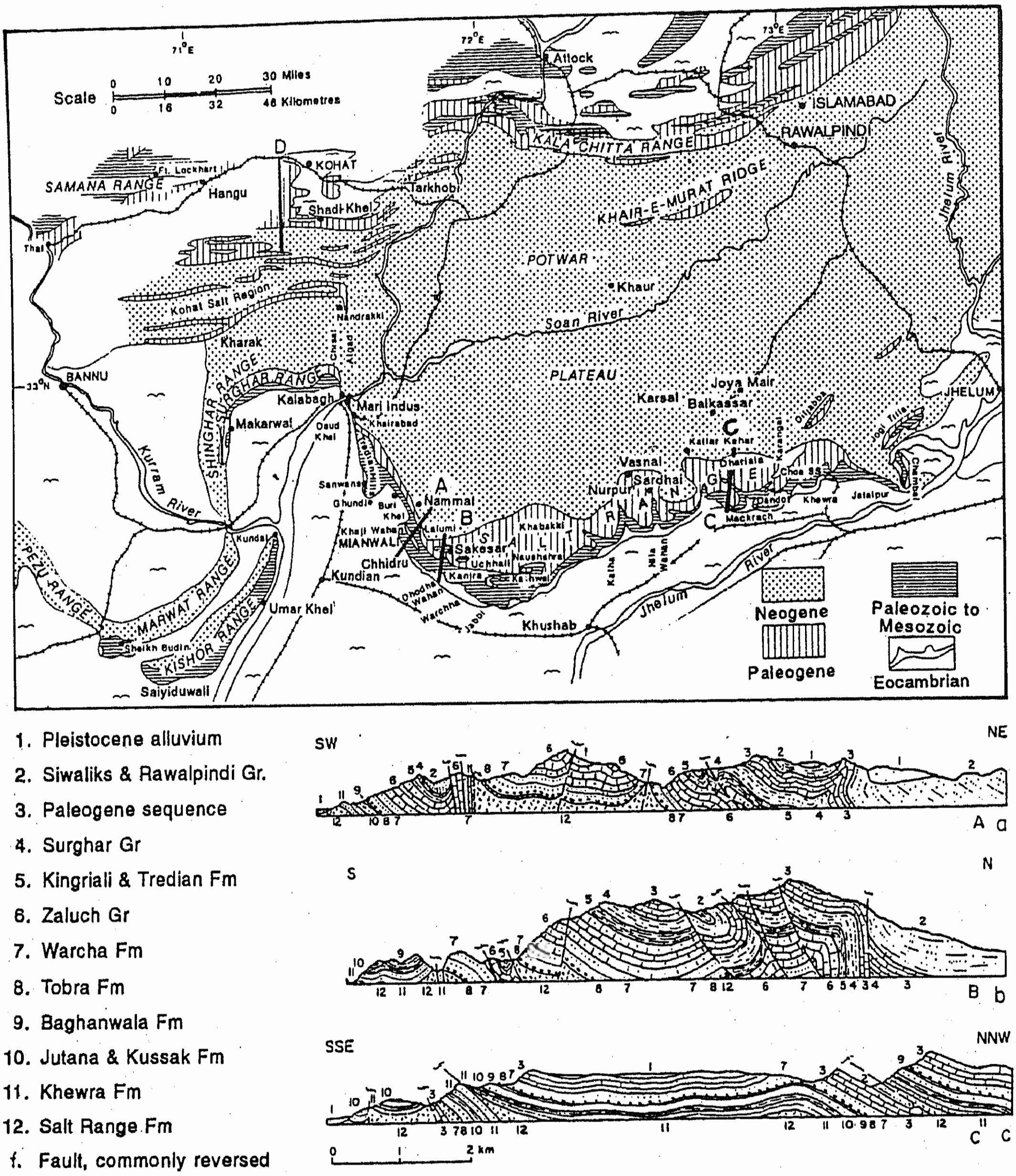


Fig. 4.26. Geological sketch map and sections of the Salt Range and Kohat-Potwar Plateau. (Modified from Gee 1945, 1989. For section D see Fig. 4.31).

sequence (Kuldana Formation) has yielded an interesting vertebrate fauna which includes three genera of primitive cetaceans (whales) *Pakicetus*, *Gandakasia*, and *Ichthyolestes* (Gingerich et al. 1990, 1994). Pakicetus is believed to be the oldest and most primitive cetacean known. Other vertebrate fossils found with the cetaceans are rodents, bunodonts, artiodactyls and anthracobunid proboscideans.

The Siwalik rocks of this region contain a rich mammalian fauna which has attracted scientists for more than a century. Of particular interest are hominoid primates whose earliest remains date back to 14.5 m.y. One of these, *Ramapithecus panjabicus*, is widely believed to be the earliest recognisable hominoid or human ancestor (Raza et al. 1983).

The stratigraphic sequence of Kohat-Potwar region also contains record of interesting and unique events. The Talchir Boulder Bed (Tobra Formation), near the base of the Permian, records a global glacial event and has been instrumental in provoking reconstruction of the Gondwana supercontinent. The 15–20 m thick Bain Boulder Bed in Pleistocene Marwat Formation (Morris 1938) in Bhittani and Shinghar Ranges have been interpreted by Nio and Hussain (1984) as a channelled debris flow deposit formed by a single catastrophic flood event. Khan et al (1985) identify this as Bain diamictite which formed due to a catastrophic lahar event (volcanic debris flow following volcanic eruption). The Plio-Pleistocene Dasht-i-Nawar Caldera, about 300–400 km to the west in Afghanistan, is believed to be the most likely source of this diamictite.

Geological Structure

The Kohat-Potwar fold belt is composed of a series of east-west trending folds and thrust faults. Its southern part comprises the Salt Range composite orocline which forms a serpentine chain of hill ranges, namely the Manzai, Bhittani, Khisor-Marwat, Surghar and the Salt Range (Fig. 4.25 and Photo. 11). These ranges consist of Eocambrian to Paleogene rock formations and constitute a distinct geotectonic zone. To their north lie the undulating Kohat and Potwar Plateaus which are separated by Indus River. The Kohat–Potwar Plateau is largely covered by Neogene Siwalik molasse, which forms several large folds, some of which are faulted. The very shape and orientation of these surface structures define a number of structural zones within the Kohat–Potwar Plateau (Fig. 4.25). Seismic reflection profiles and oil well data suggest that these zones reflect differences in deeper structures and lithostratigraphy.

The Salt Range

This range is essentially a complex salt anticlinorium with a series of salt anticlines (Fig. 4.26). It is widest in its central part, between Khewra and Warcha, where it also contains the best exposures of Paleozoic and Eocambrian sequence. The structure along its northern slope is comprised of simple, broad, shallow folds followed by a gentle monocline. Southward the folding becomes tighter and the folds are commonly faulted. Along the southern scarp the structures are more complicated and comprise east-west trending faults and overfolds. The Eocambrian evaporites are exposed in some of these overfolded and faulted anticlines (Fig. 4.26). Though the general trend of the folds is east-west in the Central Salt Range, a few north-south trending and northward plunging anticlines, which are actually "nose" type structures, have also formed.

Eastward the Salt Range loses its stature and bifurcates into two narrow northeast trending ridges, the Diljabba and the Chambal–Jogi Tilla. The latter comprises steeply

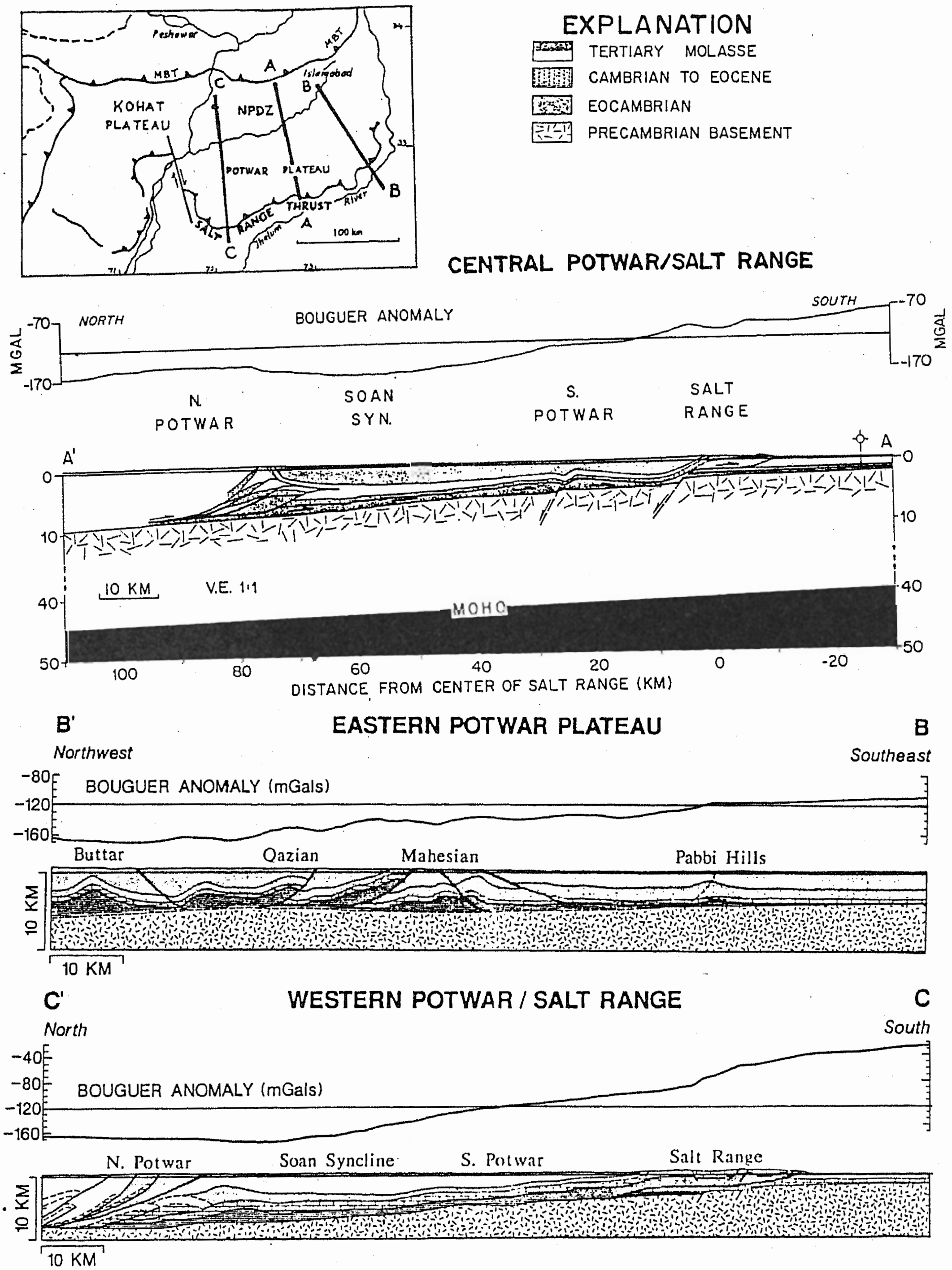


Fig. 4.27. Bouguer gravity profiles and sructural cross-sections of Potwar Plateau and Salt Range. (From Lillie et al. 1987).

dipping monoclines, complicated by complex thrusts and tear faults, whereas the Diljabba Hill is a steeply dipping anticline traversed by Diljabba–Domeli Thrust (Figs. 4.25 and 4.26). Westward the Salt Range takes a northwest bend near Warcha. Its structure remains the same and it is separated by the Kalabagh Fault from the Trans-Indus Ranges. Southward the Salt Range is truncated by the Salt Range Thrust.

Salt Range Thrust: This thrust fault runs along the southern margin of the Salt Range, between Jhelum and Indus Rivers (Fig. 4.26), and it has pushed the older rocks of the Salt Range upon the less deformed Tertiary sequence of the south-lying Jhelum Plain. The thrust zone is largely covered by Recent fanglomerates. However at places (e.g., near Jalalpur and Kalabagh), the thrust is exposed and shows the Paleozoic rocks overlying the Neogene or Quaternary deposits of the Jhelum Plain (Gee 1945, 1989, Yeats et al. 1984). Seismic reflection profiles, gravity, and drill hole data indicate that the Salt Range and Potwar Plateau are underlain by a decollement zone within Eocambrian evaporites. Along the Salt Range Thrust, effective decoupling of sediments from the basement along the salt layer has led to southward transport of the Salt Range and Potwar Plateau in the form of a large slab over the Jhelum Plain (Fig. 4.27). The Salt Range is thus the surface expression of the leading edge of a decollement thrust (Lillie et al. 1987).

[illegible]

Kalabagh Fault: This fault forms the western margin of the Salt Range and extends NNW from near Mianwali for a distance of 120 km (Fig. 4.28 and Photo. 11). It has been described as an active dextral wrench fault associated with several recorded earthquake epicentres (Kazmi 1979c). It has a long southward continuation as indicated by lineaments on Landsat photographs and buried dextral wrench faults inferred from seismic data (Seeber et al. 1979, Kazmi 1979a, Kazmi and Rana 1982). It cuts folds and faults in the Eocambrian to Quaternary rocks. Tectonic slivers of Permian and older rocks occur along the fault zone (Gee 1980). In its northern segment, the Kalabagh Fault has affected the Quaternary deposits as indicated by uplifted stream terraces, tectonic blocks of evaporite and limestone, and truncated alluvial fans. Some of these deposits (Kalabagh Conglomerate) have been dated as 2.1–1.9 Ma which also constrains the age of the Kalabagh Fault. The younger deposits are not affected by the fault. Offset of Quaternary deposits suggested 16–19 km of strike-slip movement along the fault during the Quaternary (McDougall and Khan 1990). [illegible]

Southward the Kalabagh Fault apparently displaces the Salt Range Thrust. Near its southern end (north of Khairabad), the Kalabagh Fault splays out and forms two additional subparallel faults, the Dinghot and Ainwan Faults (Fig. 4.28). Near its northern end, the Kalabagh Fault bends westward and branches out into a number of smaller, north-dipping thrust faults. About 2.5 to 3.5 km west of Kalabagh Fault and parallel to it, there is a smaller right lateral strike-slip fault which has dislocated the Surghar Range.

According to McDougall and Khan (1990), the Kalabagh Fault system forms a lateral ramp extending to the base of Salt Range-Potwar Plateau allochthon. They interpret the residual gravity anomalies in this region in terms of a NNW-trending discontinuous basement ridge that rises 2 km above the 5 km deep basement. According to this model the Kalabagh Fault over-rides this ridge (Fig. 4.28).

Trans-Indus Ranges and Bannu Plain

This region is composed of the Surghar, Marwat, Khisor, Pezu and Manzai Ranges

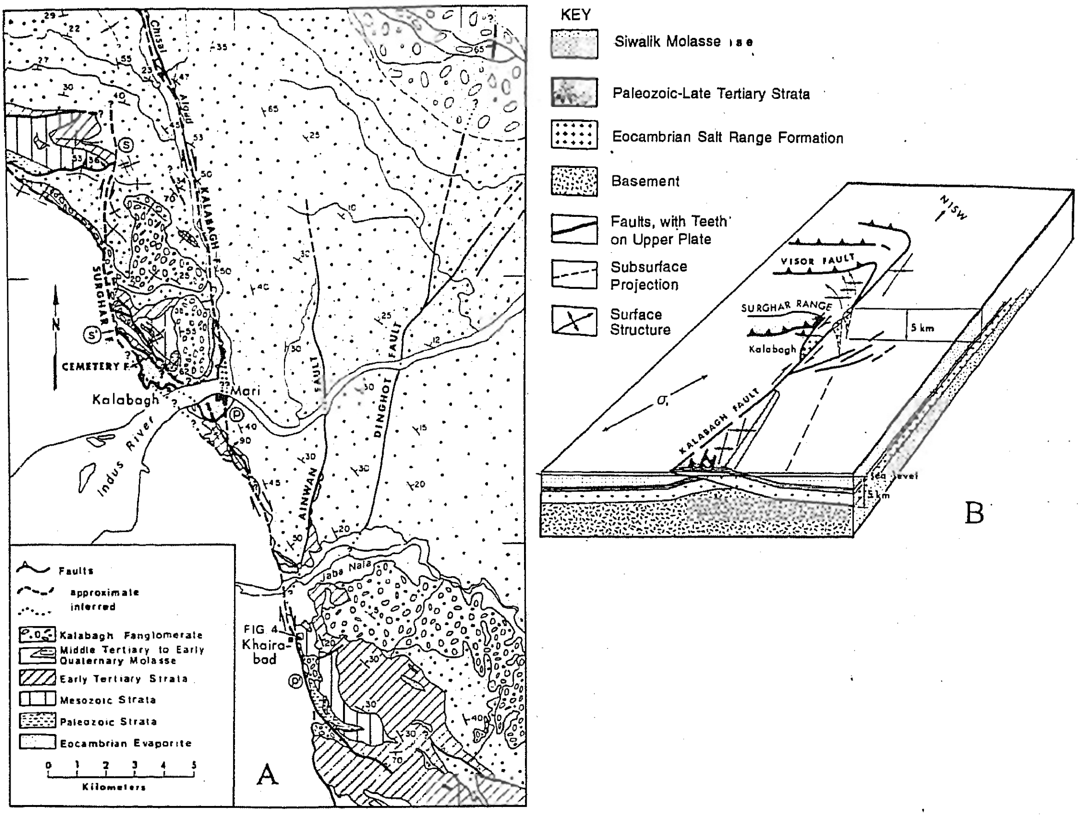


Fig. 4.28. Geological map and schematic diagram of Kalabagh Fault zone. (Modified from McDougall and Khan 1990).

A- Map of the Kalabagh area with displacement of piercing points along Kalabagh Fault (p-p'), and along Surghar Fault (s-s'). The Kalabagh Conglomerate is not offset by the Surghar Fault, but is offset 12-14 km by the Kalabagh Fault.

B- Schematic drawing of the Kalabagh Fault lateral ramp. The basement ridge extends to the Kalabagh area where it dies out. Depths to basement and approximate thicknesses of section are projected from reflection seismic and well data east of the area.

which form an 'S' shaped double re-entrant and surround the Bannu depression (Fig 4.25). The exposed stratigraphic sequence is similar to that of the western Salt Range, except that the Rawalpindi Group (Miocene) is missing in the southern part of his region and the Siwalik Group unconformably overlies upper Cretaceous rocks (Fig. 4.24).

The Surghar and Khisor Ranges are asymmetrical, overfolded anticlines, with Permian strata exposed in the core and overlain by Mesozoic and Paleogene rocks. The Marwat Range is an anticlinal feature, largely covered by the Siwaliks. Late Permian to Cretaceous rocks are exposed near its southwestern end and are unconformably overlain by the Siwaliks. The Bhittani and Manzai Ranges are also anticlinal and are entirely covered by Siwaliks. The Bhittani anticline has been faulted (Hemphill and Kidwai 1973). The Bannu Plain is covered by alluvium , and as indicated by seismic surveys, it is followed in depth by a sedimentary sequence similar to that of the surrounding hills (Khan et al. 1986).

A north-dipping thrust, the Surghar Thrust is located along the southern margin of the Surghar Range. It is probably the western extension of the Salt Range Thrust and continues along the southern/eastern margin of the Trans-Indus Ranges (Gee 1989).

Potwar Plateau

This plateau is roughly defined by the rivers Indus and Jhelum to the west and east respectively, the Kalachitta–Margalla Hill Ranges to the north, and the Salt Range to the south (Photo. 2.11). It is largely covered by the Siwalik sequence, though at places upper Eocene shales and limestones crop out locally in folded inliers. Its northern part, known as the North Potwar Deformed Zone (NPDZ) is more intensely deformed. It is characterised by east-west, tight and complex folds, overturned to the south and sheared by steep-angle faults.

NPDZ is followed to the south by asymmetrical, wide and broad Soan syncline, with a gently northward dipping southern flank along the Salt Range and a steeply dipping northern limb along NPDZ (Fig. 4.25). In the western part this basin is comprised of several east-west, broad and gentle folds (wave length 26–40 km). In its eastern part the strike abruptly changes to the northeast and the structures comprise tightly folded anticlines and broad synclines (fold wave length 10–12 km). Axial zones of most anticlines dip steeply or are overturned. Faulting of the anticlines is rare (Pennock et al. 1989). This east to west difference in the structural style has been attributed to the reduced thickness of evaporites and lesser basement slope in the eastern part of the Potwar and Salt Range. Increased drag at the base of the section has formed relatively complicated structures due to greater internal deformation (Lillie et al. 1987).

Integration of seismic reflection, gravity and drill hole data with surface geology shows that the Salt Range and Potwar are underlain by a gentle (1°–4°), northward-dipping basement, with an upward convexity, and traversed by north-dipping normal faults (Fig. 4.27). Above the basement there is a decollement zone in Eocambrian evaporites which are overlain by an overthrust wedge of narrow cross-sectional taper. The Eocambrian evaporites have been an effective zone of decoupling allowing thrusting to extend more than 100 km south of MBT without involving the basement. The Salt Range is the topographic expression of this great thrust sheet, riding up and over a down-to-the-north basement fault. This fault, with 1 km offset of the basement, is located to the north of the Salt Range and has functioned as the thrust ramp. Its position ties with the steepened (30°–45°), monoclinal dips on the northern flanks of the Salt Range (Lillie et al. 1987). Beneath NPDZ the Phanerozoic sedimentary wedge thickens to 9 km and forms a north-dipping stack of thrust faults, some of which reach the surface while others terminate at depth as blind thrusts (Fig. 4.27).

In central and western Potwar, the thrust wedge has been transported southward as a coherent slab with little internal deformation and less than 1 km of shortening between NPDZ and Salt Range (Baker et al. 19[illegible] In eastern Potwar, in contrast, the deformation has been telescoped due to increased basal traction and faults cut upsection, producing a different structural style characterised by fault-folds, triangles and pop-up zones In

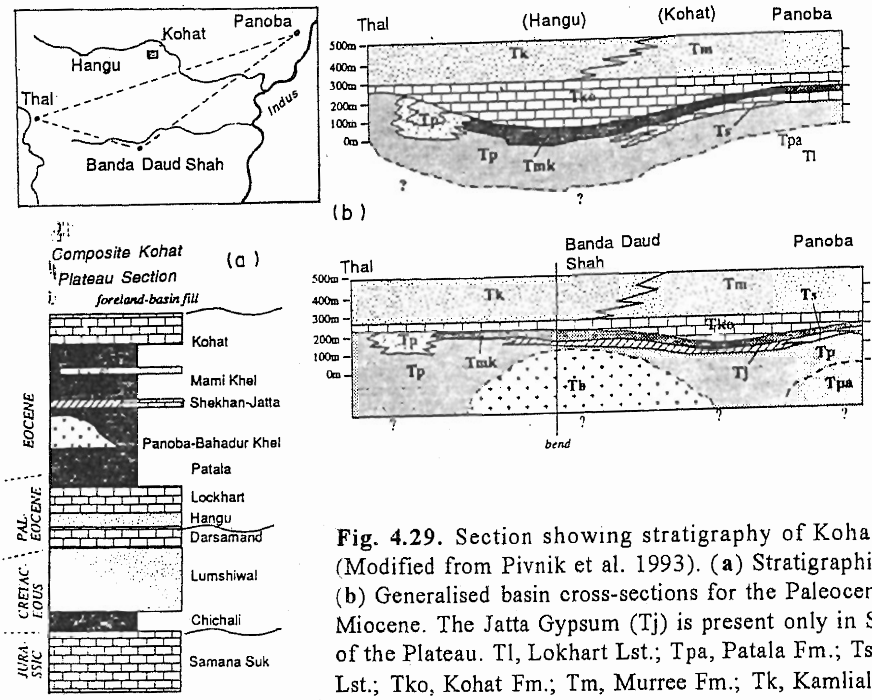


Fig. 4.29. Section showing stratigraphy of Kohat Plateau. (Modified from Pivnik et al. 1993). (**a**) Stratigraphic column. (**b**) Generalised basin cross-sections for the Paleocene through Miocene. The Jatta Gypsum (Tj) is present only in SE portion of the Plateau. Tl, Lokhart Lst.; Tpa, Patala Fm.; Ts, Shekhan Lst.; Tko, Kohat Fm.; Tm, Murree Fm.; Tk, Kamlial Fm.

this region 24 km of shortening has occurred (Pennock et al. 1989).

Kohat Plateau

West of the Potwar, the Kohat Plateau is comprised of Eocene and younger sedimentary rocks deposited in a tectonically restricted basin (Figs. 4.29, 4.30 and Photo. 11). The Eocene sequence contains evaporites (Bahadur Khel Salt and Jatta Gypsum) which are restricted to the southern part of the plateau (Fig. 4.29). On the surface these rocks form east-west trending, gentle to steeply dipping, doubly plunging, overturned folds tens of kilometre long.

The structures in the northern part are significantly different from those in the southern part. The northern region is characterised by tight, commonly overturned folds, out-of-syncline faults and several thrust faults. Some of the low angle thrust faults have been folded and form klippen (Fig. 4.31). They constitute a distinct thrust belt, the Mir Khweli Sar Thrust Belt (MKSTB). According to Pivnik et al. (1993) it is a compression-related relict, thin-skinned thrust belt. There are few faults outside the MKSTB because the Eocene Panoba Shale serves as a detachment horizon in which faults flatten near the surface and are not exposed. The Eocene evaporite sequence is greatly reduced or missing in the north. Instead the Panoba Shale is exposed in the anticlinal cores.

In the southern part of the Kohat Plateau east-west trending folds and north- and south-dipping reverse faults are common. Most of the faults tip out laterally into anticlines (fault propagation folds). The Bahadur Khel Salt is exposed in anticlinal cores whereas Jatta Gypsum is commonly imbricated and folded with slivers of Panoba Shale. In this region, the Lower Eocene rocks have been thrust over the Miocene molasse at several places (Pivnik et al. 1993).

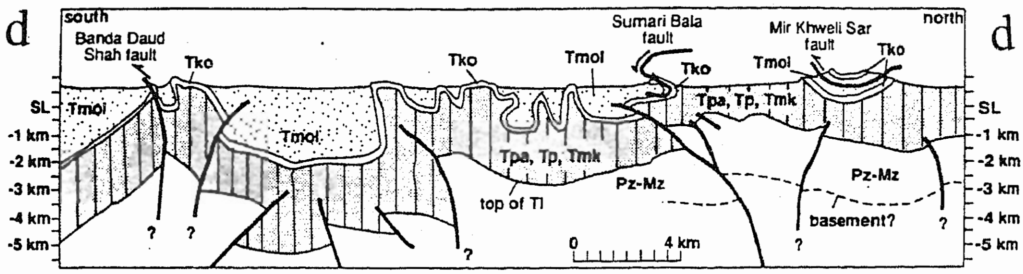


Fig. 4.30. Geological cros-section across the Kohat Plateau (from Pivnik et al. 1993). Tmol, Rawalpindi and Siwalik Gr ; Tko, Kohat Fm ; Tmk, Mamikhel Fm ; Tp, Panoba sh ; Tps, Patala Fm ; Tl, Lochart Fm ; Pz-Mz, Paleozoic to Mesozoic. (see Fig. 4.25 for location).

North of Shakardarra the structures have an east-west orientation. However, near Shakardarra, the structures abruptly change direction from east-west to north-south. This region is traversed by a large thrust fault, the Hukni Fault. According to Abbasi et al. (1991) the Hukni Fault is located above a north-south lateral ramp in the basement and the swing in strike near Shakardarra records progressive fading away of the lateral and frontal ramp complex at depth (see section on Kalabagh Fault).

With limited seismic data and the assumption that (a) low-angle faulting has been dominant, (b) there has been insignificant out-of-section transport of material and (c) the Eocambrian evaporites are present in depth, Baker et al. (1988), Abbasi et al. (1991) and McDougall et al. (1991) have used imbricate thrust and passive-roof duplex geometries in construction of balanced structural cross-sections of the Kohat Plateau (Fig. 4.31). Based on extensive field surveys and seismic profiles across the Kohat Plateau, Pivnik et al. (1993) demonstrated the absence of duplexes, passive roof thrusts, antiformal stacks related to north-south progressive thrusting in this region. According to them the Precambrian evaporite has not been identified in this region. The Kohat basement is traversed by high angle conjugate reverse faults with throw of up to 500 m. This shows a significant degree of wrench faulting and positive flower structures. Thus the geometries of structures in Kohat Plateau are the product of both compressional and transpressional tectonics.

Evidence of Deformation from the Stratigraphic Sequence

In the Kohat-Potwar fold belt, apart from the geological structure, evidence for deformation has accumulated from stratigraphic features as well. It includes unconformities, tectonic rotation, and changes in paleocurrents, provenance facies and sediment accumulation rates constrained through paleomagnetic dating. These studies provide a rational basis to reconstruct the deformational history of the Kohat–Potwar fold belt. A brief chronology of the tectonic events has been summarised in Table 4.4.

Kurram–Cherat–Margalla Fold-and-Thrust Belt

This arcuate and narrow (20 to 30 km wide) thrust belt lies to the north of Kohat-Potwar fold belt. From near Balakot (Hazara–Kashmir Syntaxis), it extends southwestward through Margalla Hills, Attock-Cherat and Kalachitta Ranges to the Sufaid Koh Range on Afghanistan border, a distance of about 350 km (Fig. 4.6) and Photo. 11). It is an intensely deformed and tectonised belt with isoclinal folds and several south-verging

Table 4.4. Chronology of significant events in Kohat-Potwar Fold Belt and adjacent regions.

Age (Ma)	Event	Manifestations	References
1.2 to 0.4 1.7 - 0.4 < 2.4 - < 0.7 < 2.4 - < 1.5	Folding/uplift Pabbi Hills Folding/uplift Rohtas Hill Folding/Uplift Chambal Ridge Folding/Uplift Mangla-Samwal	Change in fluvial sedimentation rate, attainment of surface expression and initiation of degradation.	-(1)
2.0 - 1.0	Thrusting–Jogi Tilla Thrust, Chambal Thrust, Salt Range Thrust	Tectonic rotation, changes in paleocurrents, appearance of Talchir-bearing conglomerate.	-(2)
2.0 - ~ 0.1	Strike slip faulting-Kalabagh Fault	Deformation of Late Pleistocene to Recent conglomerate.	-(3)
2.1 - 1.8	-Accelerated uplift of Pir Panjal -Major movement along MBT -Kalachitta uplift -Soan Syncline uplift	Syndepositional unconformity, sediment accumulation declines, basinward paleocurrents, Lei Conglomerate, lacustrine deposits in ponded basins, in Bannu changes in fluvial regime, 1.9 Ma Lei Conglomerate unconformable over 2.1 Ma folded strata.	-(4),(5),(6).
2.5	Thrusting-Jogi Tilla Thrust	Tectonic rotation.	-(4)
3.5 - 3.0	Thrusting - Riwat fault	Syndepositional unconformity (~3.5 Ma), change of provenance (limestone conglomerate replaced by crystalline conglomerate), changes in drainage, tectonic rotation of strata below unconformity.	-(4)
5.4 - 4.5	Thrusting - Salt Range Thrust, MBT. Uplift—Salt Range, Pir Panjal. Basin formation—Potwar, Attock, Kashmir	Angular unconformity, tectonic rotation of strata beneath unconformity, facies and provenance changes, increase in sedimentation rate, facies migration.	-(4), (7), (8).
7 - 5	Uplift - Pir Panjal	Pre-Karewa thick paleosols, 45° clockwise tectonic rotation.	-(9)
7.4	Changes in mammalian fauna.	Major mammalian fauna turn over event	-(10)
9 - 11	Renewed uplift of Kohistan, N. of Potwar	Changes in provenance, arkosic sst+high feldspar, blue-green hornblende. Major mammalian faunal turnover event (9.5 Ma).	-(11), (12), (13). -(10).
13 - 14	Renewed uplift and unroofing of Himalaya	Change in provenance, minor mammalian turn over event (13.2 Ma).	(10).
15-	Renewed uplift and unroofing—Kohistan Arc	Change in provenance, high grade rock fragments in sediments.	(11), (14).
>20	Changes in mammalian fauna	Major mammalian fauna turn over event.	(10)
26	Rapid uplift of Himalayas	Angular unconformity base Murree Fm., change marine to fluvial continental deposits; quartz, garnet, tourmaline rich and clasts metamorphic rocks from Indian plate margin. Rapid uplift S. of MMT, accelarated uplift of Nanga Parbat (fission track and Ar/Ar ages).	(11), (14), (15), (16).

References: (1) Johnson et al (1979), (2) Burbank and Beck (1989), (3) McDougall and Khan (1990), (4) Burbank et al. (1988), Johnson (1982), (6) Abbasi and Friend (1989), (7) Johnson (1986), (8) Opdyke et al. (1982), (9) Burbank et al. (1986), (10) Barry et al. (1985), (11) Abbasi et al. (1989), (12) Abid et al. (1983), (13) Cerveny et al (1989), (14) Zeitler et al. (1982), (15) Zeitler et al. (1985), (16) Zeitler et al. (1989).

thrust sheets. Eastward it has been cut by the Jhelum Fault. Southward it has been thrust over the Kohat-Potwar fold belt. This thrust zone is now being referred to as the MBT (Lillie et al. 1987, Yeats and Hussain 1987, Burbank et al. 1988, McDougall and Hussain 1991, Abbasi and McElroy 1991).

Tectonostratigraphic Setting

The Panjal Thrust and its western continuation, the Khairabad Fault, define the northern limits of Kurram–Cherat–Margalla thrust belt. Across the Panjal–Khairabad Thrust the metamorphic grade and stratigraphy differs significantly. South of the thrust, Tanawal Formation is missing and there is a wide hiatus in the Paleozoic sequence (Table 4.5). North of the Panjal–Khairabad Thrust the Precambrian Tanawal Formation crops out extensively, the Paleozoic sequence is more complete, and the region is dominated by metamorphic, magmatic and intrusive rocks (Fig. 4.32).

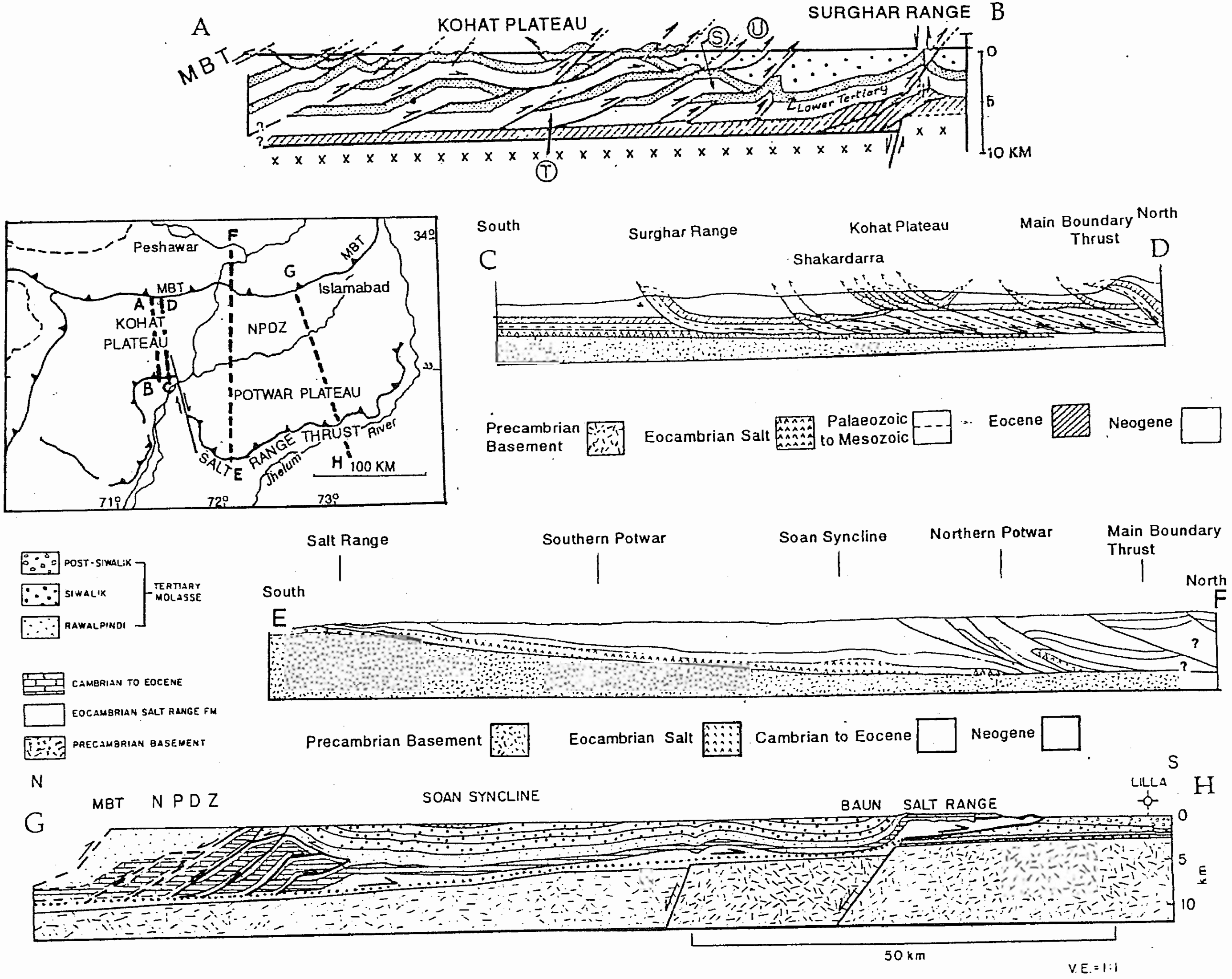
The stratigraphic sequence in various thrust blocks ranges from Proterozoic to Neogene, though the Ordovician to Permian sequence is largely missing (Table 4.5). The unfossiliferous Proterozoic rocks (Hazara Formation) are mainly comprised of a flysch type sequence of dark grey slate, phyllite, quartzite and subordinate limestone with Rb/Sr dates of 740 ± 20 and 930 ± 20 Ma (Crawford et al. 1975). These rocks exhibit low grade metamorphism and are intruded by basic dykes and sills. The Mesozoic and Cenozoic rocks consist of platform-type shallow marine to non-marine sequence characterised by many unconformities (Table 4.5), but devoid of basic dykes and sills. The Tertiary sequence (Lockhart, Patala and Murree Formations) probably represents a sediment pile derived from the northern highlands and thrust sheets (Yeats and Hussain 1987).

Geological Structure and Deformation

The geological map and sections in Figures 4.21 and 4.32 show the structure of this fold-and-thrust belt. The faults in the southern part bring strongly deformed Mesozoic and Early Tertiary rocks to the surface, while those in the north bring up Precambrian and Paleozoic sequences. Westward, in the Sufaid Koh some of the thrusts apparently merge and pass into the active right-lateral Sarobi Fault (Kazmi and Rana 1982, Yeats and Lawrence 1984).

Stratigraphic, tectonic, isotopic and paleomagnetic data indicates that the Kurram–Cherat–Margalla fold-and-thrust belt has gone through at least six periods of deformation, i.e., during Precambrian, Permo-Triassic, Late Cretaceous, pre-Pliocene, Pliocene, and Quaternary. Kazmi and Rana (1982) drew attention to these orogenic/epiorogenic, metamorphic and magmatic events which have been also confirmed by subsequent workers. The earliest orogenic event, the Precambrian Hazaran Orogeny (Baig et al. 1987, Chaudhry et al. 1989), is constrained by the angular unconformity between the Precambrian low-grade metamorphosed Hazara Formation and the unmetamorphosed Abbottabad Group. The Tanakki Conglomerate at the base of Abbottabad Group contains no cleavage but carries cleaved metamorphic clasts from the underlying Hazara Formation. In southern Hazara, south of the Panjal Thrust, the sequence can be conveniently seen near Tanakki, Sobrah, Mirpur, Sangargali and Daultmar.

The absence of post-Cambrian Paleozoic sequence south of the Panjal Thrust may be attributed to erosion during the Hazaran Orogeny. According to Baig et al. (1987) the 500–600 Ma granites north of Panjal Thrust may belong to a Late Hazaran orogenic phase.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NPDZ
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POTWAR PLATEAU
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Jhelum
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Surghar Range
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Kohat Plateau
Main Boundary Thrust
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Precambrian Basement
Eocambrian Salt
Palaeozoic to Mesozoic
Eocene
Neogene
POST-SIWALIK
SIWALIK
RAWALPINDI
TERTIARY MOLASSE
CAMBRIAN TO EOCENE
EOCAMBRIAN SALT RANGE FM
PRECAMBRIAN BASEMENT
Salt Range
Southern Potwar
Soan Syncline
Northern Potwar
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50 km
V.E.=1:1

In the Thandiani region, east of Abbottabad, there is a prominent angular unconformity between the Cambrian Hazira Formation and Jurassic Datta Formation. The entire Cambrian to Triassic sequence is missing (Figs. 4.2 and 4.32). Furthermore, the Precambrian to Paleozoic strata are intruded by basic dykes and sills which, however, do not cut the Mesozoic sequence and represent a pre-Mesozoic magmatic event. In Kaghan area, northeast of Kurram–Cherat–Margalla thrust belt, the Permo-Triassic metamorphosed (greenschist facies) sequence is unconformably overlain by unmetamorphosed Triassic limestone (Wadia 1957, Chaudhry et al. 1989). These field evidences suggest an orogenic event during Permo-Trias.

In the Attock–Cherat Range, there is an angular unconformity between Paleocene Lockhart Limestone and the Precambrian sequence (Fig. 4.32). In Kalachitta Range this limestone unconformably overlies Cretaceous and earlier rocks. The Cherat and Hissartang Faults juxtapose older rocks against Lockhart Limestone. According to Yeats and Hussain (1987), the displacement on these faults is pre-Tertiary and the Attock–Cherat Range was uplifted prior to the Paleocene. During this uplift the Mesozoic sequence was eroded and south-verging imbricate thrusting preceded the deposition of Lockhart Limestone (Yeats and Hussain 1987).

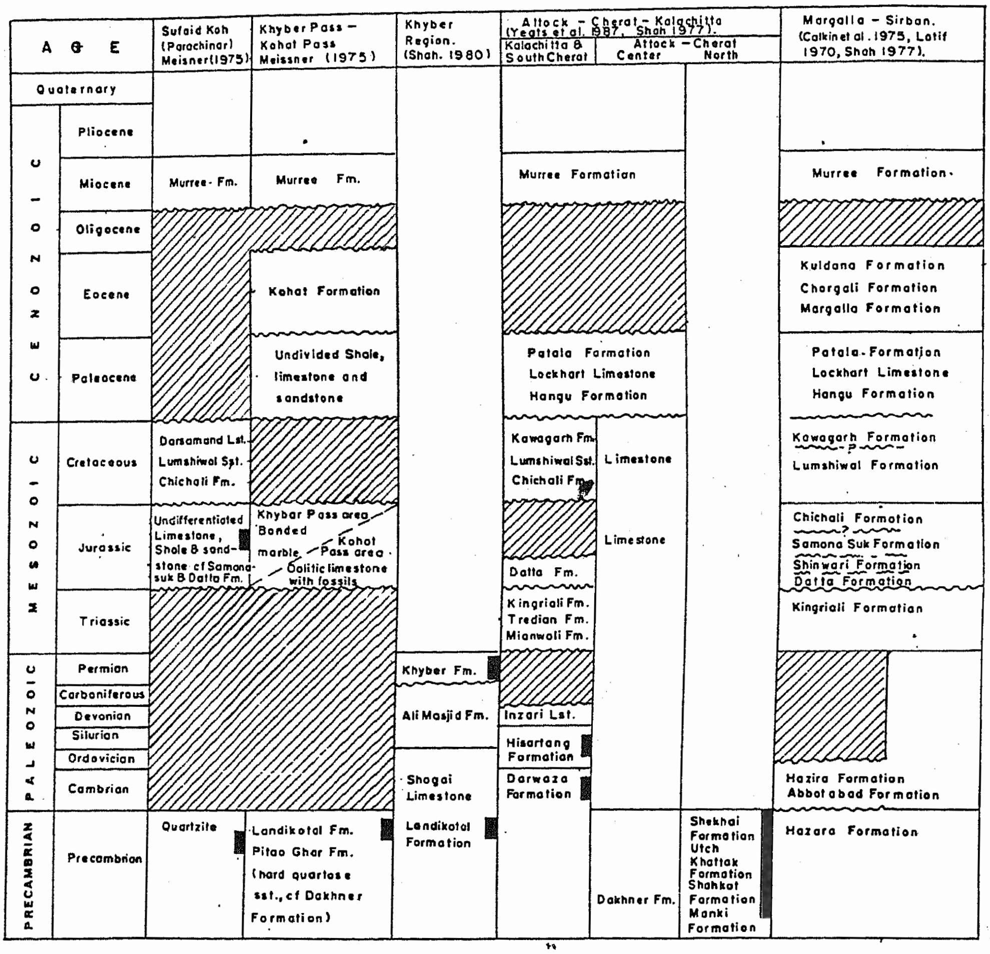
During the pre-Pliocene deformation the Miocene Murree Formation was involved in folding, imbricate thrusting and local development of slaty cleavage. This was followed by uplift of the Kalachitta Range, and thrusting and ramping on the MBT between 2.1 and 1.9 Ma (Burbank et al. 1989). The latest deformation, however, occurred during Late Quaternary after deposition of the Peshawar basin-fill, dated 2.8 to 0.6 Ma by Burbank and Tahirkheli (1985). Yeats and Hussain (1987) have reported four en echelon, left-stepping pressure ridges parallel to the northern front of the Attock–Cherat Range. The Quaternary sediments in these ridges have been folded and faulted. The left-stepping pattern suggests that they formed by oblique-slip faulting.

Himalayan Crystalline Nappe-and-Thrust Belt

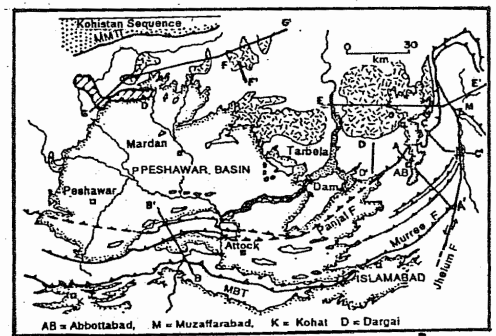
This 100 km wide, intensely tectonised zone forms the northwestern margin of the Indo-Pakistan crustal plate, and lies between the Khairabad–Panjal Thrust and the Indus Suture zone (MMT). It extends westwards from the Nanga Parbat–Haramosh Massif up to the Sarobi Fault in Afghanistan, a distance of over 450 km (Fig. 4.6). Its southern part is largely covered by Quaternary deposits of Peshawar and Haripur Basins. North of Khairabad–Panjal Thrust, these basins are surrounded by low hill ranges comprised of Precambrian metasediments and a near-complete sequence of fossiliferous Paleozoic and early Mesozoic rocks (Fig. 4.21). In the southern part, this sedimentary sequence, has been affected by low-grade metamorphism. Northwards, near the Indus Suture zone, the tectonometamorphic setting changes from an essentially sedimentary fold-and-thrust belt to a metamorphic and magmatic terrain which is characterised by thick stacks of nappes, thrust sheets and mylonitised shear zones (see Chapters 6 and 7). In this complex fold-and-thrust belt three major structural zones are quite evident. From

Fig. 4.31. Balanced cross-sections across Kohat-Potwar Plateau. AB, balanced cross-section of Surghar Range and Kohat Plateau (from McDougall et al. 1991); S= fault bend folding, T= backlimb thrusting, U= Kalabagh Fault zone. CD cross-section through Kohat Plateau to Surghar Range (from Abbasi et al. 1991). EF, cross-section from Attock-Chert Ranges to western Salt Range (from Leathers 1987). GH, balanced cross-section of Potwar Plateau and Salt Range (from Baker et al 1988).

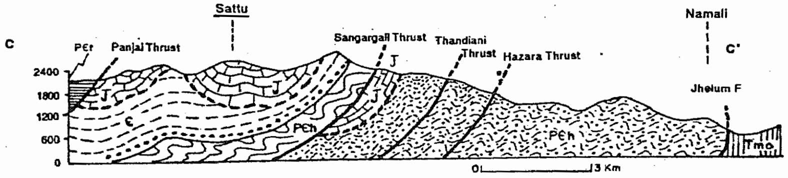
Table 4.5. Stratigraphic sequence in the Kalachitta-Margalla Fold-and-Thrust Belt. Striped columns indicate mafic dkes, triangles represent volcanic rocks.

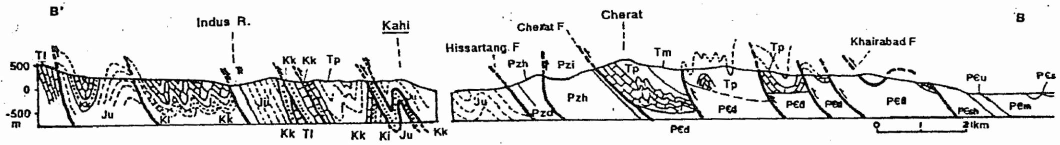


AGE		Sufaid Koh (Parachinar) Meisner(1975)	Khyber Pass – Kohat Pass Meissner (1975)	Khyber Region. (Shah. 1980)	Attock – Cherat – Kalachitta (Yeats et al. 1987, Shah 1977).			Margalla – Sirban. (Calkin et al. 1975, Latif 1970, Shah 1977).
					Kalachitta & South Cherat	Attock – Cherat Center	North	
Quaternary								
CENOZOIC	Pliocene							
	Miocene	Murree- Fm.	Murree Fm.		Murree Formation			Murree Formation·
	Oligocene	▨	▨		▨			▨
	Eocene	▨	Kohat Formation		▨			Kuldana Formation Chorgali Formation Margalla Formation
	Paleocene	▨	Undivided Shale, limestone and sandstone		Patala Formation Lockhart Limestone Hangu Formation			Patala-Formation Lockhart Limestone Hangu Formation
MESOZOIC	Cretaceous	Darsamand Lst. Lumshiwal Sst. Chichali Fm.	▨		Kawagarh Fm. Lumshiwal Sst. Chichali Fm.	Limestone		Kawagarh Formation ? Lumshiwal Formation
	Jurassic	Undifferentiated Limestone, Shale & sand-stone cf Samona-suk & Datta Fm.	Khybar Pass area Banded marble Kohat Pass area Oolitic limestone with fossils		▨ Datta Fm.	Limestone		Chichali Formation ? Samona Suk Formation Shinwari Formation Datta Formation
	Triassic	▨	▨		Kingriali Fm. Tredian Fm. Mianwali Fm.			Kingriali Formation
PALEOZOIC	Permian	▨	▨	Khyber Fm.	▨			▨
	Carboniferous	▨	▨		▨			▨
	Devonian	▨	▨	Ali Masjid Fm.	Inzari Lst.			▨
	Silurian	▨	▨		Hisartang Formation			▨
	Ordovician	▨	▨					▨
	Cambrian	▨	▨	Shogai Limestone	Darwaza Formation			Hazira Formation Abbottabad Formation
PRECAMBRIAN	Precambrian	Quartzite	Landikotal Fm. Pitao Ghar Fm. (hard quartose sst., cf Dakhner Formation)	Landikotal Formation		Dakhner Fm.	Shekhai Formation Utch Khattak Formation Shahkot Formation Manki Formation	Hazara Formation



Location map of geological sections shown in Figs. 4.32 & 4.33.





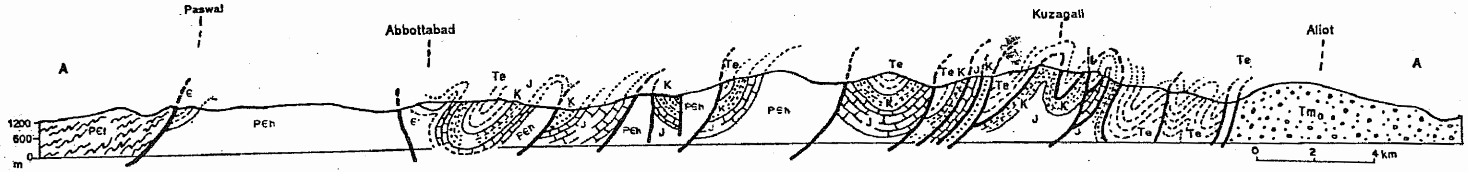


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Fig. 4.32. Geological sections across Kalachitta-Margalla fold-and-thrust belt.

AA'. Paswal-Abbottabad-Kuzagali-Aliot section. PЄt= Tanawal Fm.; PЄh= Hazara Fm.; J= Jurassic-Datta, Shinwari and Samana Suk Fms.; K= Єretaceous-Єhichali, Lumshiwal and Kawagarh Fms; Te= Paleocene to Eocene-Hangu, Lockhart, Patala, Margalla, Shekhan and Kuldana Fms., Tmo= Oligo-Miocene, Murree Fm. (modified from Latif 1970). BB'. Attock-Єherat-Kalachitta section. PЄm= Manki Fm.; PЄd= Dakhner Fm.; PЄs= Shahkot Fm.; PЄsh= Shekhai Lst.; PЄu= Utch Khattak Lst.; Pzd= Darwaza Lst.; Pzh= Hissartang Fm.; Pzi= Inzari Lst.; TR= Triassic; Ju= Jurassic; Kl= Lumshiwal Fm.; Kk= Kawagarh Fm.; Tl= Lochart Fm.; Te= Patala Fm.; Tmo= Murree Fm. (modified from Yeats et al. 1987). Є-Є'. Sattu-Sanargali-Namali section. PЄt= Tanawal Fm.;PЄh= Hazara Fm.; Є= Abbottabad Gr with Tanakki Єonglomerate at base; J= Jurassic (modified from Baig et al. 1987).

north to south these include the crystalline nappe zone, the Khyber–Hazara metasedimentary fold-and-thrust belt, and the Peshawar Basin. Regional variations and differences in stratigraphy, structural style, metamorphism and magmatism help to broadly outline a number of even smaller structural units within this belt (Fig. 4.21). These structural zones are briefly described below.

Peshawar Basin

The Peshawar Basin covers a vast area (over 5,500 km^2) in the southwestern part of this nappe-and-thrust belt. Quaternary fanglomerates form the basin margins, whereas fluvial micaceous sands, gravels and lacustrine deposits cover its central part. The fluvial deposits have a northern provenance and were probably deposited by the ancestral Kabul and Indus Rivers. However, in the southern part of the basin, fanglomerate and lacustrine deposits have been apparently derived from the Attock–Cherat and adjacent ranges (Photo. 11). Burbank and Raynolds (1988) have suggested that the Peshawar Basin sediments may have been ponded by the uplift of the Attock–Cherat Range and movement on MBT. Fault traces within the alluvial sequence indicate Late Quaternary deformation of the basin (Yeats and Hussain 1987). Magnetic-polarity stratigraphy and fission-track dates suggest that the Peshawar basin-fill is 2.8 to 0.6 Ma old (Burbank and Tahirkheli 1985). Sedimentation in southern part of the basin terminated about 600,000 years ago due to accelerated uplift of the the Attock–Cherat Range. However, along the courses of main streams, sedimentation is still continuing and Burbank (1983) has documented at least 40 catastrophic floods which inundated the basin in pre-historic times, probably within the last 60,000 years.

The Peshawar Basin contains a number of small monadnocks comprised of Paleozoic rocks, i.e., in the vicinity of Nowshera and Mardan (Fig. 4.21). The stratigraphic sequence exposed in this region has been described in Chapter 5. Inasmuch as the Peshawar Basin covers a part of the Himalayan fold-and-thrust belt, it is believed to have been carried passively on the back of low-angle detachment faults and thrust sheets, some of which find surface expression in hill ranges to the south of the basin. It is thus classified as a piggyback-type basin (Ori and Friend 1984).

Khyber–Lower Hazara Metasedimentary Fold-and-Thrust Belt

This metasedimentary fold-and-thrust belt lies to the north of Khairabad–Panjal Thrust and extends eastward from Khyber Pass region to Garhi Habibullah (Fig. 4.21). The Peshawar Basin covers a large part of this belt. To the northeast, along the Hazara-Kashmir Syntaxis, this belt is wedged in between the Panjal Thrust and Balakot Shear Zone. The Mansehra Thrust, which comprises a mylonitised shear zone, possibly an extension of the Balakot Shear Zone (Baig and Lawrence 1987), separates the Khyber–Hazara metasedimentary belt from the Hazara Crystalline Nappe Zone. Farther to the west, the northern margin of this metasedimentary belt extends up to the Besham and Swat Crystalline Nappes (Treloar et al. 1989a,c). Palmer-Rosenberg (1985) has mapped a tectonic boundary – the Nikanai Ghar Fault Complex – between Ilam Nappe (a part of the Swat Crystalline Nappes) and the metasediments (Nikanai Ghar Marble) of this belt (Fig. 4.33).

The Khyber–Hazara metasedimentary belt is largely composed of Precambrian to Early

Mesozoic sediments (Fig 4.21). The Precambrian sequence is mainly comprised of slates and phyllites with subordinate quartzites and marbles which crop out in the southern part of the belt. The Precambrian section is largely in the form of thrust blocks with variations in metamorphic grade at some places. This has led to a profusion of formation names and considerable confusion in establishing a satisfactory lithostratigraphic sequence in the region (see Chapter 5). In the eastern part of the belt (Hazara region), according to Calkins et al. (1975), Shah (1977) and Yeats et al. (1987), the Precambrian sequence is comprised of (a) quartz schist, graphitic schist, marble and quartzofeldspathic gneiss, overlain by (b) thick widespread slate, phyllite, and little-metamorphosed greywacke sandstone (Hazara Formation). This sequence is unconformably overlain by quartzites and argillites (Tanawal Formation). These rocks continue up to Tarbela.

Farther westward, a series of new names have been used to describe the sequence. In Gandghar Range, south of Tarbela, argillites, slates and phyllites of the Precambrian Manki Formation are conformably overlain by slates, phyllites and limestones/marbles of Shekhai Formation (Hylland et al. 1988). The Manki Formation continues westward into the Attock Range where it is thrust over argillites, quartzites, and subordinate siltstones and limestones of the Precambrian Dakhner Formation (Fig. 5.2). According to Yeats and Hussain (1987), Dakhner may be stratigraphically equivalent to Manki Formation, though differing in degree of metamorphism. Farther westward, in the Khyber region, a series of slates and phyllites intruded by mafic dykes, and known as Landikotal Formation (Shah 1977), are considered to be Precambrian (Pogue et al. 1992).

Inliers of a fossiliferous (conodont bearing) Paleozoic rocks occur in the hills east of Peshawar Basin (near Nowshera), in Sherwan syncline of Hazara area, and in the Khyber Pass region (Pogue et al. 1992). The Paleozoic sequence mainly comprises thick beds of argillites, marbles and dolomites (Table 5.1). A widespread amphibolite horizon occurs in the upper part of this sequence and has been named Karapa Greenschist (Pogue et al. 1992). Conodonts in the overlying and underlying formations indicate Late Carboniferous to Late Triassic age for this unit. Geochemical analyses indicate that this greenschist is a tholeiitic basalt (Ahmad et al. 1987). It has been traced northward into Swat.

The Khyber-Hazara metasedimentary fold-and-thrust belt has been intruded by mafic dykes and sills and granitic rocks of which the extensive Ambela pluton is the most conspicuous. These intrusive rocks range in age from Late Paleozoic to Early Mesozoic and have been discussed in detail in Chapter 6. This metasedimentary belt is characterised by tight, asymmetrical or isoclinal folds imbricated by several thrust faults (Fig. 4.33).

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Zone of Crystalline Nappes

This zone comprises the northwestern margin of the Indian plate and forms a relatively narrow belt south of the MMT (Indus Suture zone). It extends westward from the Nanga Parbat–Haramosh Massif and continues up to the Afghan border (Fig. 4.21). It is comprised of a thick sequence of Proterozoic basement gneisses and schists unconformably overlain by variably metamorphosed Phanerozoic cover sediments (Kazmi et al. 1984, Lawrence et al. 1989). Subduction of the Indian plate has decoupled the sedimentary cover from the basement. The cover has been imbricated with an extensive series of thrust slices and crustal nappes. Crystalline Proterozoic basement rocks are also involved in thrusting. According to Coward et al. (1982, 1985) the granitic basement

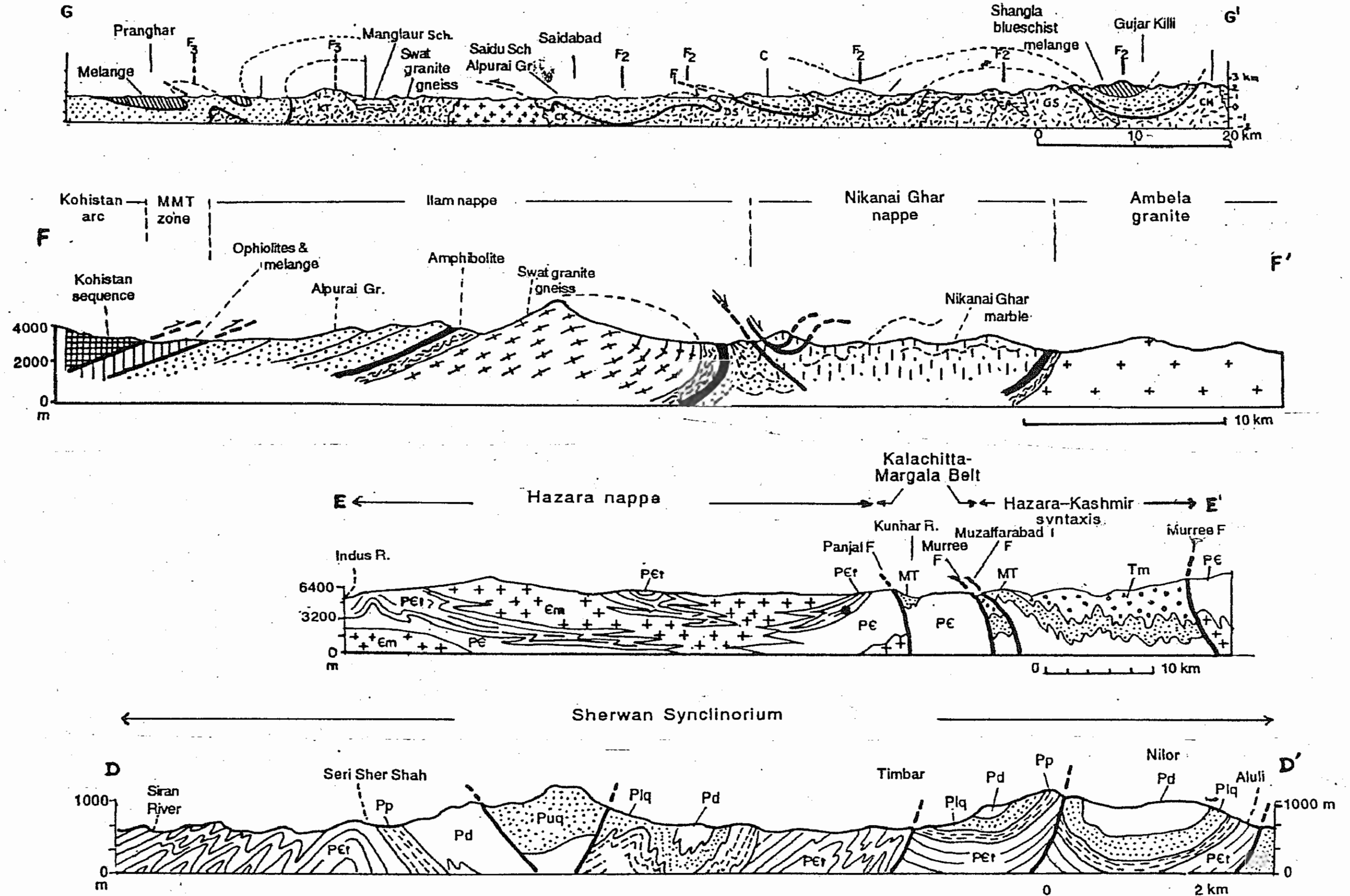


Fig. 4.33. Geological cross-sections across Swat-Hazara crystalline Thrust Belt. For location see Fig. 4.32.

D-D'. *Sherwan syncline section.* PCt= Precambrian Tanawal Fm overlain by Paleozoic to Triassic sequence as follows: Pp= phyllite, Plq= lower quartzite; Pd= dolomite; Puq= upper quartzite (modified from Calkins et al. 1975). E-E'. *Section across Hazara Nappe and Hazara-Kashmir syntaxis.* PC= Tanawal Fm,; Cm= Mansehra granite; MT= Mesozoic to Tertiary; Tm= Murree Fm. (modified from Calkins et al. 1975). F-F'. *Section across Swat Nappe and northern part of Khyber-Lower Hazara Metasedimentary Belt* (modified from Ahmad et al. 1987). G-G'. *Section across Swat Crystalline Nappe* (from Lawrence et al. (1989).

assemblage contains slices of cover rocks in synclinal folds and shear zones. In Hazara, thrust structures have developed in a large duplex with its floor at the Mansehra Thrust and roof at the MMT. Restoration of this thrust system in balanced cross- sections indicates a shortening in excess of 470 km (Fig. 4.34).

Collision and subduction have thickened the Indo-Pakistan plate margin through formation of a thick sequence of ductile mylonites, and imbrication of cover and basement along north-dipping crustal-scale thrust stacks. The latter are comprised of a number of lithologically distinct nappes which form a 5 km thick tectonostratigraphic sequence. Six major thrust nappes have been identified in this region (Fig. 4.21). According to Treloar et al. (1990) each nappe is internally imbricated by thrusts with higher grade rocks in hanging walls instead of the foot walls, resulting in inverted metamorphic sequences. West to east, the south-verging zones of thrust nappes are arranged in a row and are separated by prominent shear zones and thrust faults. These structural zones are briefly discussed below.

Mohmand–Swat Region

The basement sequence in this region is comprised of Precambrian Manglaur Formation which has been intruded by Swat Granite Gneiss (Kazmi et al. 1984). The Manglaur Formation includes quartz-feldspar schist, quartz-mica-garnet schist and graphitic schist, amphibolite and calcsilicate. The Manglaur schists contain two generations of garnet porphyroblasts indicating at least two period of metamorphism separated by a retrograde episode (Photo. 17). The Manglaur Formation and Swat Granite Gneiss are un-conformably covered by Late Paleozoic Alpurai Group (Kazmi et al. 1984, Pogue et al. 1992, DiPietro 1993). In its lower part garnet-quartz-mica schists (Photos. 18 to 20), quartzofeldspathic gneisses, metapelites and amphibolites dominate whereas the upper part is comprised of calcschists and marbles (see Chapter 5).

The Swat Granite crops out in a number of separate inliers which are largely in the form of antiforms. These antiforms are separated by synclinal troughs filled with cover sediments (Alpurai Group). North of Nikanai Ghar, the Ilum Granite forms a south verging nappe zone (Fig. 4.33). In the Malakand–Mohmand area, large (several kilometre long) slabs of ophiolites and ophiolitic melanges derived from the Indus Suture zone form southwest verging klippe and nappes. The general geology and structure of the Mohmand–Swat region is shown in Figures 4.21 and 4.33.

Besham Nappe

Eastward the Mohmand–Swat sequence is truncated by the north-south trending Puran Fault which, according to Baig et al. (1989) and DiPietro et al. (1993), is an oblique, high-angle, left-lateral shear zone. However, Treloar et al. (1989c,d) name this fault as the Alpurai Thrust and consider it as a south vergent thrust, folded by the Besham antiform. Along this fault, the Swat sequence overlies Late Archean to Late Proterozoic sequence of the Besham Nappe. This nappe zone comprises a N-S oriented antiform at the apex of the Indus Syntaxis (Calkins et al. 1975, Baig et al. 1989). The core of the antiform is comprised of quartzo-feldspathic basement gneisses, granites, metasediments and metavolcanics (komatiites) of the Besham Group (Tahirkheli 1979). These are imbricated with metasediments of younger cover sequence of the Karora Group. The cover rocks in the core of the anticline have been metamorphosed to chlorite grade whereas

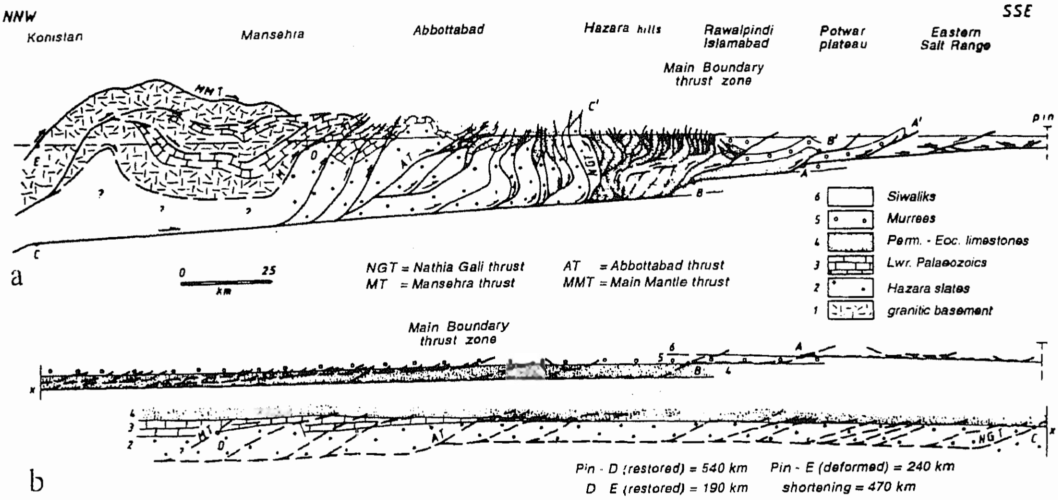


Fig. 4.34. Simplified balanced (a) and restored (b) cross-section through Pakistan Himalayas, from Main Mantle Thrust (MMT) to foreland and footwall cutoffs as close as is compatible with adjacent geology. Structure south of Main Boundary zone based on commercial seismic data, and to north on reconnaissance traverses. Termination of thrust traces in northern part of section do not imply tips. (From Coward et al. 1985).

the tectonised cover slices towards the flanks are comprised of higher grade garnetiferous rocks. This feature gives an overall sense of metamorphic inversion (see Chapter 7).

^{40}Ar/^{39}Ar dating of amphibolites from the Besham Group records an 1,850 Ma mid-Proterozoic thermal event (Treloar and Rex 1990) and confirms its Late Archean to Early Proterozoic age. ^{40}Ar/^{39}Ar ages indicate metamorphic and deformational events at >2,000 ± 6 Ma, >1,950 ± 3 Ma and magmatism at about 1,500 and 550 Ma (Baig et al. 1989, Treloar et al. 1989b).

The Karora Group (Ashraf et al. 1980, Baig et al. 1989) is composed of graphitic and psammitic phyllite, calcpelite, quartzite and dolomite and it commonly overlies the Besham Group along north trending shear zones. However, 7 km west of Besham, the Karora Group sediments have an undisturbed contact and unconformably overlie the Besham Group with a 5 m thick basal conglomerate (Amlo Conglomerate). At this locality the Karora Group is characterised by lower greenschist facies, whereas the underlying Besham Group has been multiply metamorphosed to epidote amphibolite and upper amphibolite facies. Furthermore, the Amlo Conglomerate contains gneissic clasts of underlying Besham Group and also clasts of mafic and felsic rocks that intrude the Besham Group. These features suggest that the Besham Group is characterised by at least two phases of Precambrian deformation and metamorphism prior to the deposition of the Karora Group (Baig et al. 1987, Treloar et al. 1989a). According to Baig et al. (1989), the Karora cover sequence may be correlated with Manki and Hazara Formations and was probably deposited between Early Middle Proterozoic (<1,517 Ma) to Early Late Proterozoic (728–950 Ma).

North of Besham, the Kohistan island arc sequence (Jijal Complex) has been thrust over the Besham sequence along the MMT (Figs. 4.21, 6.2 and Photo. 21). The thrust zone is characterised by breccia with serpentinite matrix and blastomylonites derived

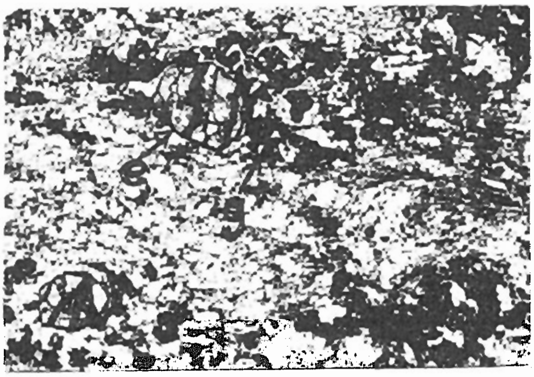


Photo 17.

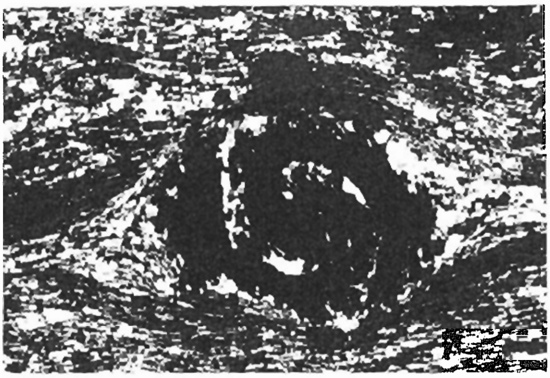


Photo. 18.

Photo. 19.

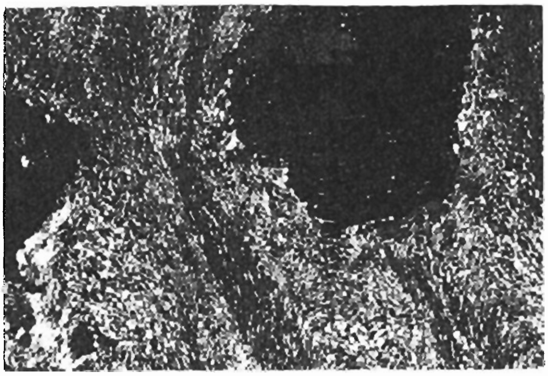


Photo. 20.

Photo. 21.

Photo 22.

Photo. 17. Microphotograph showing garnet of two generations in Manglaur crystalline schist; fresh euhedral crystals (e) growing across earlier crushed granoblasts (g). (Photo. *A.H. Kazmi*).

Photo. 19. Microphotograph showing garnet porphyroblast with curving inclusion trails from the Alpurai Schist, near Alpurai. Garnet growth was metamorphism and synchronous with deformation. (Photo. *P. J. Treloar*).

Photo. 21. Folded mylonites (z-fold) on the footwall of the MMT, south of Jijal. (Photo. *A.H. Kazmi*).

Photo. 18. Microphotograph showing rotated garnet porphyroblast in Alpurai Schist, near Alpurai. Garnet growth was synchronous with shearing deformation, S, fabric. (Photo. *P. J. Treloar*).

Photo. 20. Microphotograph of Alpurai Schist. Garnet crystal, growth of which postdates an early fabric (as shown by straight quartz inclusion trails) but predates development of an S2 crenulation cleavage. The crenulation is early in the deformation history, thus garnet growth was early in the deformation. (Photo. *P. J. Treloar*).

Photo. 22. Mylonitic, sillimanite grade Tanawal Formation sediments in the Thakot Shear Zone, 6 km south of Thakot. (Photo. *P. J. Treloar*).

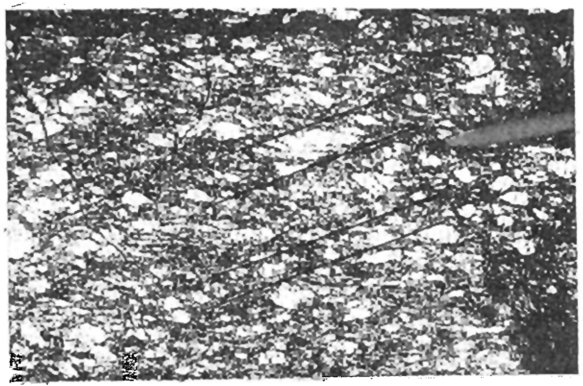


Photo. 23

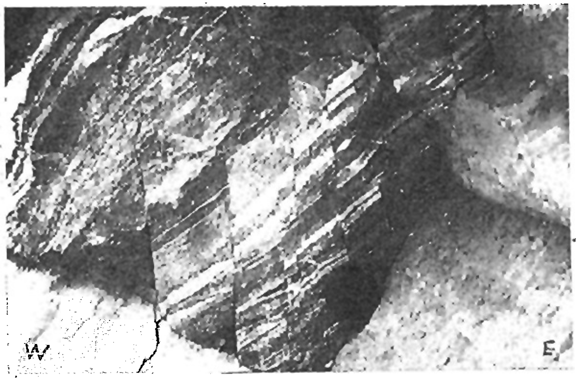


Photo. 24

Photo. 23. Photograph showing early ductile shear bands along Sassi Shear Zone in Nanga Parbat Gneiss near Sassi. The shear bands (marked) are east-side up. (Photo. *P. J. Treloar)*

Photo. 24. Late stage brittle faults in the Nanga Parbat gnesiss near Sassi, showing east-side-up displacement. This displacement of the Nanga Parbat Gneiss was accomodated within the Sassi shear zone by ductile through to brittle thrust type displacement along what are now steeply dipping structures (Photo *P. J. Treloar*).

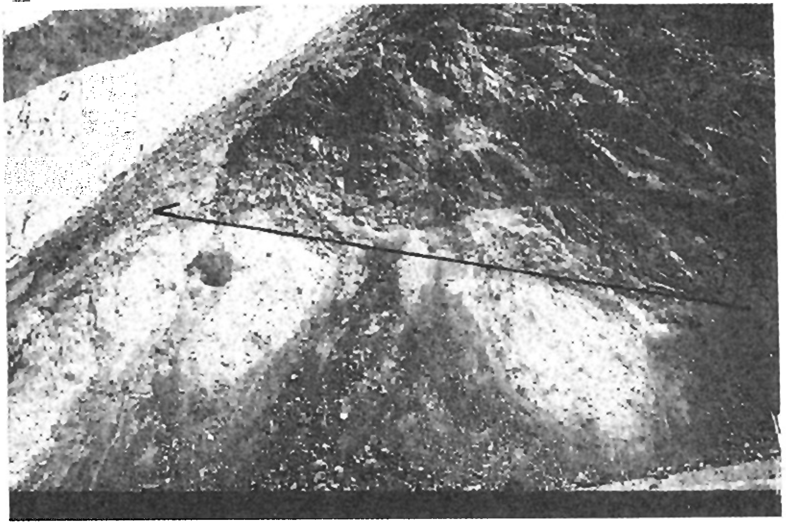


Photo. 25. Photograph showing the Liachar Thrust. Along the western margin of the Nanga Parbat Massif, this thrust places high grade Nanga Parbat Gneisses on its hanging wall onto Recent Indus Valley glacial sediments (< 10,000 years old). Position of thrust shown by black line. (Photo. *P. J. Treloar*).

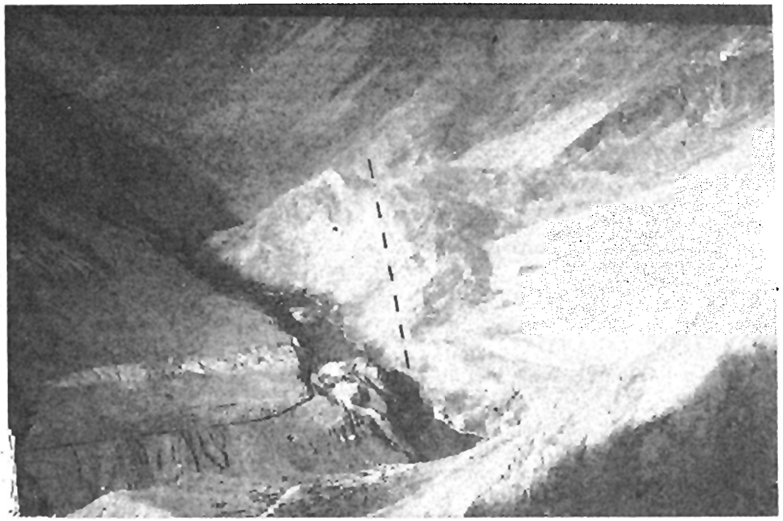


Photo. 26. Photograph showing the Liacher Thrust and the remains of the 1894 landslide (preserved in the hammocky ground on east bank of the Indus. Within the mound, the Liacher Thrust (marked by a dashed line), separates the Recent glacial sediments on the left from gneisses on the right. (Photo. *P. J. Treloar*).

from paragneisses. Eastward the Besham sequence has been cut off by steeply dipping, dextral strike-slip, mylonitic Thakot shear zone which also shows a late component of east-side-down semi-brittle movement (Lawrence and Ghauri 1983). Its surface trace gives the impression of dextral strike-slip motion, though according to Treloar et al. (1989c,d) it is a folded thrust fault. The Thakot Fault juxtaposes the Besham Nappe sequence against the Hazara Nappe. The Besham rocks are largely biotite grade with tectonic garnetiferous pods whereas the adjacent Hazara Nappe sequence is sillimanite grade (Photo. 22). Thus apart from the older age of the Besham sequence, there is also a metamorphic discontinuity across the Thakot Fault. In the southern part the Besham Nappe sequence is truncated by the Darband Fault and pinches off north of Darband. Only a narrow outcrop of the cover sequence continues southward along the western edge of the Tarbela Lake. Treloar et al (1989c,d, 1991a) suggest that the Besham Nappe sequence underlies the Hazara Nappe and Swat Nappe sequences and that the Besham sequence is apparently a tectonic window.

Hazara Nappe

This nappe is 40 to 50 km wide and the Mansehra Thrust forms its southern boundary. It extends northwards for about 75 km, up to the MMT and the Banna Thrust (Fig. 6.2). The Balakot shear zone separates it from Hazara–Kashmir Syntaxis and the Kaghan Nappe. The basement rocks in Hazara Nappe are comprised of metaquartzites and garnet-mica schists of the Precambrian Tanawal Formation which is intruded by thick sheets of porphyritic, two-mica Mansehra Granite (516 m.y. Rb/Sr isochron age, Le Fort et al. 1980). The Cambrian Abbottabad Group unconformably overlies the Tanawal Formation and forms the cover sequence (Latif 1974).

One characteristic structural feature of the Hazara Nappe is the presence of a number of north-dipping, south-verging, curved thrust faults or shear zones which terminate northwards in the Thakot and Batgram shear zones (Fig. 4.21). The latter shear zones are apparently crustal scale structures like a pair of lateral ramps to a thrust system. The other thrusts and shear zones mentioned earlier, and which internally imbricate the Hazara Nappe, are apparently hanging wall splays of the linked lateral ramp system. The Hazara Nappe is further characterised by an inverted Barrovian-type of metamorphism. This is due to the parallelism between isogrades and thrusts. The metamorphic grade changes across the thrusts in a manner that while it remains constant within any one thrust slice, it

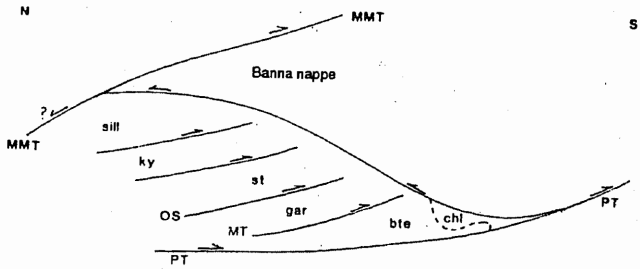


Fig. 4.35. North-south section through the Hazara nappe showing the internal imbrication within the nappe and its structural relationships within the underlying Besham nappe and overlying low grade rocks of the Banna nappe. The Panjal Thrust forms the sole thrust for the two internal zone deformations. Upward the increase in metamorphic grade is the result of post-metamorphic stacking and disruption of the metamorphic pile. (From Treloar 1989).

differs from the adjacent slice. Each major N-dipping thrust has higher-grade rocks in the hanging-wall than in the foot-wall. The metamorphic grade increases from biotite grade rocks south of Mansehra Thrust, to kyanite and sillimanite grade near the MMT (Fig. 4.35 and Photo. 22).

Banna Nappe

The Banna Nappe lies to the south of MMT and crops out in Alai Kohistan, along the northern margin of the Mansehra Granite. It comprises calcareous schists, slates, phyllites and marbles. The Banna Thrust juxtaposes this group over the high-grade Tanawal Formation of the Hazara Nappe (Fig. 4.35). This thrust, however, has a backthrust sense of movement towards the northwest and it is characterised by brittle deformation (Treloar et al. 1989a,d).

Kaghan Nappe

The Kaghan Nappe forms the northern portion of Hazara-Kashmir Syntaxis. It lies between Panjal Thrust and the MMT. Westward, the Mansehra Thrust–Balakot shear zone separate it from the Hazara Nappe (Figs. 4.21 and 6.2). The Balakot shear zone forms a north verging loop, following the general trend of Hazara–Kashmir Syntaxis. According to Greco (1989) and Greco and Spencer (1993), it swings southeastward to join the MCT. South of Naran, this shear zone has a southeasterly trend and has been named Batal Thrust by Chaudhry et al. (1986) and Ghazanfar et al. (1986). This fault divides the Kaghan region into a northern Sharda zone and a southern Kaghan zone. Treloar et al. (1989c,d) suggest that these are two separate nappes.

The Kaghan zone is comprised of unfossiliferous low grade metamorphic rocks, mainly pelitic schists, quartzite and mafic dykes. According to most workers, this sequence belongs to the Salkhala Formation (Wadia 1931, Calkins et al. 1975, Yeats et al. 1984, Greco et al. 1993). However Ghazanfar et al. (1985) and Chaudhry et al. (1986) have described this sequence as the Kaghan Group (see Chapters 5 and 7).

The Sharda zone is composed of basement paragneisses (Precambrian ?) intruded by porphyroblastic, two mica granite gneisses (c.f., Mansehra Granite) and a cover sequence of "carbonitic and pelitic" rocks with interlayered basaltic flows (Greco and Spencer 1993). This sequence is characterised by recumbent folding, ductile thrusting and several nappes. Chaudhry et al. (1986) refer this basement cover sequence as the Sharda Group and they have mapped and described a number of sub-units of this group (Fig. 4.36). Greco and Spencer (1993) on the other hand divide the sequence north of the Batal Fault into (a) Cambrian and older granitic basement, (b) Lower Paleozoic cover (Naran Formation of Greco et al. 1989), and (c) Paleozoic-Mesozoic metasedimentary sequence (Burawai Formation of Greco et al. 1989). We have addressed these stratigraphic problems in Chapter 5.

The granite gneisses of the Sharda Group have been described as mainly pre-tectonic (pre-Himalayan), S-type bodies folded along with the Sharda Group paragneisses and amphibolites. They are thought to be derived from the Sharda Group paragneisses through earlier anatexis. They have been grouped with the High Himalayan leucogranites (Chaudhry and Ghazanfar 1987). The Batal Fault, which has been referred to as MCT by Ghazanfar and Chaudhry (1986) and Greco (1991), separates the low metamorphic grade (greenschist facies) rocks of the Kaghan Group from high grade (upper amphibolite facies) sequence of the Sharda Group. Thus in Kaghan region also there is an inversion of metamorphic grade which increases upwards with successive nappe.

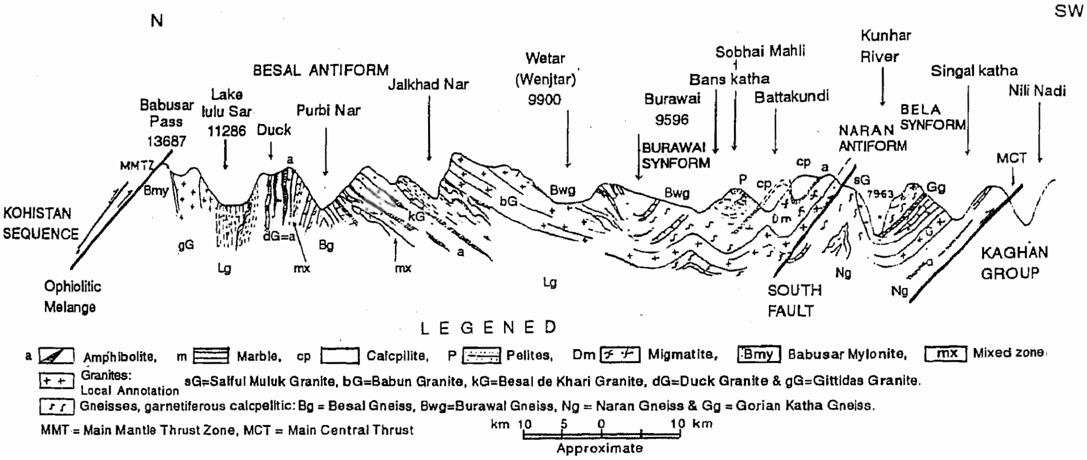


Fig. 4.36. Generalised geological cross-section through upper Kaghan, parallel to Kunhar River. (From Chaudhry and Ghazanfar 1987).

Nanga Parbat–Haramosh Massif

The high-grade metamorphic and magmatic rocks of the Kaghan region apparently continue northeastward into the Nanga Parbat Range. A large part of this range, east of the Babusar Pass, has not been mapped. The connection between the Kaghan and the Nanga Parbat–Haramosh sequence is uncertain. Wadia (1933), Misch (1935), Gansser (1964) and other workers had placed the entire metasedimentary sequence of Hazara–Kashmir Syntaxis, Kaghan and Nanga Parbat region in the Salkhala Formation. Geological mapping of Nanga Parbat–Haramosh region by Madin (1986) and Verplanck (1986), Coward (1986), Butler et al. (1988) and Lawrence et al. (1989) has shown that the metamorphic stratigraphy and possibly the age of Nanga Parbat–Haramosh sequence is distinctly different from the Salkhala Formation (see Chapters 5, 6 and 7).

The Nanga Parbat–Haramosh Massif is a major re-entrant into the Kohistan arc terrain and is commonly known as the Nanga Parbat Syntaxis (Photo. 10). It is comprised of more than 15 km thick Proterozoic gneisses and schists of the Nanga Parbat Group (Madin 1986). The lower part of the section contains amphibolite grade pelitic and psammitic Shingus Gneiss which is a paragneiss. Shingus Gneiss is interlayered by 500 Ma orthogneisses and it is tectonically overlain by the Iskere Orthogneiss (1850 Ma zircon date, Zeitler et al. 1989). The amphibolites and marbles form upper part of the sequence (see Chapter 5, 6 and 7). The contact between Iskere and Shingus Gneiss is sheared and is marked by sillimanite bearing mylonites. Treloar et al. (1991b) view the Shingus Gneiss as a cover to Iskere Gneiss and correlate it with Tanawal Formation–Mansehra Granite sequence of Hazara region. In the Haramosh region, the Iskere Gneiss is overlain by the Haramosh Schist which comprises medium- to coarse-grained, amphibolite-grade biotite schist and gneisses, marble, calc-silicate gneiss and subordinate amphibolite. Southward, the Baroluma Fault separates the Haramosh Schist from the Shingus Gneiss. At the same location where Madin et al. (1989) have mapped and described the Haramosh Schist, Butler et al. (1992) have described a sequence of interlayered ortho- and paragneisses with abundant sheets of deformed granitoids. They call this sequence "Layered Unit" and have shown that it overlies the Iskere Gneiss and crops out around the northern plunging end of the Nanga Parbat antiform (Fig. 6.1). We propose that this unit be named Haramosh Formation. This formation underlies the Kohistan metavolcanics and the two are juxtaposed along the MMT which is clearly visible along the southern aspects of the lofty watershed bearing the Darchan (5,500 m) and Haramosh II Peaks (6,666 m). The Nanga Parbat Group is intruded by younger, undeformed granites, pegmatites and mafic dykes (see Chapter 6).

The Nanga Parbat-Haramosh Massif comprises two large, upright and north-trending folds, the Iskere and Bulechi antiforms (Fig. 4.37). The massif has been multiply deformed and three folding events may be recognised. The earliest folds are tight isoclinal structures and were synchronous with peak metamorphism. The structures are ductile and formed on the western side, along the Raikot–Sassi Fault zone and pre-date 9 Ma. The second generation folds are E-W large kink folds and the later set of folds are the very large N-S Iskere and Bulechi anticlines (Madin et al. 1989, Lawrence et al. 1989). These larger folds formed around 4–5 Ma. According to Treloar et al. (1991b), the Nanga Parbat Syntaxis apparently post dates south-verging crustal stacking and thickening within the core of the syntaxis.

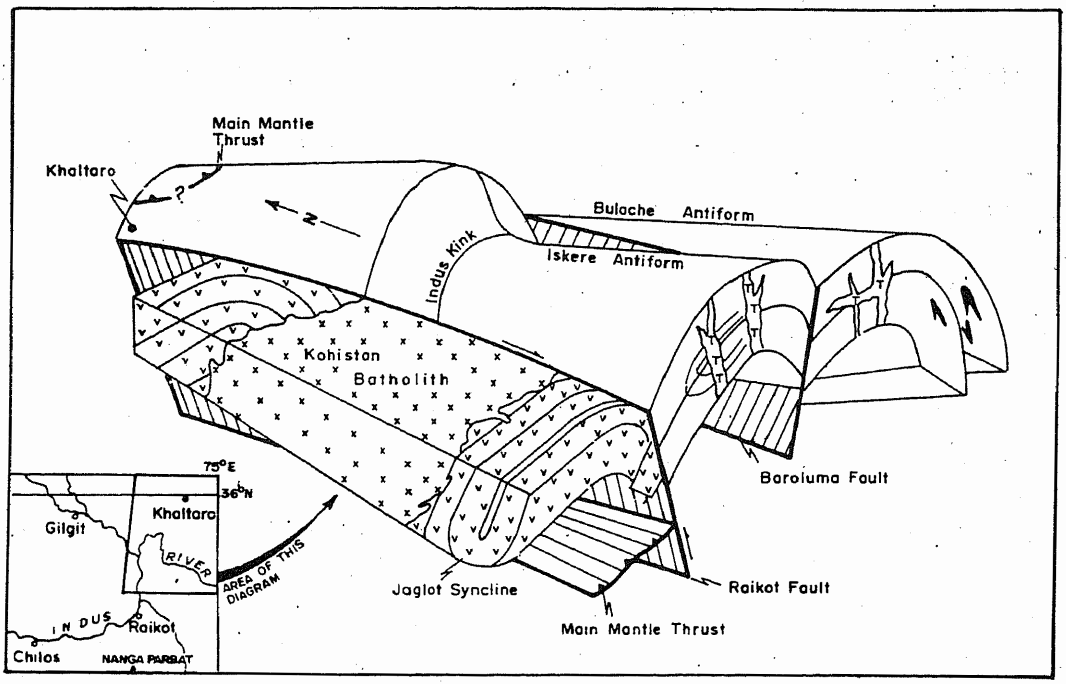


Fig. 4.37. Structural diagram of the northwestern portion of the Nanga Parbat-Haramosh Massif. In the Kohistan terrain west of the Raikot Fault "V" symbol indicates metavolcanic and metasedimentary rocks of the andesitic arc, "x" symbol indicates granodiorite and quartz diorite plutons of the Kohistan batholith. The eastern end of the Jaglot syncline shows drag by oblique right-lateral reverse motion of the Raikot Fault. In the Nanga Parbat-Haramosh terrain east of the Raikot Fault, the Nanga Parbat granite gneisses are shown without pattern, isoclinally folded and transposed amphibolite dikes are shown in solid block, and post-tectonic, cross- cutting tourmaline granite dikes and sills are indicated by "T" symbols. (From Madin 1988, Lawrence et al. 1989).

The Nanga Parbat-Haramosh Massif is bound by faults to the east, north and west. According to most investigators (Tahirkheli 1979, Bard et al. 1980, Coward et al. 1986, Butler et al. 1992), the MMT surrounds this massif. Madin (1986) and Madin et al. (1989) presented field evidence to show that along the western side of Nanga Parbat–Haramosh Massif an active dextral reverse fault, the Raikot Fault, truncates MMT and forms the western boundary of the massif. Treloar et al. (1991b) have, however, shown that between Raikot and Sassi, the MMT is comprised of a complex system of faults, thrusts and shears (Photos. 23 and 24). In the southern part it is a brittle thrust, the Liachar Thrust, that places the Nanga Parbat Gneiss on Recent Indus alluvium (Fig. 4.38, Photos. 25 and 26). South of Bunji, there are a number of smaller fault splays parallel to MMT. The Bunji–Sassi section of the fault, the Shahbatot Fault zone (Butler et al. 1989), contains outcrops of marbles and mica schist within the fault zone. Treloar et al. (1991b) contend that the MMT is present between the Raikot–Sassi zone though extensively deformed by later shears.

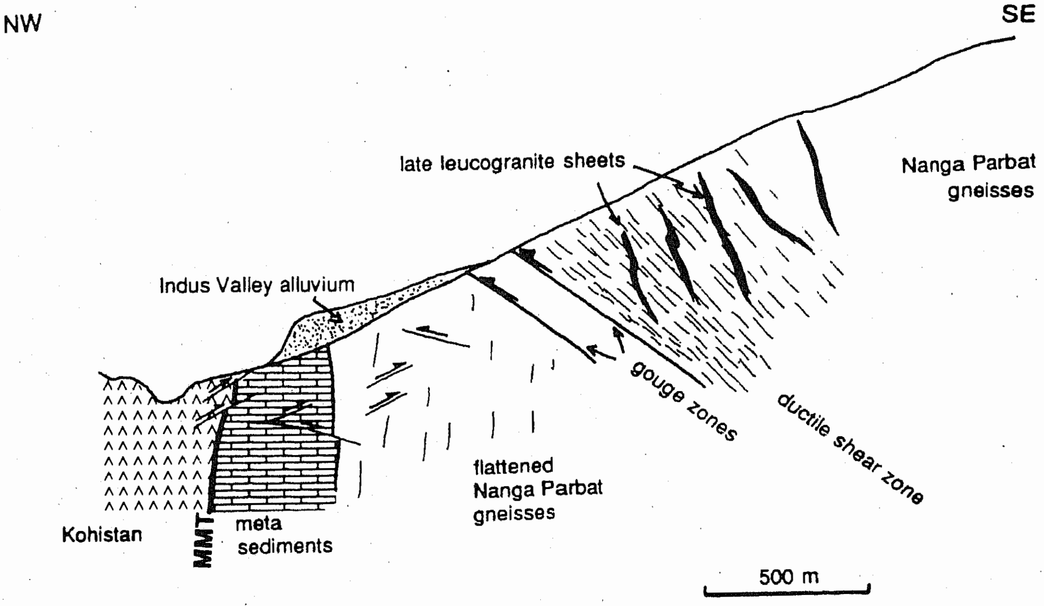


Fig. 4.38. Section across the Liachar thrust zone showing the northwest verging nature of the thrust with Indian Plate gneisses in its hanging wall and the vertical MMT sequence, unconformably overlain by recent Indus Valley alluvium, in the foot-wall. MMT= Main Mantle Thrust. (From Treloar et al. 1991).

Along the northeastern margin of Nanga Parbat–Haramosh Massif, the Nanga Parbat Gneisses are separated from the Ladakh arc terrane by a 3–5 km wide fault zone, the Stak Fault zone. Within this zone there are four major faults (Verplank 1986). West to east these are the Majupah, Ganji, Stak Chi and Askore Faults. The Majupah Fault separates Shingus Gneiss from gabbro, ultramafic pods and amphibolites of island arc affinity. A 20–30 m wide mylonite zone, which is truncated by a 5–8 m wide breccia, marks the Majupah Fault. Eastward the Ganji and Stak Chi Fault zones comprise banded gneisses and schistose rocks, containing tectonically emplaced, pods of ultramafic rocks. The Askore Fault is marked by a 50 m wide zone of chlorite-biotite schist with small inclusions of ultramafic rocks. It separates banded gneisses and amphibolites in the west from diorites in the east. Movement along the Stak Fault zone is apparently right lateral and west side up.

The Nanga Parbat–Haramosh Massif is unique due to its current extraordinarily high uplift rate of over 5 mm/year. Fission-track and $^{40}Ar/^{39}Ar$ data by Zeitler (1985) show that this uplift is 7 times higher than in the adjacent Kohistan arc terrain. Holocene movement and dislocation of Quaternary deposits along Raikot Fault (Lawrence and Ghauri 1983, Shroder et al. 1989) suggest that this differential uplift of the two blocks has been accomplished along the MMT and the complex Stak and Raikot–Sassi Fault zones. Zeitler's data (1985) show that this massif has been uplifted approximately 16 km in the last 6 to 8 m.y. and that the uplift has accelerated exponentially. On the basis of reconstruction of the original stratigraphy and subsequent postulated erosion, Madin et al (1989) estimate that the actual total uplift of Nanga Parbat may have been about 24 km.

We have earlier mentioned that the Hazara–Kashmir Syntaxis apparently originated due to anticlockwise rotation of thrust directions in combination with a sinistral shear zone that developed along the western margin of the syntaxis. By contrast, the Nanga Parbat Syntaxis formed above the pinned lateral termination of a major Himalayan thrust which created strong clock-wise rotation coupled with thrusting and strike-slip faulting (Coward et al. 1986, Butler e al. 1989, Madin et al. 1989)

Chronology of Significant Events

Radiometric data, particularly Ar-Ar and K-Ar geochronology of amphiboles and micas together with fission track zircon and apatite data, clearly indicate that the zone of crystalline nappes has a long history of multiple pre-Himalayan as well as Himalayan deformation, metamorphism and magmatism (Table 4.6). It is comprised of six nappes as described earlier. Despite general similarities there are significant differences in their deformational, metamorphic and magmatic histories, which suggest that they had initially lain at separate locations and have been juxtaposed only during relatively late Himalayan tectonic events. Each block consists of a Proterozoic crystalline basement of schists and gneisses intruded by Cambrian granites and unconformably (or tectonically) overlain by a metasedimentary sequence of cover rocks varying in age from Precambrian to Jurassic. The Besham basement rocks are the oldest (Late Archean to Early Proterozoic) and have been deformed by three separate Early Proterozoic orogenic episodes (Table 4.6). These events were accompanied by amphibolite facies metamorphism, granitic and mafic dyke intrusions, and were followed by another magmatic event during Early Middle Proterozoic (Baig et al. 1989). The Besham cover sequence (Karora Group) may also be the earliest cover and has been tentatively shown as Middle Proterozoic by Baig (1990). Unlike basement rocks in other blocks, the Besham basement gneisses do not bear Himalayan-age thermal overprint, though they are imbricated with a cover sequence with Himalayan metamorphic imprints. According to Treloar (1991a,c) the two may have been deformed and metamorphosed separately, prior to re-imbrication during a later Himalayan event.

The Nanga Parbat sequence differs from other blocks inasmuch as it comprises three

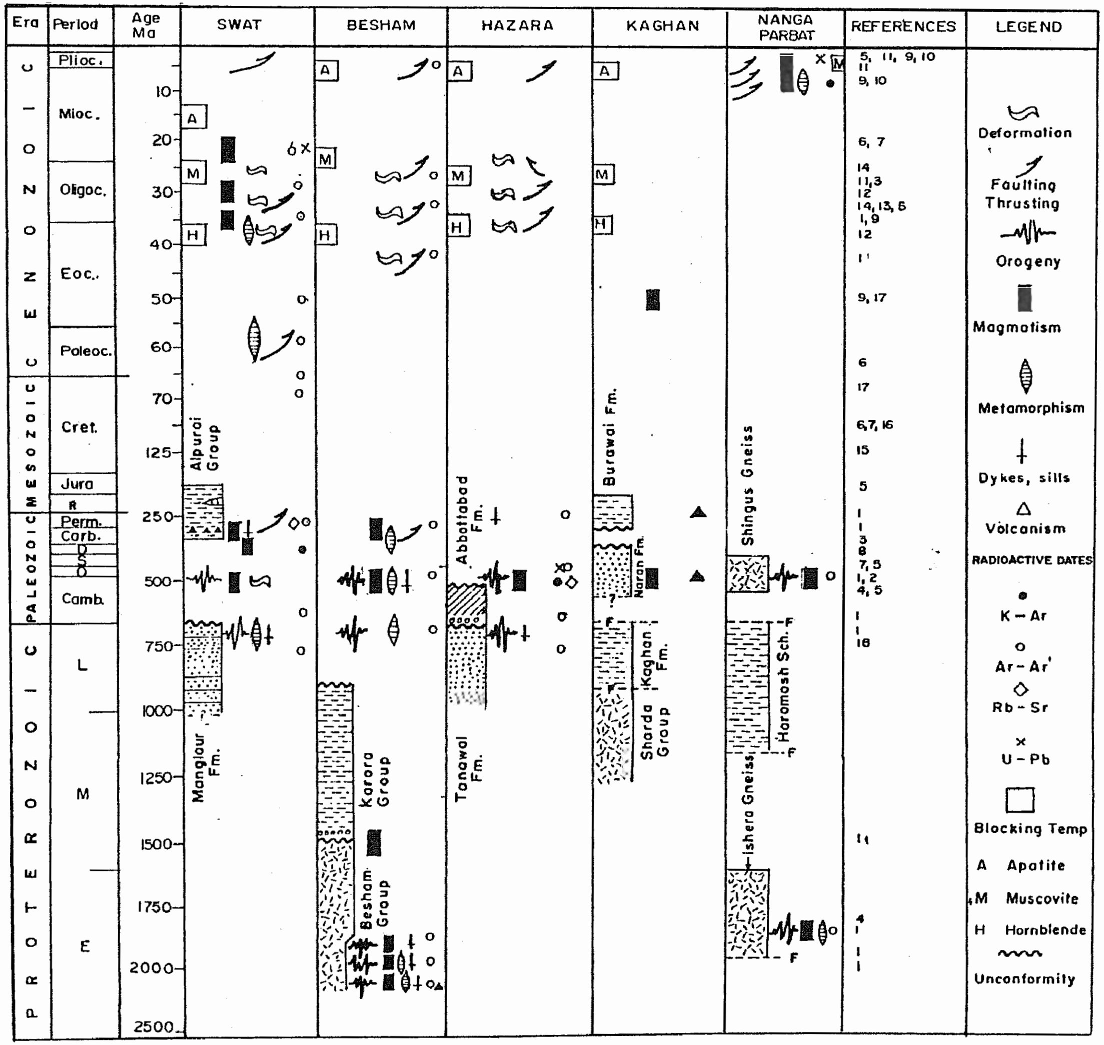


Table 4.6. Chronology of significant events in the zone of Crystalline Nappes. Compiled from: 1. Baig et al. 1991; 2. Le Fort et al. 1980; 3. Le Bas et al. 1987; 4. Zeitler et al. 1989; 5. Treloar 1989; 6. Maluski et al. 1984; 7. Zeitler et al. 1982, 1985; 8. Kemp 1986; 9. Zeitler et al. 1991, 1993; 10. Smith et al. 1990, 1992; 11. Treloar et al. 1991; 12. Rosenberg 1985; 13. Lawrence et al. 1989; 14. Treloar et al. 1990; 15. Maluski et al. 1982; 16. Shams 1980; Treloar et al. 1990; 18. Crawford 1975; 19. Greco et al. 1993; 20. Chaudhry et al. 1987.

lithostratigraphic units with fault contacts and lack of clarity regarding their actual superposition. According to Madin et al. (1989) Shingus Paragneisses are structurally the lowest, followed by the Iskere Orthogneisses and Haramosh Schist. Treloar et al. (1991b) consider the Shingus Gneiss to be the cover to Iskere Gneiss. The latter provide 1,850 Ma zircon date (Zeitler et al. 1989) which is interpreted as a deformational, thermal and magmatic event (Treloar et al. 1990b). This event was apparently wide-spread as indicated by Ar/Ar 1887-1865 Ma dates from Besham (Baig 1990) and 1820 ± 130 Rb-Sr date from Kumaon (Valdiya 1988).

The Swat–Besham–Hazara region has been affected by Late Proterozoic Hazaran Orogeny (Baig et al. 1987, 1989) with evidence of metamorphism, volcanism and mafic magmatism. Based on Ar/Ar dating Baig (1991) mentions volcanism and plutonism associated with this orogeny at 850 to 600 Ma and metamorphism and deformation at 664–625 Ma. These dates compare favourably with Rb/Sr isochron age of 728 ± 20 Ma and 889 ± 40 (K-Ar) for Hazara Formation (Crawford and Davies 1975), 865 ± 20 to 809 ± 20 Ma for the Kirana volcanics (Davies and Crawford 1971, Ahmad et.al. 1997), 721 ± 12 Ma K/Ar from Nepal Higher Himalayan meta-sediments, 819 ± 80 Ma from metasediments and volcanics in Nepal (Krummenacher 1961, 1966) and several Rb/Sr isochron ages of 719 ± 10 to 743 ± 10 from Rajasthan granites and volcanics (Crawford and Compston 1970). These dates indicate that the Hazaran Orogeny is a widespread feature which may be correlated with Pan-African and Baikalian Orogenies in Africa and Asia.

Another significant tectonomagmatic event is indicated by the Cambrian-Ordovician granitic intrusions in Swat, Hazara, Besham, Kaghan and Nanga Parbat. This event included mafic dykes and volcanism (Table 4.6). The last pre-Himalayan deformation affecting these regions occurred during the Permo-Carboniferous with extensive volcanism (Panjal Volcanics), mafic dykes and granite intrusions (see Chapter 6).

Inasmuch as the Himalayan Orogeny is the result of India-Asia collision, the earliest Himalayan tectonometamorphic events are constrained by the time of the collision. The blueschist and greenschist facies melanges, which have been thrust southward over crystalline nappe zone, have provided Ar-Ar and K-Ar dates of 67 ± 12 to 100 ± 20 Ma (Maluski et al. 1984, Shams 1980, Treloar et al. 1990) for metamorphism. This is probably related to precollision subduction of the Indian plate margin. From Besham and Swat region, amphibolite dykes have given 67, 65 and 50 Ma ages. The 67 to 65 Ma dates have been interpreted as a thermal event indicating ophiolite obduction which may have been accompanied by collapse of passive continental margin and the 50 Ma thermal event suggests collision of the Indo-Pakistan plate with Kohistan (Treloar et al. 1990). The initial collision may have been earlier as suggested by the angular unconformity along which deformed Late Cretaceous sequence is overlain by undeformed Paleocene carbonates in the Attock–Cherat and Kalachitta Ranges (Yeats and Hussain 1987). Several other geological observations also provide additional support for an early India-Asia contact:

(a) Biostratigraphic studies indicate that the accretionary-prism and trench strata of the Katawaz Flysch Basin were first thrust over the Indo-Pakistan passive plate-margin between 66 m.y. to 55.5 m.y. (Beck et al. 1995).

(b) Indo-Asia faunal exchange at the Cretaceous-Tertiary boundary vide evidence from the Intertrappean Beds of south India (Jaeger et al. 1989).

(c) Pre-Paleocene south-vergent thrusting of southern Tibet over northern India (Searle 1988).

(d) Regional unconformities prior to Cretaceous-Tertiary boundary in the Zanskar region (Gaetani and Garzanti 1991).
(e) Collision-related crustal thickening in the Northwestern Himalaya that started at least 20 m.y. before closure of hornblende isotopic system at about 40–35 Ma (Treloar et al. 1991c) which suggests a pre-60 Ma India-Asia contact.
(f) Extrusion of Deccan Traps (Klootwijk et al. 1991, White and McKenzie 1989).
(g) Increase in oceanic $^{87}Sr/^{86}Sr$ ratios during Late Maestrichtian which may be due to an increased riverine input due to India-Asia contact, or related to a pre-cursor of Deccan Trap volcanism (Martin and MacDougall 1991, Javoy and Courtillot 1989).
(h) Significant plate reorganisation in western India Ocean at about 65 Ma (Courtillot et al. 1986).
(i) Change in the Indian apparent polar-wandering path around 62 Ma (Klootwijk et al. 1985).
(j) Isotopic evidence that suturing in Ladakh was complete before 60 Ma (Klootwijk et al. 1985).

According to Treloar et al. (1991a,c) decoupling of the cover and basement sequence and crustal thickening may have occurred prior to 60 Ma. This was followed by metamorphism which spanned the period 50–30 Ma (Maluski and Matte 1984). According to Treloar et al. (1989c, 1991c) peak metamorphism occurred prior to 35–40 Ma which is the age of cooling through hornblende blocking temperature. Steady cooling sharply increased between 25–18 Ma due to accelerated uplift and a fast rate of exhumation which was probably in excess of 10 km (Figs. 7.10 and 7.16). The cooling and uplift rates in various blocks of the crystalline nappe zone vary, with the Nanga Parbat block showing an extremely rapid uplift and cooling during the past 9 m.y.

Deformation and metamorphic chronologies of this region have been worked out by various workers. Deformational phases identified by them overlap. Whereas Palmer-Rosenberg (1985) and DiPietro (1991) mention three deformational phases, Treloar (1989a, d) suggests five, though he has split (D_2) into two phases. To avoid confusion we have followed Treloar's terminology. In the crystalline nappe zone, the early deformation has been largely ductile during Himalayan metamorphism. Deformed porphyroblasts in the rock fabric, particularly spectacular spiralled garnets (Photos. 18-20),, indicate that metamorphism was synchronous with ductile shear associated with early deformation due to the thrusting of Kohistan over the zone of crystalline nappes (Indian plate margin). This event (D_1) produced simple shear zones characterised by ductile blastomylonites and thrust related folds in MMT, Thakot, Balakot, Oghi, Alpurai and Batal shear zones. These earliest folds are recumbent, isoclinal nappes formed at or near the metamorphic peak constrained by the $^{40}Ar/^{39}Ar$ hornblende date of 39 to 40 Ma (Lawrence et al. 1985, Treloar et al. 1991c).

In the Besham region the onset of this event may have been earlier as indicated by the $^{40}Ar/^{39}Ar$ dates of 51–36 Ma from shear zones (Baig 1991). This early phase of Himalayan deformation was probably accompanied by uplift and formation of a mountain front, perhaps of low relief which was quickly eroded away prior to 35 Ma and provided molasse type sediments to the foreland basin (Kazmi and Rana 1982, Treloar et al. 1991c). Stratigraphic evidence for this early Himalayan episode during Late Eocene is provided by the interfingering of Eocene marine and continental sediments, presence of MORB type volcanic clasts in Eocene red beds in the Hazara–Kashmir Syntaxis (Bossart et al. 1989), and serpentinite boulders in Late Eocene Kuldana Formation near Murree (Treloar et al. 1990).

The second phase of deformation (D_2) was also synmetamorphic and occurred shortly after D_1 (Lawrence et al. 1989); it may have been a diachronous continuation of D_1 (Treloar et al. 1989c). The second deformation formed upright, N-S open folds, 12–18 km wide, and 3–5 km high in Swat region (Lawrence et al. 1989). They are characterised by crenulations and are sheath-like features in the northern part. Palmer-Rosenberg (1985), on the basis of muscovite dates, places this event between 37–30 Ma. Data by Treloar et al. (1990, 1991c) suggests D_2 event may predate 34 to 24 Ma muscovite blocking temperature dates (Figs. 7.16 and 7.17).

The third major deformation (D_{2a}) involved restacking and reimbrication of the metamorphic pile involving crystalline basement and the cover sequence and formation of late shears. This phase of thrust stacking apparently covered the 34–24 Ma period of muscovite blocking temperature. Treloar et al. (1990) report that muscovite ages from shear zones and unsheared rocks are similar, thus stacking occurred no later than cooling through the muscovite blocking temperature. This Oligocene deformational phase may be considered as the second major Himalayan orogenic phase (Middle Himalayan Orogeny of Kazmi and Rana 1982; second mountain front of Treloar et al. 1991c). It is also characterised by widespread intrusion of leucogranites (see Chapter 6).

The period 25–18 Ma (Early Miocene) saw rapid cooling through muscovite-apatite blocking temperatures with accelerated uplift, exhumation and deposition of extensive molasse in the foreland basin (Murree Formation). This uplift represents the fourth deformation (D_3) associated with hinterland directed extensional (normal) faults, back folds and backthrusts (Banna Thrust).

The fifth and the latest deformation (D_4) covers the last 9 m.y. and its effects are most dramatically seen in the Nanga Parbat–Haramosh Massif which is characterised by faulting, thrusting and mylonitisation along the Liachar–Shahbatot Faults or MMT, Plio-Pleistocene magmatism, metamorphism and the extraordinary rapid uplift and exhumation. The late east-west folds in Swat area, apparently, also formed during this period.

KOHISTAN MAGMATIC ARC

Kohistan is an intraoceanic island arc bounded by the Indus Suture zone (MMT) to the south and the Shyok Suture zone (Main Karakoram Thrust or MKT) to the north. This E-W oriented arc is wedged between the northern promontory of the Indo-Pakistan crustal plate and the Karakoram block (Photo. 6). Gravity data modelling indicates that the MMT and MKT dip northward at 35° to 50° and that the Kohistan arc terrain is 8 to 10 km thick (Malinconico 1989). Seismological data suggests that the arc is underlain by the Indian crustal plate (Seeber and Armbuster 1979, Fineti et al. 1979). The northern and western part of the arc, along MKT, is covered by a sequence of Late Cretaceous to Paleocene volcanic and sedimentary rocks (Fig. 4.39). This sequence is comprised of several formations which have been described in detail in Chapters 5, 6 and 7. The central part of the arc terrain is mainly composed of Kohistan Batholith which comprises an early (110–85 Ma) suite of gabbro and diorite, followed by more

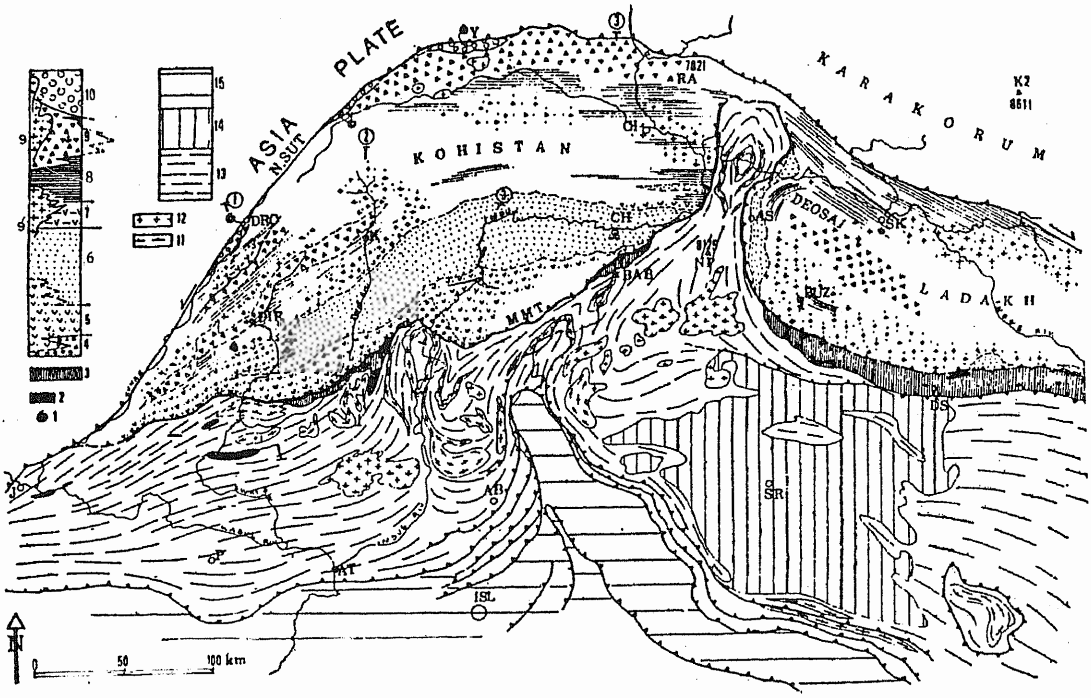


Fig. 4.39. Geological map of Kohistan arc and the Indo-Pakistan plate margin (from Bard et al. 1980). *Kohistan arc sequence:* 1. Cretaceous (Eocene in Dir); 2. Ultramafic klippes; 3. Greenschist belt with local blueschist slices; 4. Jijal ultramafic complex; 5. southern amphibolitic belt; 6. "Pyroxene Granulite" belt; 7. Northern amphibolitic belt; 8. Kalam-like oceanic series; 9. Utror volcanic and volcanosedimentary equivalents; Syn- to post-kinematic dioritic and granodioritic intrusives; 10. Detrital upper series. *Indo-Pakistan Plate:* 11. Orthogneiss (Paleozoic granites ?); 12. Undeformed granitoids; 13. Precambrian and Paleozoic series; 14. Upper Paleozoic to Triassic series of the Srinagar basin including Panjal Traps; 15. Foreland Cenozoic deposits. AB= Abbottabad, AS= Astor, AT= Attock, BAB= Babusar Pass, BUZ= Burzil Pass, DRO= Drosh, Ds= Dras, Gl= Gilgit, ISL= Islamabad, J= Jalalabad, K= Kalam, N. SUT= Northern Suture (MKT), MMT= Main Mantle Thrust (southern suture), NP= Nanga Parbat, P= Peshawar, SR= Srinagar, SK= Skardu.

extensive intrusions of gabbro, diorite and granodiorite (85–40 Ma) which are intruded by much younger dykes and sills of leucogranite (30–26 Ma).

The southern part of Kohistan is comprised of a thick sequence of mafic and ultramafic rocks. These rocks may be divided into three tectonometamorphic complexes separated by major thrust zones. The Chilas Complex forms the northern and upper unit. It comprises layered norites and gabbros metamorphosed to granulite facies. It is characterised by a series of south-verging folds. It has been thrust southwards over the Kamila Amphibolites. The latter consist of amphibolites, meta-gabbro and orthogneisses. This sequence comprises a highly tectonised shear zone. Southward, it is thrust over the Jijal Complex which forms a tectonic wedge between the Kamila Shear zone and the MMT (Fig. 6.24). The Jijal Complex is largely comprised of garnet-pyroxene-granulites and ultramafic rocks (Tahirkheli and Jan 1979, Coward et al. 1986, Khan et al. 1993, Treloar et al. 1990, Miller et al. 1991).

A summary of the tectonic, metamorphic and magmatic events in Kohistan is given in Table 4.7, and details in Chapters 6 and 7.

Table 4.7. Chronology of deformation, metamorphism and magmatism in the Kohistan Island Arc.

Age (Ma)	Event	Reference
9 - 5.	Tectonism. Faulting along Raikot (Liachar - Shahbatot faults-MMT).	(1)
15.	Cessation of movement on MMT. Homogenisation of uplift and cooling rates across MMT.	(2)
20 - 10.	Rapid uplift and cooling.	(3)
20.	Juxtaposition of Kohistan and Indo-Pakistan plate sequence.	(2)
25.	Continued faulting along MMT.	(2)
30 - 26.	Post - collision magmatism. Leucogranite dykes and sills.	(4)
53 - 12.	Initiation and continuation of uplift.	(5)
56 - 37.	Deformation, metamorphism and magmatism.	(6)
60 - 40.	Syncollision magmatism; granodiorite, granite intrusions of Kohistan Batholith.	(9)
55 - 50.	Himalayan collision	(7)
58 - 55.	Volcanism	(8)
75 - 70.	Rapid cooling (hornblende-biotite blocking temps) following magmatism.	(3)
75.	Magmatism; intrusion of mafic dykes.	(9) (4).
85 - 60.	Magmatism; gabbro-diorite suite of Kohistan Batholith.	(10) (4) (12)
85 - 80.	Obduction–Jijal pyroxene-granulites and blueschist melange; deformation, metamorphism.	(5)
102 - 85.	Collision-Kohistan arc with Karakoram block.	(4,8)
102.	Precollision magmatism–Kohistan Batholith.	(8) (4).
104.	Formation of arc sequence–volcanics, mafic and ultramafic complexes.	(11)

(1) Zeitler et al. 1991, (2) Zeitler et al. 1982, (3) Zeitler et al. 1985, (4) Petterson and Windley 1985, (5) Zeitler 1984, (6) Bard 1983, (7) Rex et al. 1987, (8) Treloar et al. 1989, (9) Petterson et al, 1991, (10) Petterson et al 1990, 1993, (11) Coward et al. 1982, (12) Debon et al. 1987.

KARAKORAM BLOCK

This 70 to 120 km wide and 1,400 km long structural zone comprises the Karakoram crustal plate which is one of the fragments of the Cimmerian collage derived from Gondwanaland and accreted to Eurasia (see Chapter 3). In the north, the South Pamir Fault (Desio 1979) separates it from the Southern Pamir Block. To the east it is terminated by the Karakoram Fault and to the west by the Sarobi Fault. The Shyok Suture zone (MKT) forms its southern margin (Figs. 3.3 and 4.6). Parts of Karakoram have been mapped by many workers, noteably Ivanac et al. (1956), Desio (1964), Zanettin (1964), Stauffer (1975), Gamerith (1982), Calkins et al. (1981), Pudsey et al. (1985), Leake et al. (1989) and Searle (1991). They have introduced several overlapping formation names and a standard lithostratigraphic nomenclature for this region has not been formalised. In this section we give a general outline of the tectonostratigraphic sequences and their structure whereas the sedimentary, magmatic and metamorphic rocks of the region are discussed in greater detail in Chapters 5, 6 and 7 respectively.

Karakoram Batholith

The most ubiquitous feature in the Karakoram is the Axial Batholith (Ivanac et al. 1956, Desio 1979), which forms the central part of this belt and hosts the tallest mountain peaks of the region, including K2 (Photos. 2 and 27). It is comprised of a number of large parallel or en echelon plutons. From west to east these are Buni Zom, Zagar–Umalsit–Ghamu Bar, Dobargar–Darkot, Hunza and Baltoro plutons. They range in age from Jurassic to Miocene (Searle 1991). Along the northern margin of the Karakoram Batholith, near Baroghil Pass, there are small outcrops of pre-Ordovician granites. The granitic rocks and the surrounding sedimentary sequence have been metamorphosed to varying extent during at least three main thermotectonic events and a later retrograde phase (Chapters 6 and 7). The Karakoram Batholith divides the region into a northern and southern sedimentary belt. The western part of the region, the Tirich Mir zone, is separated from the main body of the Karakoram belt by the Tirich Fault (Fig. 4.40).

Tirich Mir Zone

This zone is largely comprised of the highly deformed and imbricated metasediments of the Arkari Formation (Jurassic ?) which mainly consists of dark grey phyllites with subordinate quartzites and marbles. Southward it is intruded by numerous stocks and sills of leucogranites and the metamorphic grade increases to the amphibolite facies (Chapters 5 and 7). The Kafiristan, Garam Chashma and Tirich Mir plutons are the largest batholiths in the region having K-Ar biotite ages of 48 ± 2 Ma, 20 ± 1 to 19 ± 5 Ma and 117 ± 5 Ma respectively (Searle 1991). They mainly consist of biotite granodiorite and augen gneiss.

Southern Sedimentary Belt

The Southern belt forms an arcuate, north-verging belt of highly deformed and imbricated metasediments thrust southward over the Kohistan–Ladakh sequence along the Shyok Suture zone or MKT. In the western part of the belt, between Chitral town and Mastuj, Late Paleozoic to Tertiary rocks are exposed. The Darkot Group is the oldest sedimentary unit with Carboniferous to Permian fossils. It consists of slates, schists and quartzites intercalated with crystalline limestones (Ivanac et al. 1956). It has been intruded by granites and is unconformably overlain by the Tertiary (?) Reshun Formation (Pudsey et al. 1985) which comprises red shales, limestones and conglomerates with clasts of underlying rocks. Near the Reshun village, the Darkot Group wedges out to be followed southward by a large anticline in which Chitral Slates (Jurassic ?) are exposed. Koghozi Greenschist and Gahirat Limestone (Cretaceous) crop out on its southern limb and the Krinj Limestone (Cretaceous) occurs on its northwestern limb (Fig. 4.40). The Krinj Limestone is unconformably overlain by the Reshun Formation. The Reshun Thrust Fault juxtaposes the Paleozoic sequence of the Northern sedimentary belt over the Reshun Formation and truncates the western part of the southern sedimentary belt.

In the region between Mastuj and Hunza, the southern sedimentary belt is almost entirely comprised of the Darkot Group. Large plutons of the Karakoram Batholith have intruded the Darkot Group. In the Hunza area and farther southeastward, in the Baltoro–Shigar region, portions of the Darkot Group have undergone higher grade metamorphism. In this region the belt is comprised of Dumordu, Ganchen and Askore Formations (Desio 1963, 1964, 1979).

The Shyok Suture Melange forms a thin tectonic wedge between the Kohistan–Ladakh arc sequence and the Southern sedimentary belt. It consists of grey to green slates, interbedded clastic sediments and blocks and clasts of greenstone, limestone, red shale, melange and sheared lenses of ophiolite.

Northern Sedimentary Belt

The Northern sedimentary belt extends westward from Shaksgam, through Upper Hunza and Chapursan Valleys, up to Baroghil Pass. It then bends southward and follows the Yarkhun Valley (Fig. 4.40). In the Upper Hunza region, it consists of three tectonostratigraphic units separated by two major faults (Fig. 5.4). An E-W trending reverse fault, the Upper Hunza Fault, has placed the southerly Gujhal Unit over the centrally located Sost Unit. The latter forms an antiformal stack and northward, across the Northern Fault, it is structurally overlain by the Misghar Unit (Gaetani et al. 1990). The sedimentary sequences in these tectonostratigraphic units range from Permian to Cretaceous in age and are largely comprised of interbedded limestones, shales and sandstones (Zanchi 1993, Gaetani et al. 1995). A conspicuous feature of the northern sedimentary belt is a black carbonaceous shale which overlies the Karakoram Batholith (Photo. 28). In different regions it has been variously named as Pasu Shale, Shimshal Shale, Singhie Shale, Baltoro and Rimo Shale. In the Upper Hunza–Chapursan region it is overlain by fossiliferous Permian carbonates (Guhjal dolomite).

The region between Upper Hunza and Upper Yarkhun Valleys (from Yashkuk to Showar Shur) is unmapped and the westward continuation of the Upper Hunza tectonostratigraphic units is uncertain. The Baroghil region contains Early Palaeozoic rocks and it is characterised by four tectonostratigraphic units. To the northwest lies the Tas Kupruk Unit which consists of Ordovician (?) to Permian metasediments, dolomites and volcanics (Fig. 5.4). Southeastward the Tirich Fault (?) separates it from the Karambar Unit which contains a Silurian (?) to Permian (or younger) terrigenous to shallow-water carbonate sequence. Southward the Karambar Unit is faulted against the Baroghil Unit which comprises a Devonian to Triassic sedimentary sequence. Farther to the southeast the Baroghil Unit has been thrust over the Axial Unit which consists of the Cretaceous to Lower Tertiary (?) Reshun Conglomerate, unconformably underlain by Upper Permian Carbonates. South of the Baroghil Pass, the Baroghil Unit unconformably overlies the pre-Ordovician Ishkarwaz Granite. This granite intrudes a metasedimentary sequence (LeFort et al. 1994) which we refer to as the Chikar Formation (Chapter 5). Southward the Chikar Formation and the Ishkarwaz Granite are covered and their contact relationship with the Karakoram Batholith (Darkot pluton) is not known. It is likely to be a tectonic boundary.

Dark splintery slates with siltstone laminae (not unlike the Pasu shale and other black shales mentioned earlier) unconformably overlie the Ishkarwaz Granite (Le Fort et al. 1994). At the base of the sedimentary sequence there is a conglomerate bed rich in chert followed by arkosic sandstones with brachiopods and overlain by slates containing Early Ordovician palynological assemblage (acritarchs). This sequence has been referred to as Baroghil sediments (Tongiorgi et al. 1994, Le Fort et al. 1994). In the upper part the slates contain metadolomites and quartzites with Middle to Late Ordovician conodonts (Talent et al. 1979). We propose that this sequence be named Baroghil Formation instead of Baroghil sediments. Above the Baroghil Formation there is an angular unconformity followed by a thick sequence of slates, dolomitic limestone and quartzites with Devonian fossils (Chilmarabad Formation). The latter is unconformably overlain by Permian carbonates (Hayden 1915, Talent et al. 1979, Tahirkheli 1982).

How far to the west and southwest the tectonostratigraphic units of the Baroghil region extend is not known because the area between Baroghil Pass and Mastuj has not yet been adequately mapped. The Northern sedimentary belt continues southwestward up to Chitral and beyond. In this region it is apparently confined to a narrow belt between the Tirich and Reshun Faults. Here it consists of an imbricate sequence of Devonian to Triassic (or Jurassic ?), fossiliferous quartzites, shales, crystalline limestones and dolomites (Charun Quartzite, Shogram Formation, Lun Shale). In this region the Reshun Fault.

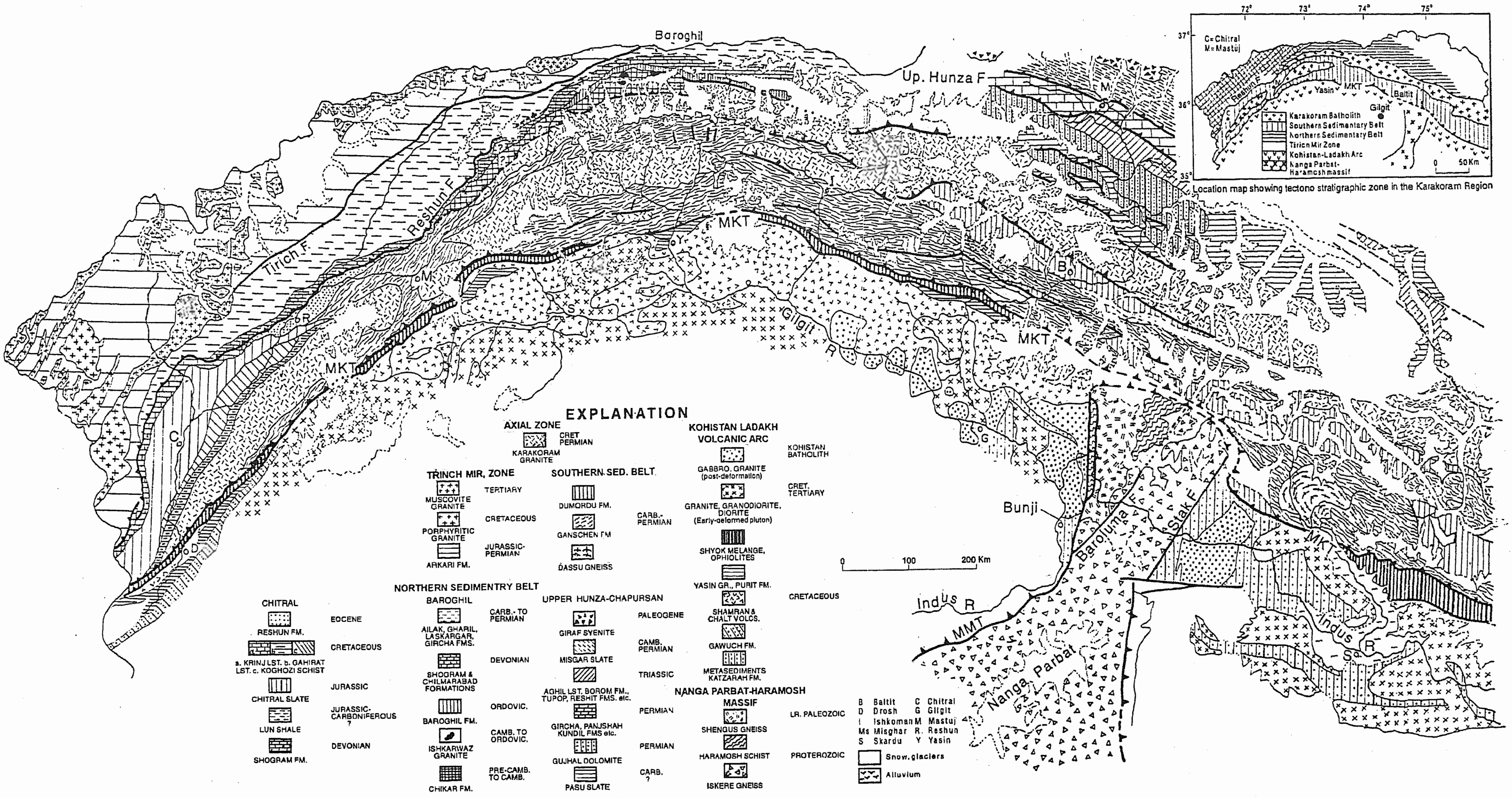


Fig. 4.40. Geological map of the Karakorams. (Basaed on data from Ivanac et al. 1956, Tahirkheli and Jan 1979, Gamerith 1982, Kazmi and Rana 1982, Pudsey et al. 1985, Leake et al 1989, Searle 1991, LeFort et al. 1994).

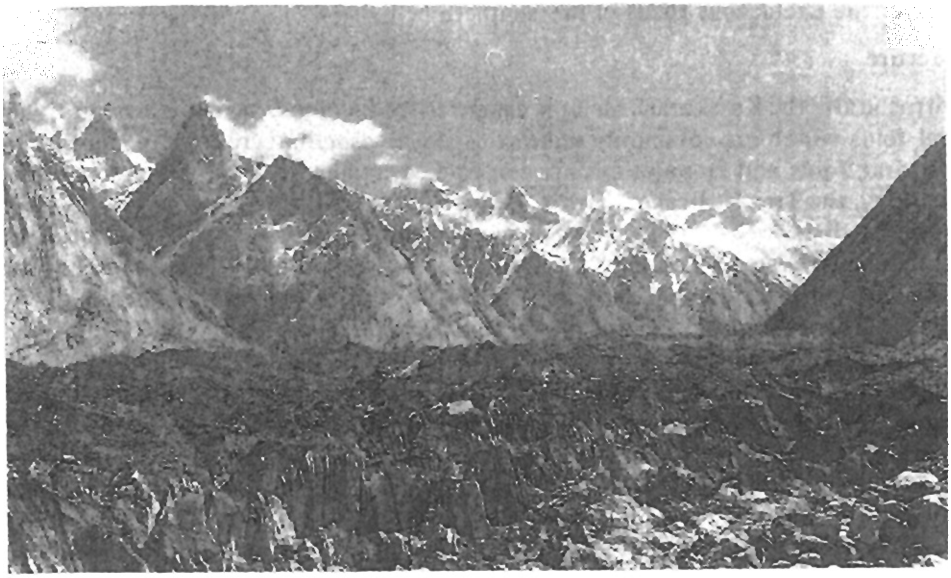


Photo. 27. View of Baltoro Glacier (foreground), Lobsang Spires (left) and Broad Peak 8047 m (right background). Baltoro granite crops out in the mountains above the glacier. Note the debris covered surface of the Baltoro Glacier. (Photo: *M. P. Searle*).

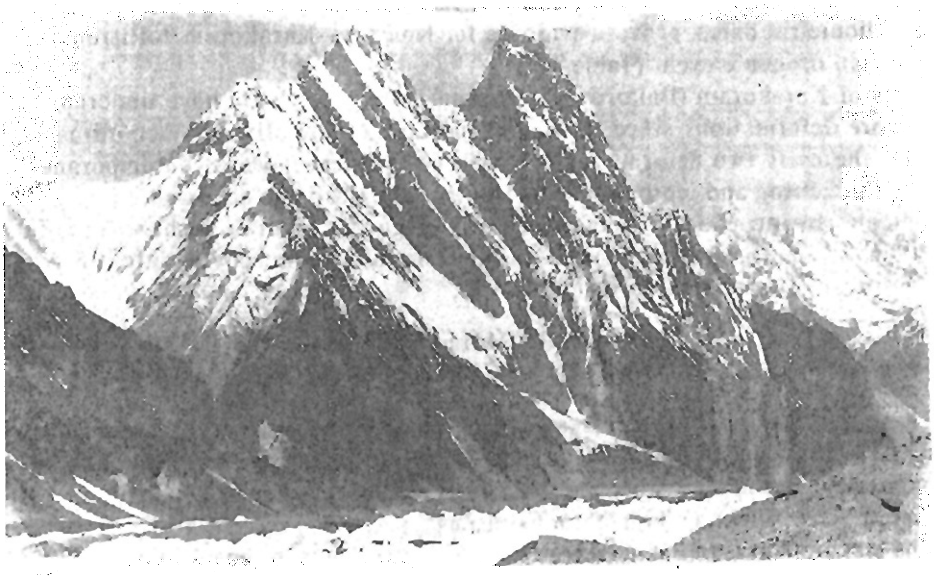


Photo. 28. View of the Mitre peak. The northern contact of Baltoro Granite is along this peak and the contact zone comprises a metamorphic aureole (andalusite hornfels) superimposed on sediments of northern Karakoram terrain (Baltoro black shale). g= Baltoro Granite, a= andalusite hornfels, M= Mitre peak, V= Vigne glacier (tributary of Baltoro). Photo: M.P. Searle.

which is a north dipping reverse fault, juxtaposes the Paleozoic sequence of the Northern belt over the Cretaceous rocks of the Southern belt.

Structure

Structurally the Karakoram block is characterised by broad, asymmetrical to tight, isoclinal folds which are commonly imbricated. Small isoclinal, recumbent folds, chevron folds, drag folds and crenulations are common. A number of large regional scale thrust faults, e.g., the Tirich , Reshun, Baltit, Upper Hunza and Chapursan Faults and the MKT, extend for distances up to 100 km or more. Smaller thrusts commonly imbricate the metasedimentary sequences and form thrust stacks. Normal and wrench faults also occur extensively but these are less conspicuous than the thrusts and reverse faults.

The Karakoram block has undergone multiple deformation since the Cambrian. The Baroghil section bears evidence of the earliest pre-Ordovician, probably Precambrian to Cambrian deformational, magmatic and metamorphic events. There is apparently a major hiatus during the Silurian because Silurian fossils have not yet been reported. In the Baroghil section the Ordovician succession underlies the Devonian with an angular unconformity (Tahirkheli 1982). The Batura region contains some evidence of deformation (molasse type Urdok Conglomerate), probably related to the rifting of the Karakoram plate from Gondwanaland during Late Permian or Early Triassic. In Upper Hunza and Chapursan an angular unconformity below the Middle to Late Cretaceous Tupop Formation and lack of Paleontological evidence for post-Middle or Late Jurassic sedimentation, coupled with the well constrained 208–163 Ma radiometric dates (Searle 1991) on the Hushe Gneiss Complex in the Shigar region suggest deformation during the Jurassic. In Chitral, Upper Hunza, Shaksgam and Baltoro, angular uncoformities above the Lower to Middle Cretaceous, and magmatism and metamorphism well constrained with radiometric dates, provide evidence for Kohistan–Karakoram collision and a Late Cimmerian orogenic event (Table 4.9).

Parts of Karakoram (Baltoro, Masherbrum, Mango Gausar) have undergone at least four more deformations since the Kohistan–Karakoram collision. According to Searle (1991), the early two deformations (Paleocene to Eocene) were contemporaneous with crustal thickening and regional metamorphism, the third one occurred during Oligocene as indicated by intrusion of undeformed Early Oligocene granites in sheared thrust sheets, and the last one took place during Mio-Pliocene (biotite-muscovite dates 15–5 Ma) and was associated with rapid uplift (Table 4.9).

KAKAR KHORASAN FLYSCH BASIN AND MAKRAN ACCRETIONARY ZONE

These two tectonic zones presently form separate and distinct structural units northwest and west of the Sulaiman–Kirthar fold-and-thrust belt (Fig. 4.6). However, both share a common depositional history and are comprised of Cenozoic Khojak Flysch. They are separated by the Chaman Transform Fault Zone.

The Kakar Khorasan Basin (Kazmi and Rana 1982) lies to the north of Zhob ophiolite-and-thrust belt (Photo. 12). As part of the large Katawaz Basin, it extends southward from near Kabul to Zhob and links up with the Makran Basin. The Kakar Khorasan Basin is bound to the south and east by the Zhob and Shinghar–Chukhan Manda Faults respectively. Westward it is terminated by the Chaman Fault (Fig. 4.6). It is filled by 4 to 6 km

Table 4.8. Stratigraphy and chronology of significant events in the Karakoram Block. (Compiled from Ivanac et al. 1956, Desio 1976 Calkin et al 1981, Pudse et al. 1985, Leake et al. 1989, Gaetani et al. 1990, 1993, Searle 1991, Zanchi 1993, Le Fort et al. 1994). Numbers under "magmatism" indicate: 1-Hunza dykes, Masherbrum and Baltoro plutons, Biange gneiss; 2-Yasin, Chingkiangla, Thallis, Batura granites; 3- Yasin Muztagh Tower, Hunza, Khunjerab, Darkot Pass granites and K2 gneiss; 4-Hushe Complex.

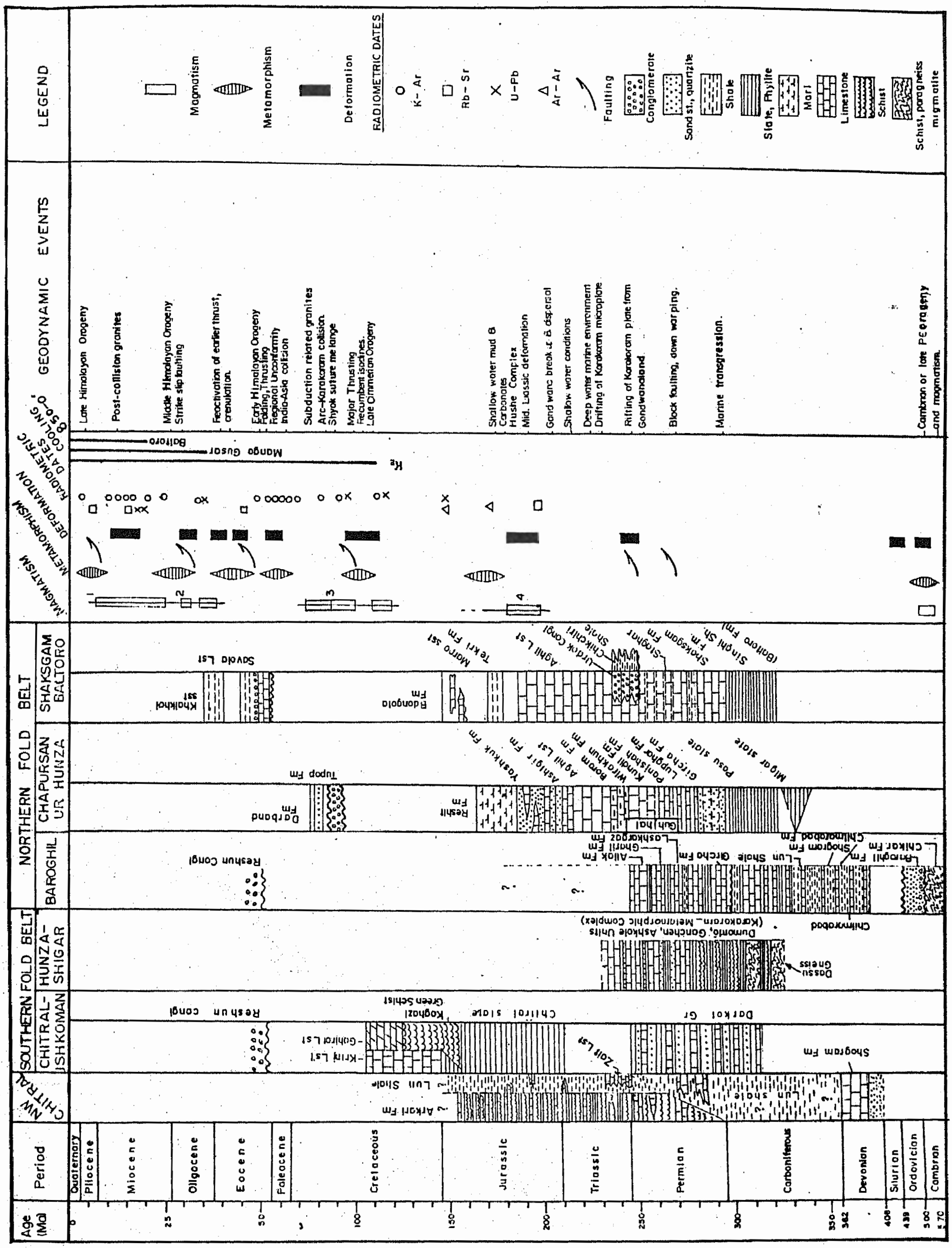
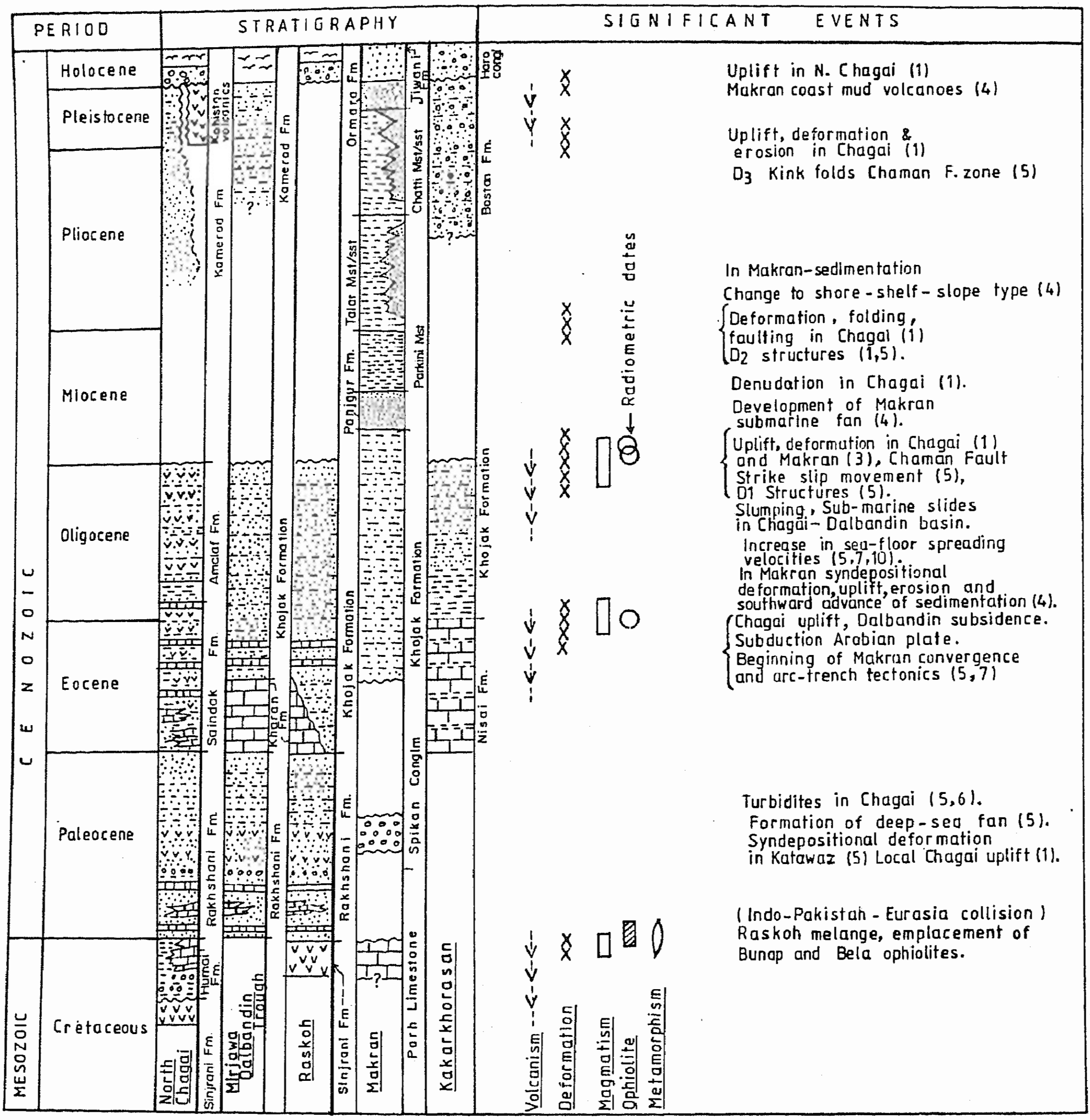


Table. 4.9. Stratigraphy and chronology of significant events in the Chagai and Makran Regions. Compiled from (1) HSC 1960; (2) Schreiber et al. 1972; (3) Whitmarsh 1974; (4) Mc Kenzie et al. 1976b;(5) Sillitoe et al. 1977; (6) Dykstra 1978; Arthurton et al. 1979; (8 and 9) Lawrence et al. 1979. 1982; (10) Harms 1984).



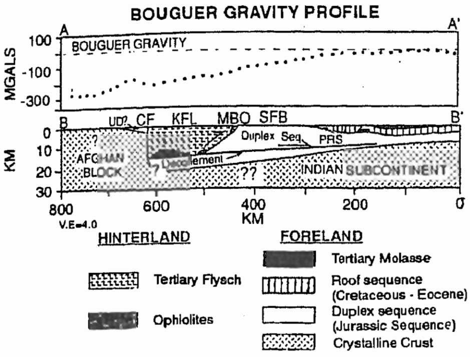
Uplift in N. Chagai (1)
Makran coast mud volcanoes (4)

Uplift, deformation & erosion in Chagai (1)
D3 Kink folds Chaman F. zone (5)

In Makran-sedimentation
Change to shore-shelf-slope type (4)
Deformation, folding, faulting in Chagai (1) D2 structures (1,5).
Denudation in Chagai (1).
Development of Makran submarine fan (4).
Uplift, deformation in Chagai (1) and Makran (3), Chaman Fault Strike slip movement (5), D1 Structures (5).
Slumping, Sub-marine slides in Chagai-Dalbandin basin.
Increase in sea-floor spreading velocities (5,7,10).
In Makran syndepositional deformation, uplift, erosion and southward advance of sedimentation (4).
Chagai uplift, Dalbandin subsidence. Subduction Arabian plate. Beginning of Makran convergence and arc-trench tectonics (5,7)

Turbidites in Chagai (5,6).
Formation of deep-sea fan (5).
Syndepositional deformation in Katawaz (5) Local Chagai uplift (1).

(Indo-Pakistan - Eurasia collision)
Raskoh melange, emplacement of Bunap and Bela ophiolites.



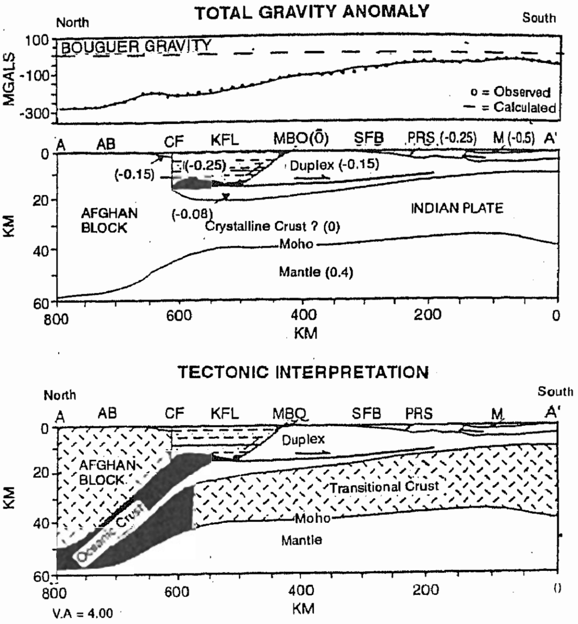


Fig. 4.41. Bouguer gravity anomaly profile, densityand tectonic model across Sulaiman Fold Belt, Chaman fault and Katawaz Basin CF= Chaman Fault, KFL= Katawaz Flysch Basin, MBO= Muslim-bagh Ophiolite, PRS= Passive roof sequence, SFB=Sulaiman Fold Belt and UD= undifferentiated (from Jadoon 1992).

thick sequence of flysch, deltaic and molasse type sediments (Ahmed 1991, Qayyum et al. 1996). Gravity modelling (Jadoon et al. 1992) shows that the transitional crust of the Sulaiman zone becomes thinner and deeper beneath Kakar Khorasan Basin and is overlapped by oceanic crust (Fig. 4.41). According to this model, the oceanic crust at the distal end of the Indian plate has underthrusted Afghan block. As a result, the Afghan block has thickened to about 57 km.

Eocene Nisai Limestone and Oligocene to Miocene Khojak Flysch are exposed in the Kakar Khorasan Basin. This sequence has been thrust southward over the Zhob Ophiolites. The flysch is capped by Pliocene or younger molasse (Table 4.9). This sequence forms broad synclines and tight anticlines cut by reverse faults (Fig. 4.12). The folds form a part of the Sulaiman oroclinal flexure. The structural trend changes from SW in the basin to NS near the Chaman Fault.

The **Chaman Fault** is a major left-lateral strike-slip active fault (Griesbach 1893, HSC 1960, Kazmi 1979, Lawrence et al. 1979) that has been the site of moderate to large earthquakes. This fault extends northward from Kharan to Kabul for 850 km and is considered as the western transform boundary of the Indian plate (Molnar and Tapponier 1975, Stöcklin 1977). It connects the Makran convergence zone with the Himalayan convergence zone (where Indo-Pakistan plate is underthrusting Eurasia). In Pakistan the fault marks the boundary between the Chagai magmatic arc to the west and the Khojak flysch basin to the east. The active trace of the fault is clearly seen on satellite imageries (Photo. 15). In the Chaman area the fault is characterised by a kilometre wide zone of fault-gouge, tectonic melange and exotic blocks (Lawrence et al. 1979). It contains clasts of volcanic rocks from Chagai region, limestone clasts from Kharan Formation, mafic-ultramafic rocks and sheared granitic rocks derived from Spina Tezha area (Fig. 4.42). The active fault splays out into several smaller faults. According to Lawrence et al. (1981), there has been 450 km of left lateral motion along the Chaman Fault.

The Chaman Transform Fault zone is 30–60 km wide and lies between the Chaman and

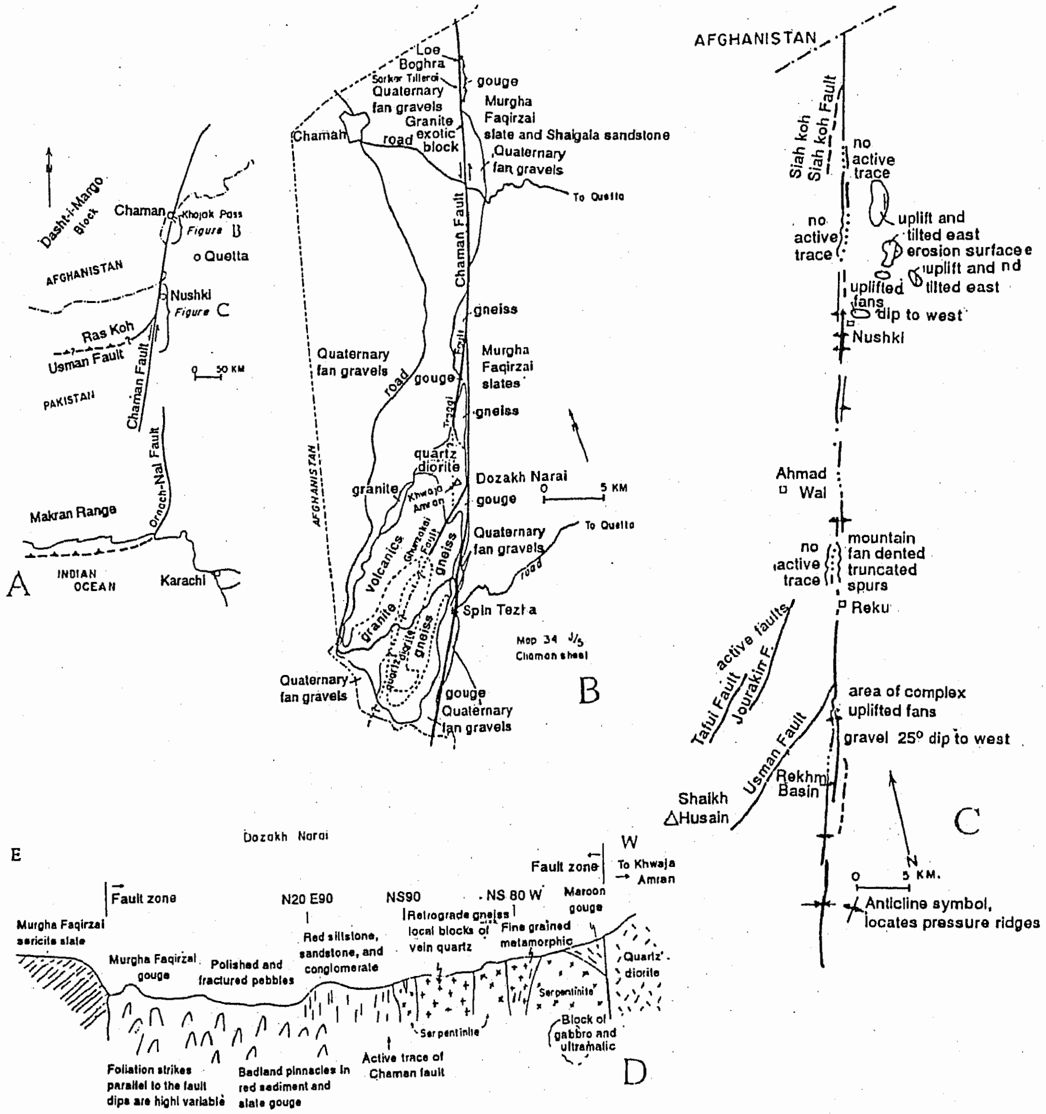


Fig. 4.42. Geological maps and section of the Chaman Fault zone in the Chaman-Nushki region. (From Lawrence et al. 1979).

Bhalla Dhor Faults. Between Qila Abdullah and Garruk (SE of Kalat) it extends for about 275 km in a N–S direction. It consists of tectonised and metamorphosed (slate to phyllite grade) sediments of Khojak Flysch. This region is characterised by small folds, strong axial plane cleavage and thrust faults. Numerous strike slip faults, parallel or subparallel to the Chaman Fault, form a dense network of fractures (Qureshi et al. 1993). Southward this zone widens and undergoes transition from a transform to a convergence regime and the structures gradually swing from north-south to southwest and then towards the west (Fig. 4.43).

In the Makran region, the Khojak Flysch forms a 7,000 m thick sedimentary prism

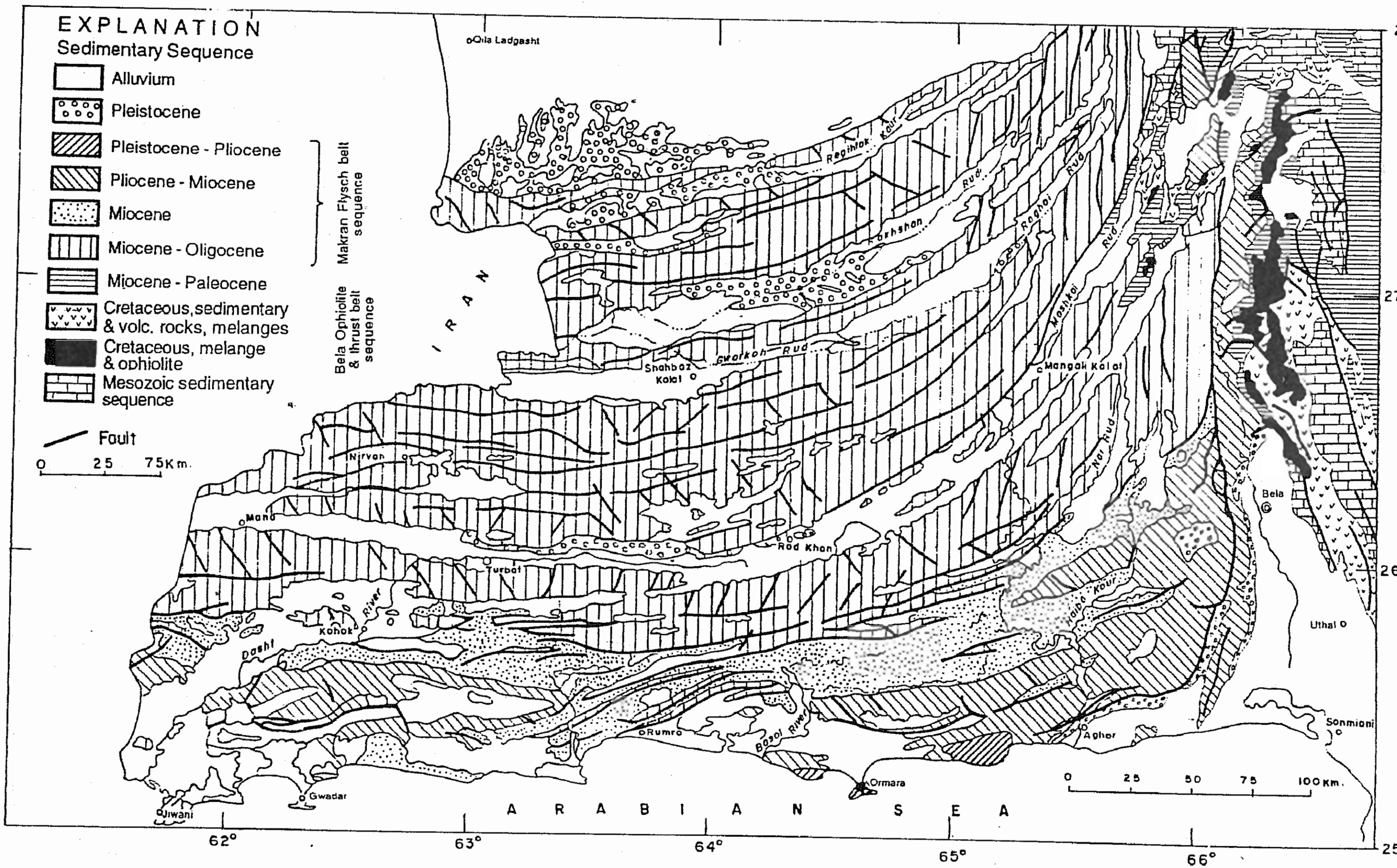


Fig. 4.43. Geological map of the Makran accretionary zone and the sothern part of the Bela Ophiolite and Thrust Belt. (Modified from Qureshi et al. 1993). See Photo. 14 also.

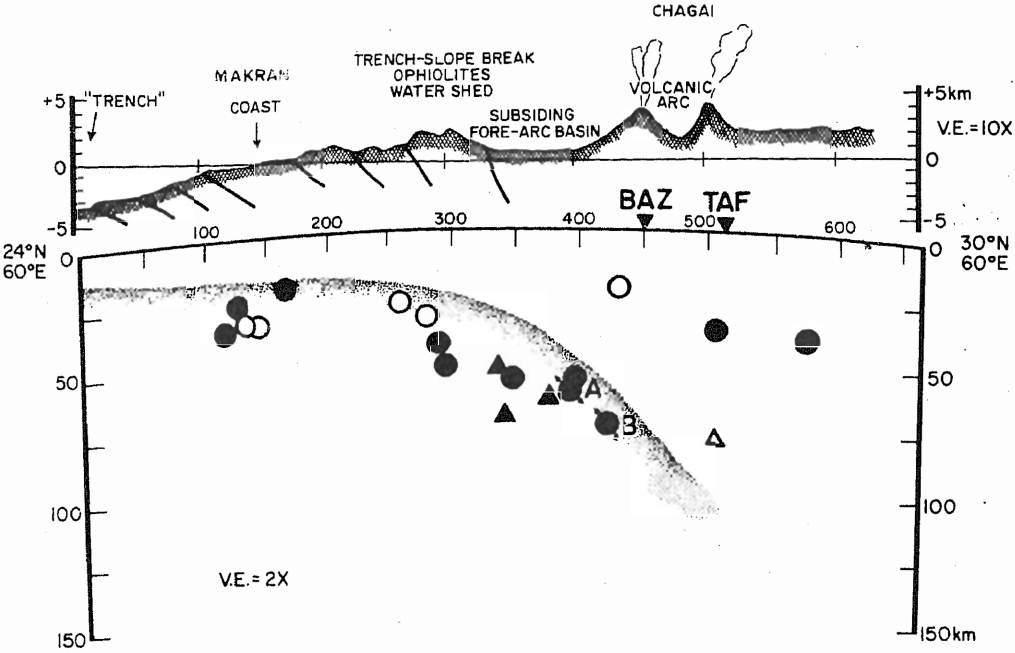


Fig. 4.44. Cross-section through the Makran region showing the earthquake hypocenters and inferred dipping Benioff zone (bottom), and topography and some surface tectonic features (top) along the same profile. *Bottom*: Circles represent events up to 200 km to the east of the section line, triangles up to 200 km to the west. Filled symbols represent events for which depth is constrained by at least one depth phase; open symbols represent events for which the depth is determined by minimising the residuals of first P arrivals only. Average depths determined independently by these two methods usually differ by no more than 10 km in cases when both are available in this region. (From Jacob and Quittmeyer 1979).

up to 15,000 m thick. The Ornach–Nal Transform Fault forms the eastern, and the Lut Massif in Iran forms the western margin of this zone. Northward it is covered by the Quaternary deposits of the Kharan Basin. The earliest rocks exposed in this region are Paleocene Ispikan Conglomerates (HSC 1960, Raza et al. 1991), overlain by Oligocene shales and a thick sequence of Miocene sandstones and shales (Hinglaj Formation). On the Makran Coast this sequence is overlain by Pliocene shelf-sandstones which dovetail with the mudstones of the neritic zone farther to the south (Table 4.9). The Khojak Flysch was deposited as a vast deep sea submarine fan similar to the present day Indus and Bengal Fans (Harms et al. 1984, Lawrence et al. 1981). The sedimentation pattern however changed to less extensive slope-shelf-shoreline deposits since Late Miocene.

The Makran convergence zone comprises an east-west trending fold-and-thrust belt which is intensely deformed in the north with tight asymmetrical, recumbent or isoclinal folds and well developed cleavage. The folds are cut by reverse and wrench faults (Photo. 14). The Nai Rud, Hoshab, Panjgur and Siahan are the major thrusts faults that divide the Makran region into a number of thrust-bound blocks (Kazmi and Rana 1982). This structure has produced an extensive valley-and-ridge topography. In southern Makran doubly plunging tight anticlines are separated by broad synclines.

Geophysical and geological data indicate that the Makran region is an active sub-

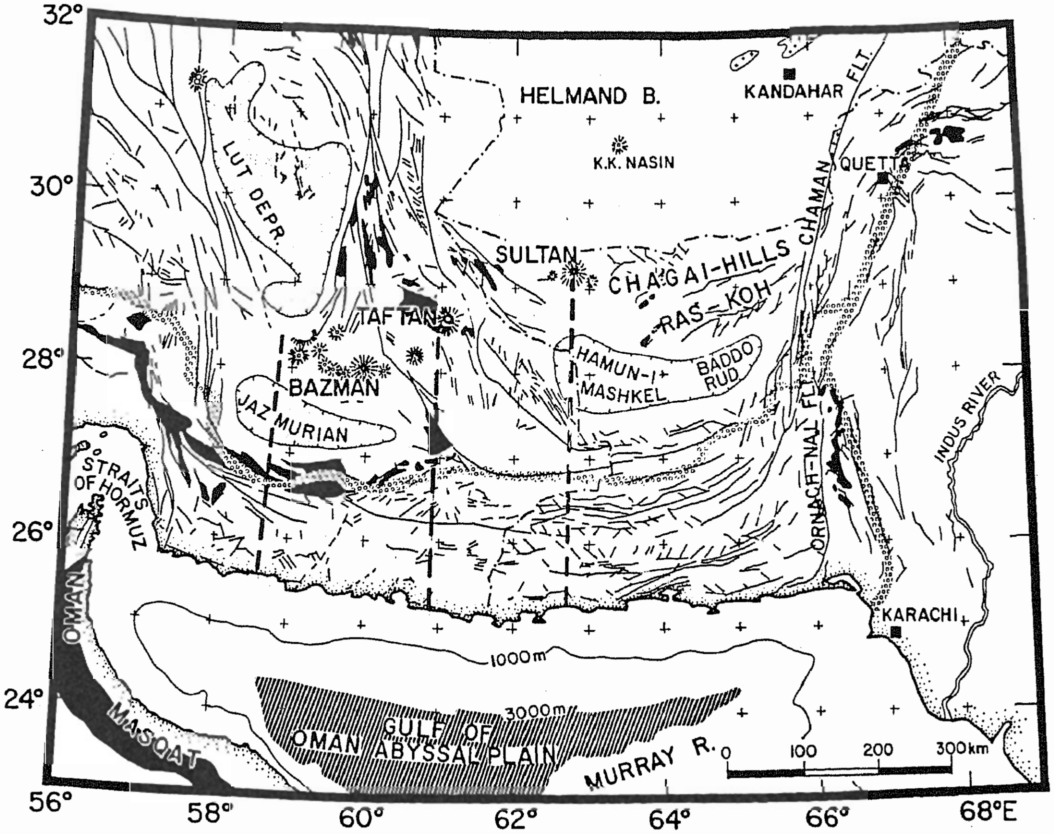


Fig. 4.45. Tectonic sketch map of Makran and surrounding regions (from Kazmi and Rana 1982, Berberian 1976). The abyssal part of the oceanic portion of the Arabian plate (in the Gulf of Oman), that is, the area with water depths larger than 3,000 m, is shown in dark diagonal hatching. Ophiolites and coloured melange shown in solid black. The pattern of present day drainage divides (light ring patterns) mimic approximately the plate boundaries and the triple junction, probably as they may have existed some 30 (or more) m.y. ago. Deep sediment-filled interior basins, probably with recent histories of subsidence, are shown in light shading. Of these, the Jaz-Murian and the Hamun-i-Mashkhel depression are interpreted by Farhoudi and Karig (1977) as forearc basins, they lie seaward of the volcanic arc. Transverse breaks defining the segmented subduction zones (Dykstra and Birnie 1979) are shown by thick dashed lines. (From Jacob and Quittmeyer 1979).

duction zone. Earthquake hypocentres define a dipping seismic zone which is at least 80 km deep (Fig. 4.44) and earthquake focal mechanism suggest down-dip tension associated with subcrustal events in the seismic zone. North of Makran, the Chagai volcanic arc coincides with the 100 km depth contour of a down-dip projected extension of this seismic zone. Earthquake-related coastal uplift has occurred in the region. The 400–800 km wide arc-trench gap in the Makran subduction zone is exceptionally wide. A large part of the accretionary prism is subaerially exposed (Jacob et al. 1979).

The pattern of the Quaternary volcanoes, the physiography, and the seismological data suggest that the Makran subduction zone is segmented (Fig. 4.45) and that the

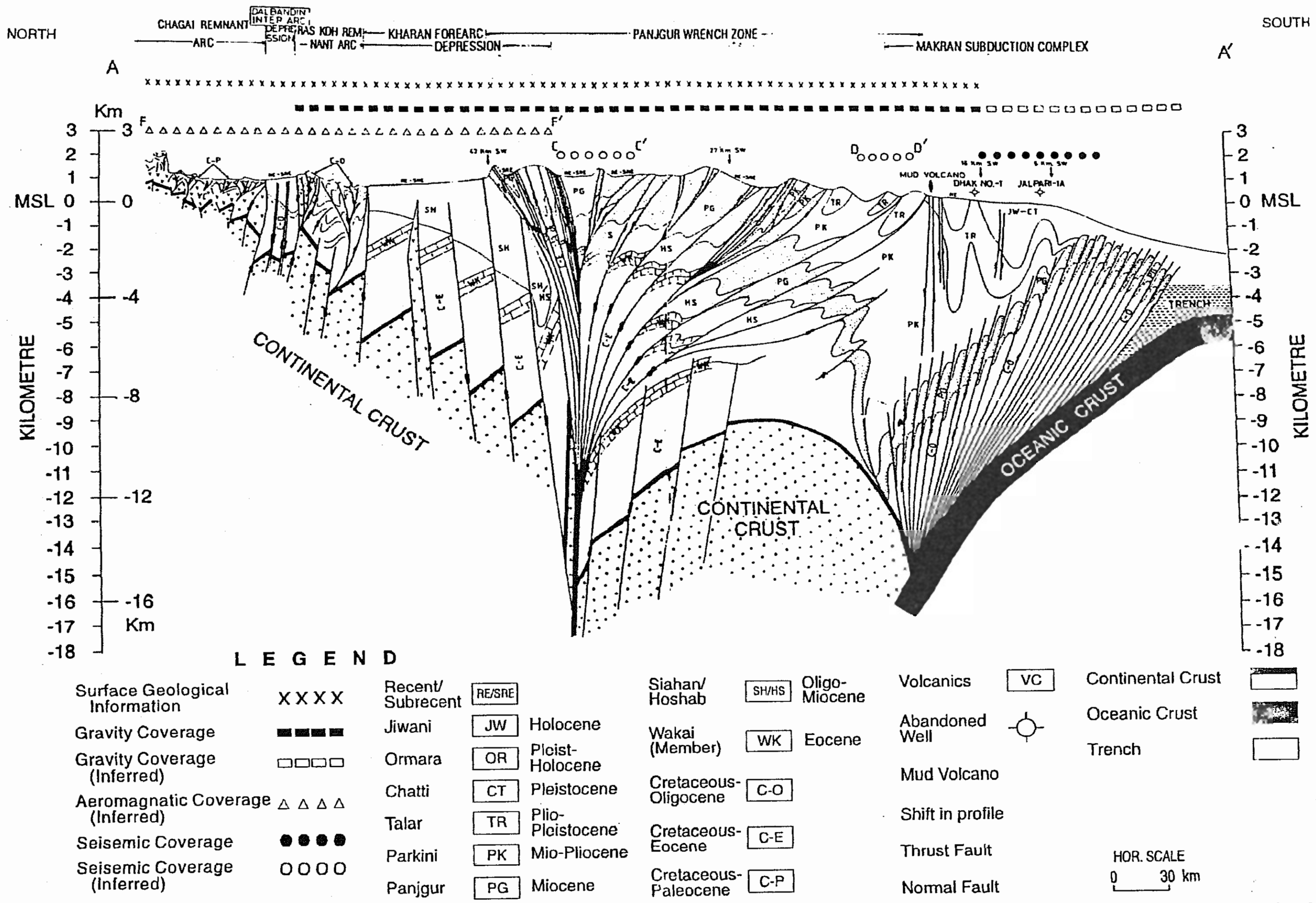


Fig. 4.46. Structural and tectonic model of the Chagai Volcanic and Magmatic Arc and the Makran Accretionary Zone. (From Raza et al. 1991).

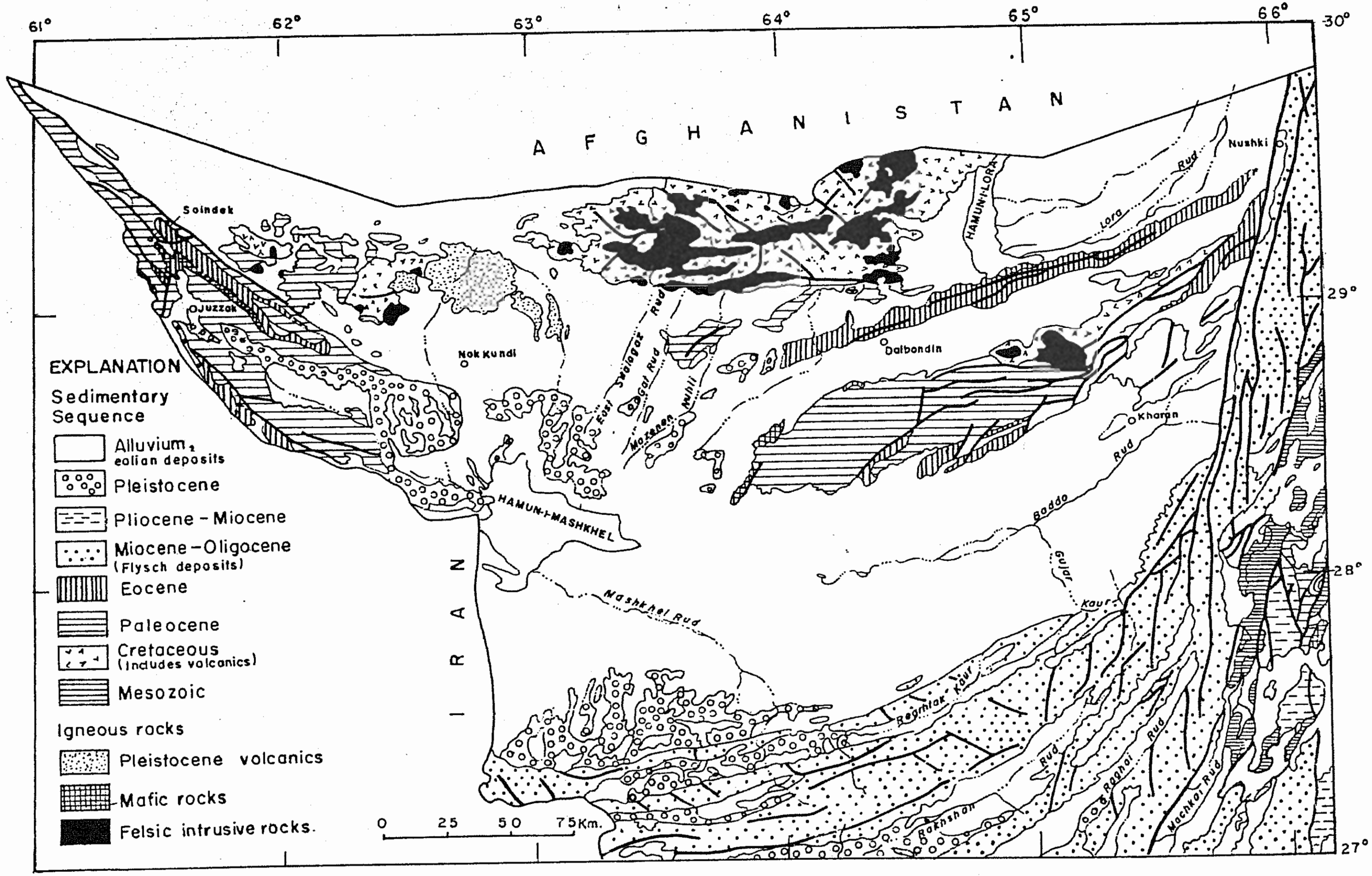


Fig. 4.47. Geological map of the Chagai Volcanic and Magmatic Arc and a part of the Chaman fault zone and northern part of the Makran Accretionary Zone. (Modified from Qureshi et al. 1993). See Photo. 15 also.

subducting lithosphere is broken along large basement faults perpendicular to the trench (Dykstra et al. 1979). These faults, known as transverse breaks, divide the subducting plate into various dipping digitate segments (Carr et al. 1973, Barazangi et al. 1976).

Raza et al. (1991) have, however, proposed a different structural model to explain the tectonic features of the Makran arc-trench system (Fig. 4.46). According to them the arc was destroyed in Late Tertiary and modified to a non-arc subduction regime. The hind part of the formerly leading edge of the subducted oceanic slab became the new leading edge and the active subduction shifted to the line of present mud volcanoes strung along the Makran Coast. Transform movement on the east (Ornach–Nal Fault) and the west (Neh Fault) and digitation of the continental slab instead of the oceanic slab, as postulated by Dykstra et al. (1979) have contributed to the formation of present day structural features of this region.

KHARAN BASIN

The Kharan fore-arc depression lies to the north of the Makran convergence zone. This is an enclosed drainage basin filled with Quaternary deposits and apparently underlain by Tertiary sediments. Aeromagnetic and gravity data suggest that the sedimentary pile in the basin may be 5,000 to 7,000 m thick (Raza et al. 1991). A model of Kharan Basin largely based on available integrated aeromagnetic, gravity and geological data is shown in Figure 4.46.

CHAGAI MAGMATIC ARC

North of the Kharan depression, the Chagai arc is an east-west-trending, arcuate, south-verging magmatic belt comprised of Cretaceous to Tertiary sediments and volcanics. It is intruded by Tertiary granites, diorites and mafic dykes and sills. Small tectonised blocks of ultramafic rocks occur in the southern part (Ras Koh Range). Quaternary volcanoes dot the landscape in the northwestern section of the arc (Fig. 4.45). This magmatic arc forms a part of the Makran trench arc system on the southern margin of the Afghan and Lut blocks (Jacob et al. 1979). It comprises three main tectonic components. From north to south these are: (i) the North Chagai calc-alkaline and magmatic belt, (ii) Mirjawa–Dalbandin trough and (iii) the Ras Koh uplift block (Kazmi and Rana 1982). Eastward the Chagai arc is abruptly truncated by the Chaman Fault (Photo. 15,Fig. 4.47)).

The Chagai magmatic arc has been discussed in detail in Chapters 5, 6 and 7. Its general stratigraphy and geochronology has been summarised in Table 4.9. From the fore-going account it is clear that the Kakarkhorasan, Makran and Chagai regions are all parts of Makran–Chagai trench-arc system which developed in response to northward movement of the Indian and Arabian plates and subduction of the Neotethys under the southern margin of the Afghan block (see Chapter 3). Cretaceous submarine volcanism associated with periodic uplift eventually led to the formation of Chagai andesitic volcanic arc. It appears that the Ras Koh tectonic block formed at about the same time south of the Mirjawa–Dalbandin trough. The origin of this block is not clear though various models have been proposed (see Chapter 6). There is evidence of ophiolite obduction, metamorphism, magmatism and deformation between Late Cretaceous and Early Paleocene. This was followed by deposition of turbidites with syndepositional deformation and Paleocene volcanism. Northward subduction of the oceanic lithosphere and development of the Makran trench started between Late Eocene and Early Oligocene and apparently

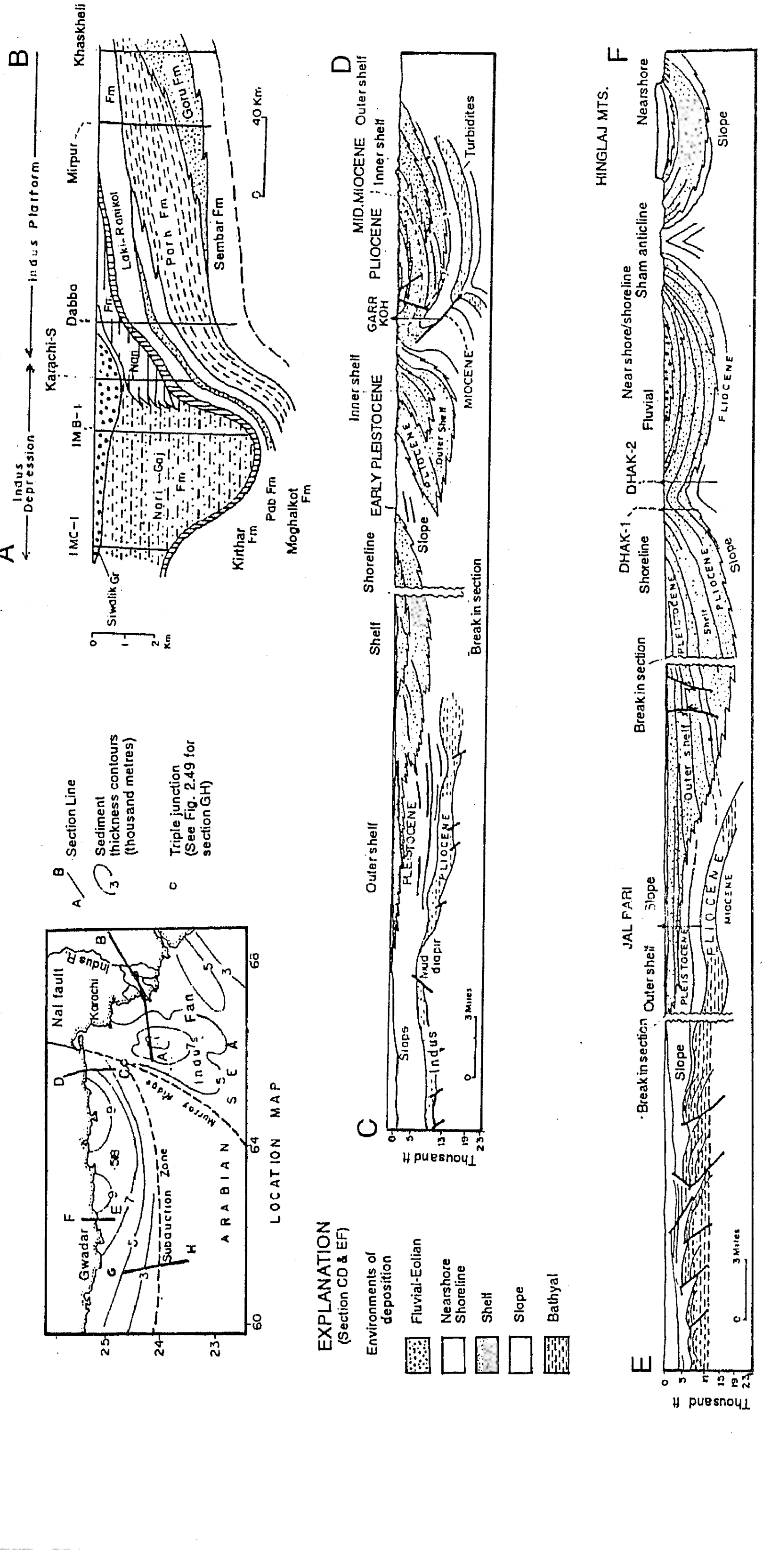


Fig. 4.48. Geological cross-sections of Pakistan Offshore. (AB based on data from Quadri and Shuaib 1986; CD, EF modified from Harms et al. 1984).

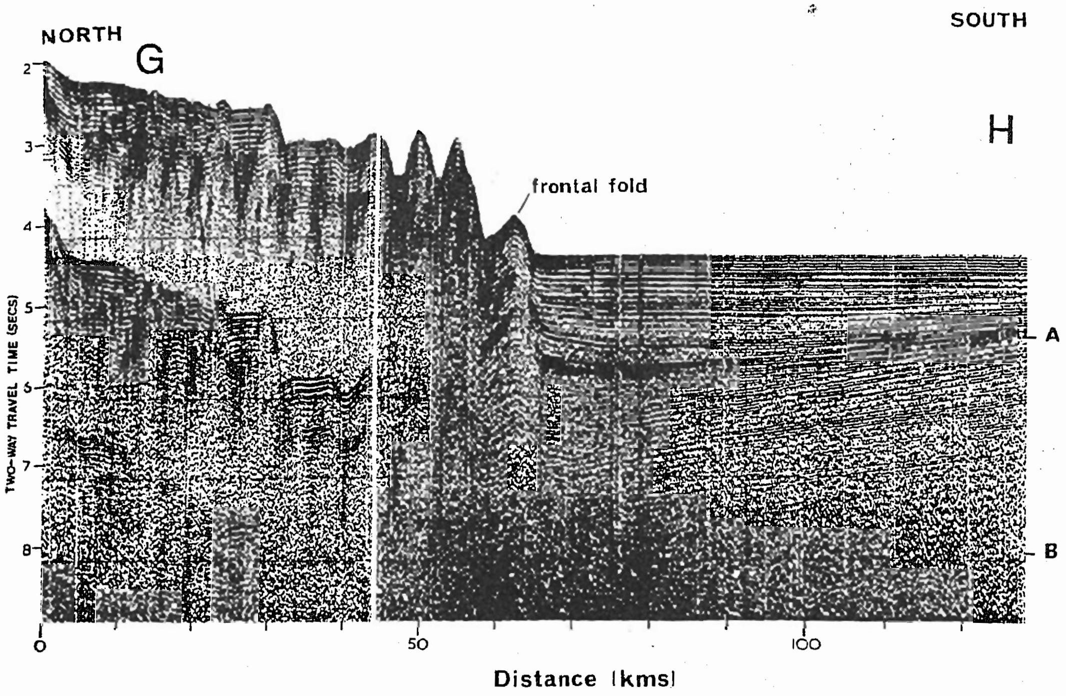


Fig. 4.49. Continuous seismic profile (line GH in Fig 4.48) across the Makran continental margin. Notice the frontal fold at 65 km, the inter-fold basins partially filled with sediment between 40 and 60 km and the ridges buried beneath secondary sediments between 0 to 40 km. Horizon 'A' divides the topmost flat-lying sediments from uniformly dipping layers below. Horizon 'B' is the deepest observable reflector, comprising numerous hyperbolae. Vertical exaggeration at sea floor approx. 10x. (From White 1979).

coincided with a second uplift and deformation in Chagai region as indicated by 35 Ma zircon fission track age (Dykstra 1978) and by the increase in sea-floor spreading velocities around 35–30 Ma (Lawrence et al. 1981, McKenzie et al. 1976). This may have been the period when the Chaman Fault developed into an effective transform system with subsequent, short and intermittent episodes of rapid movement.

The Makran trench sediments formed as a deep-sea fan. Deformation caused shoreward uplift of the sedimentary prism with the concomitant southward shifts in the depositional front. In Chagai–Ras Koh region and in Kakar Khorasan a long gap in the stratigraphic sequence during Miocene, and 20 Ma zircon fission track age (Dykstra 1978) and 21 Ma K-Ar hornblende and biotite date (Sillitoe and Khan 1977) highlight the most significant orogenic event in the whole region. It was associated with volcanism in northern Chagai, extensive magmatism in Chagai and Ras Koh, intensive folding, faulting and thrusting in the entire region, tectonism in Chaman Transform Fault zone, bifurcation of Makran and Kakar Khorasan Basins and uplift and erosion of the arc. In the Makran region it was followed by a change in sedimentation pattern from a deep-sea fan type of flysch to a shore-shelf-slope type of sedimentary sequence. Further volcanism occurred in the Chagai arc during Pleistocene and Early Holocene accompanied by two more phases of uplift and faulting. During this latter period uplift, folding and faulting occurred on Makran Coast, and several fault related mud volcanoes formed in southeastern part of the region.

PAKISTAN OFFSHORE

The Pakistan Offshore extends for 700 km from Rann of Cutch to the Iranian border (near Gwadar). It comprises two distinct structural and sedimentary basins, the Indus and Makran Offshore Basins, which are separated by the Murray Ridge. This ridge is an extension of the Owen Fracture Zone and forms the boundary between Indian and Arabian plates (Figs. 4.4 and 4.48).

Makran Offshore

This zone is characterised by relatively narrow shelf and slope which form the northern margin of the Gulf of Oman Abyssal Plain. In this region the Arabian plate is being subducted beneath Makran and the subduction zone is located near the base of the continental slope. As discussed earlier, the Makran region is comprised of an accretionary wedge of deformed sediments piled up at an oceanic subduction margin and it is part of a vast trench-arc system (Fig. 4.44). Seismic profiles indicate that the sediment pile on the Makran continental margin is about 5 km thick (White 1979). In the abyssal plain these sediments are horizontally disposed but shoreward they are abruptly deformed into an approximately 50 km wide fold and thrust belt (Fig. 4.49). Only the upper 2.5 km of the sedimentary sequence is folded and forms a decollement zone. As revealed by offshore drilling, most of this sequence consists of Early Miocene and younger sediments (Fig. 4.48).

Indus Offshore

This region lies between the Murray Ridge and the Indian Coast. It is characterised by a wide shelf and slope comprised of the vast Indus Fan (see Chapter 2). It is comprised of two main structural zones, the Indus platform zone to the east and the Indus depression to the west. The Indus Offshore basin contains a thick sedimentary sequence ranging from Cretaceous to Recent (Table 4.2, Fig. 4.48). Seismic profiles of this region indicate gently dipping structures with abundant down-to-the basin type normal faults (Raza et al. 1990b).

5

Stratigraphy

Rock formations ranging in age from Early Proterozoic to Recent are well exposed and extensively developed in Pakistan. Precambrian metasediments cover the Archean gneisses of the Indian craton and form the upper part of the basement complex on the Indus platform. Inliers of these metasediments are exposed in the upper Indus Plain (Fig. 4.9). Tectonised Precambrian and Paleozoic metasediments, intruded by various types of magmatic rocks, are exposed in anticlinal cores and in thrust sheets in the Himalayas and the Karakoram. Mesozoic and Paleogene sedimentary rocks form a thick sequence of pericratonic shelf deposits in the Himalayan–Sulaiman–Kirthar fold belts. Tectonically mixed sequences of Cretaceous and Paleogene sedimentary and volcanogenic rocks are found in the Chagai–Ras Koh region. The Cenozoic sedimentary cover attains its maximum thickness in the foredeeps and the adjacent areas of the fold belts and in the Makran flysch zone. Unconsolidated Quaternary sediments cover vast areas in the intermountain basins, Indus Plain and the Thar Desert. Glacial and fluvioglacial deposits are confined to the valleys and gorges of the Karakoram and the Himalayas. These sedimentary sequences have been briefly described in this Chapter. The stratigraphy of Pakistan has been often discussed by previous workers (William 1959, HSC 1960, Shah 1977) in terms of depositional basins. Pakistan is comprised of two major sedimentary basins namely the Indus Basin, which covers the northwestern portion of the Indo-Pakistan crustal plate (regions which include the Indus platform and the Himalayan and Sulaiman–Kirthar fold belts), and the Makran Basin which covers the Tethyan zone west of the Indus Basin. Raza et al. (1989) have neatly defined the architecture of these two basins (Fig. 5.1) and have divided them into a number of sedimentary zones which closely correspond to the tectonic zones defined by us in Chapter 4 and as such, in this chapter, we have largely discussed the stratigraphy in terms of these tectonic zones.

Stratigraphic research in Pakistan may be traced back to the middle of the last century. Since then a large and impressive array of authors have contributed to the stratigraphy of Pakistan. As may be expected there has been a proliferation of stratigraphic nomenclature, and different authors have coined different names for the same suite of rocks. To systematise these, the National Stratigraphic Committee has reviewed and approved a set of standard lithostratigraphic names (Cheema et al. 1977, Fatmi 1973, Shah 1977). But these are incomplete and partly out of date because in recent years there has been a great spurt in geological research in Pakistan, new discoveries have been made, stratigraphy of many regions has been revised and new lithostratigraphic names have been proposed. In this Chapter, as far as possible, we have summarised the results of these recent stratigraphic investigations and integrated them in the existing scheme.

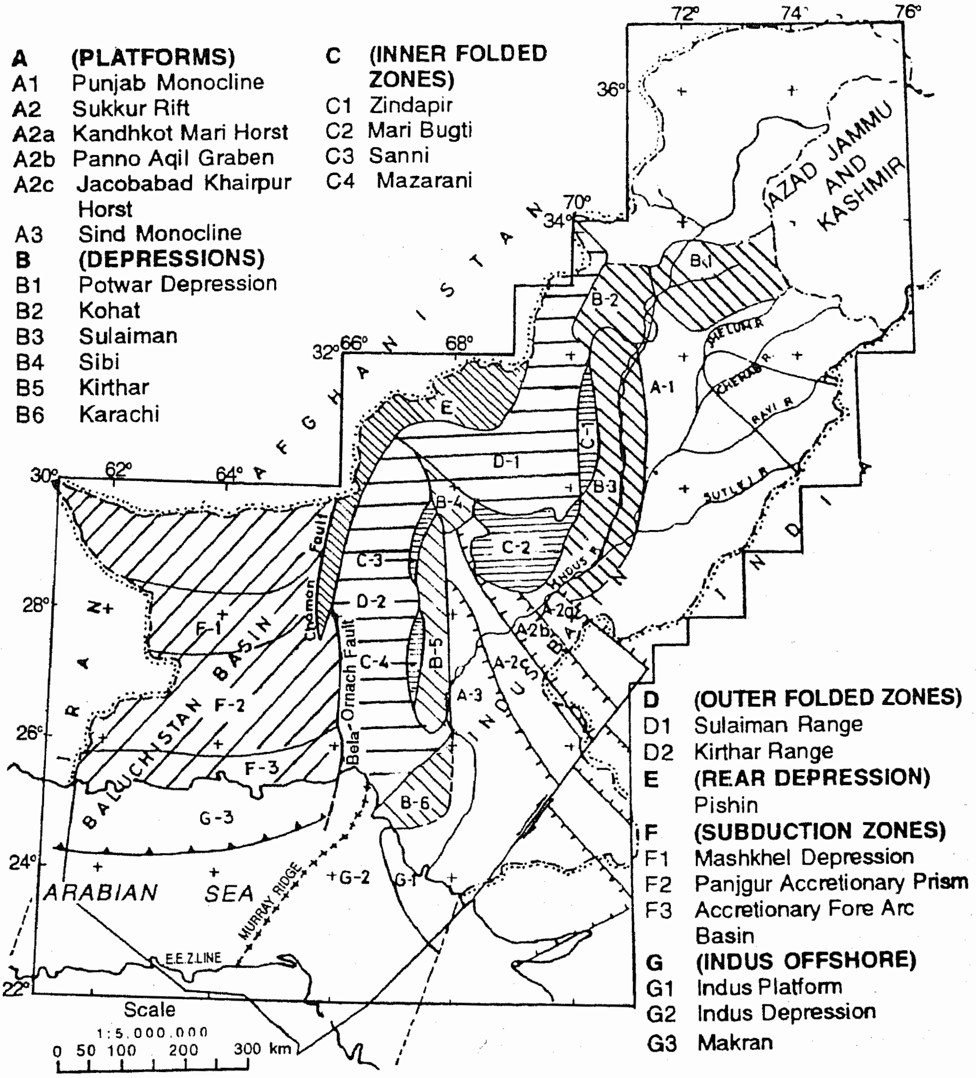


Fig. 5.1. Sedimentary zones of Pakistan (from Raza et al. 1989).

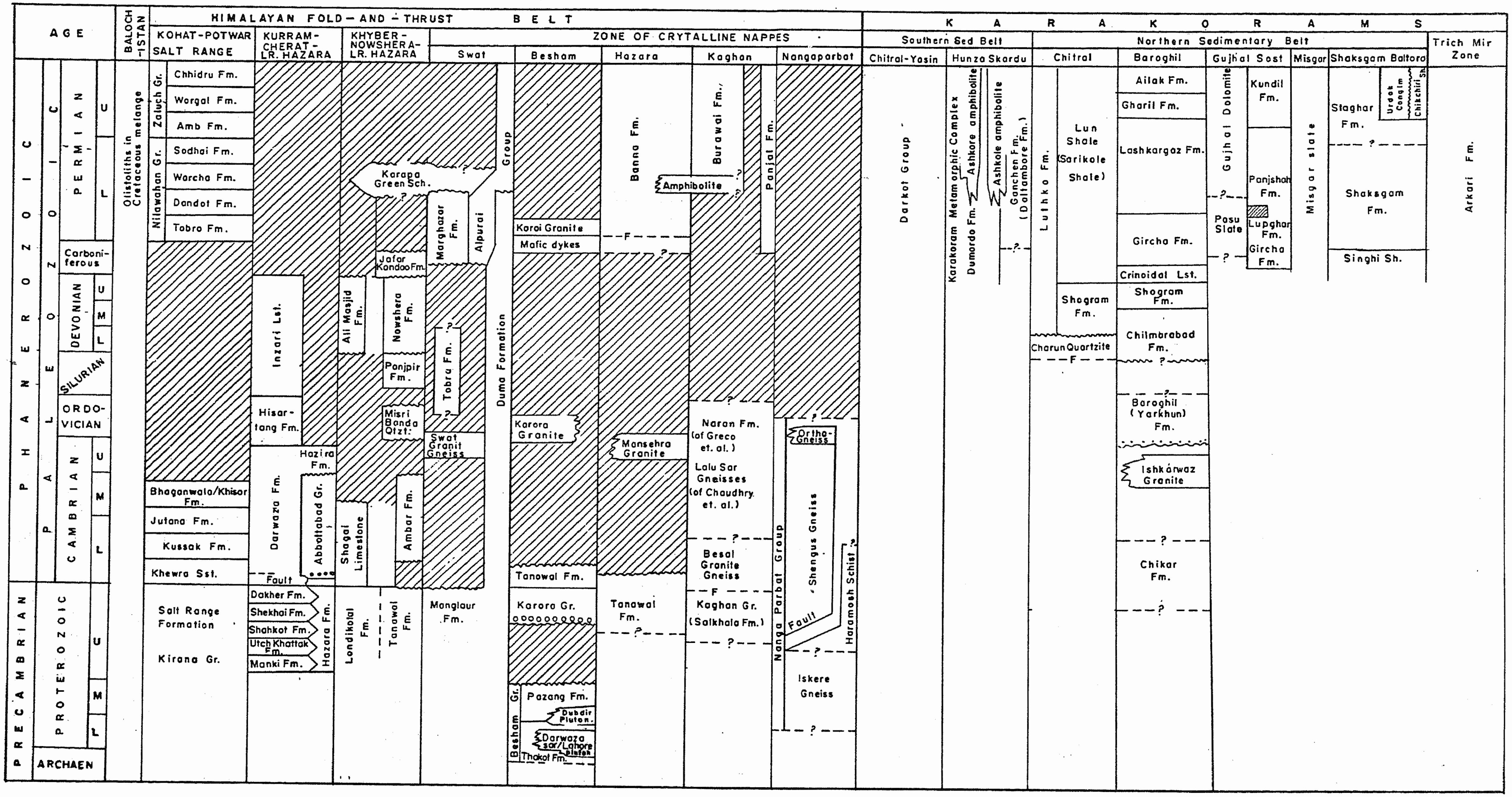
PRECAMBRIAN

Precambrian sedimentary and metasedimentary rocks crop out in the Upper Indus Plain, Salt Range, Attock–Cherat Ranges, the Himalayas and possibly in the Karakorams also. Because of their unfossiliferous nature, metamorphism and occurrence in nappes, disjointed thrust slices or fault blocks, their correlation has been difficult and is shown in Table 5.1.

Indus Platform

Kirana Group: Unfossiliferous, Precambrian metasediments and metavolcanics are exposed in the form of several small hillocks between Sargodha and Shahkot and were earlier described by Heron (1913). Kazmi (1964) introduced the name "Kirana Group"

Table 5.1. Correlation of Precambrian and Paleozoic sequences.



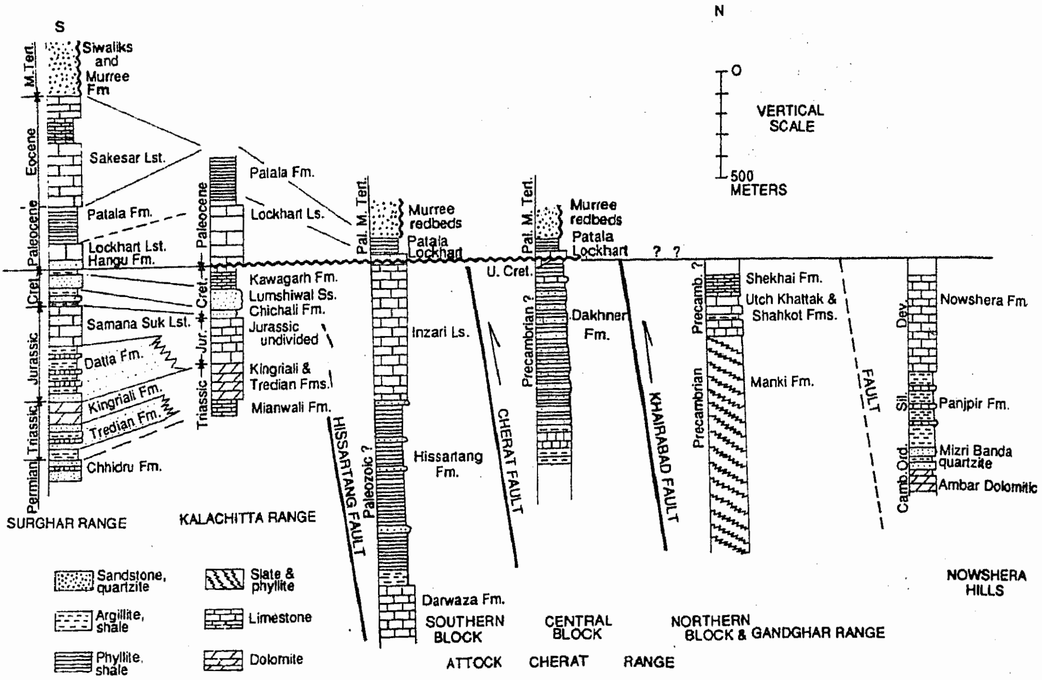


Fig. 5.2. Tectonostratigraphic sequence in Surghar, Kalachitta, Attock–Cherat Ranges and the Nowshera Hills (from Yeats et al. 1987, Hyalland et al. 1988, Hussain et al. 1989).

for these rocks, for which a Rb/Sr date of 870 ± 40 m.y. has been obtained by Davies and Crawford (1971). The Kirana Group consists of phyllites, slates, quartzites and subordinate conglomerates and volcanics. It has been subdivided into five formations: the **Hachi**, **Taguwali**, **Asianwala**, **Hadda**, and **Sharaban Formations** (Alam 1983). The total thickness of the sequence is quoted as more than 2,330 m. Metamorphic basement rocks have also been encountered by a number of wells drilled to the north and south of the Sargodha High (Figs. 4.8 and 4.9). For further details see Chapter 6.

Salt Range

Salt Range Formation: Along the northwestern flank of the Indian Shield, the basement rocks are overlain by the non-metamorphic sediments of the Salt Range Formation (Asrarullah 1967), formerly described as "Saline Series" by Gee (1945). Type locality is the Khewra gorge in the eastern Salt Range. The Salt Range Formation is widely distributed in Salt Range, between Jogi Tilla in the east and Kalabagh in the northwest. It was also encountered by wells in the Kohat–Potwar foredeep and in the southern Punjab area, along the shallow eastern flank of Sulaiman foredeep.

The major part of the Salt Range Formation consists of red, gypsiferous claystone ("Salt marls") without any apparent bedding. Intercalated thick salt bodies are being mined at Khewra, Warchha, Kalabagh and some other localities in Salt Range.

The middle part of the Formation contains an alteration of gypsum, dolomite, shale, siltstone with oil-shale layers, particularly in the western Salt Range. The top of the Formation is formed by a gypsum layer containing high-grade oil-shale or a layer of highly altered volcanic rock known as Khewra Trap (see Chapter 6). The thickness of Salt Range Formation is more than 800 m at the type locality (base not exposed). In the Kohat–Potwar foredeep the thickness locally exceeds 2,000 m as result of secondary salt migration and accumulation induced by decollement and southward-thrusting of the overlying sedimentary sequence. The Formation is devoid of fossils. Due to its position below fossiliferous Lower Cambrian and above the metamorphic Precambrian basement, it is considered as Late Proterozoic. This is in accordance with the results of sulphur-isotope measurements carried out on gypsum samples from the top of Salt Range Formation which indicate an age of about 600 m.y. (H. A. Raza, person. commun.). The evaporitic sequence of Salt Range Formation was deposited in a restricted, shallow-marine environment under arid conditions.

Peshawar Basin

The rocks exposed south of the Peshawar depression, in the Nowshera, Gandghar and Attock–Cherat Ranges, are mostly Precambrian to Devonian metasediments. The Attock–Cherat Ranges consist of three structural blocks separated by thrust faults (Yeats and Hussain 1987). The northern block is composed of the Precambrian Manki, Shahkot, Utch Khattak and Shekhai Formations, whereas the central block is largely made up of Precambrian Dakhner Formation (Fig. 5.2). These formation names were initially defined by Tahirkheli (1970) and have been since modified and adopted by Yeats and Hussain (1987, 1989) and Hylland et al. (1988).

Manki Formation: This Formation consists of dark grey, thin-bedded argillite, slate and phyllite at least 950 m thick. The argillite and slate contain two sets of cleavage, one parallel to the bedding, the other set being the axial-plane cleavage. The metamorphic grade increases northward from argillite and slate to phyllite with extensive development of quartz veins (Hylland et al. 1988). The base of the Manki Formation is not exposed. Upwards it grades into the overlying Shahkot Formation. Based on lithologic correlation with the Hazara Formation, Hussain et al. (1990) considered the Manki Formation to be of Precambrian age.

Shahkot Formation: It is exposed in the Gandghar and Attock–Cherat Ranges and consists of limestone, argillite and shale. At the base of the Formation there is a 10 m thick fine- to medium-bedded cherty limestone. The argillite in the upper part contains thin, discontinuous algal limestone. The Shahkot Formation is about 300 m thick and it is conformable with the underlying Manki Formation and the overlying Utch Khattak Formation (Hylland et al. 1988).

Utch Khattak Formation: This Formation is exposed in the Attock–Cherat and Gandghar Ranges. It consists of limestone, argillite and shale 200 m to 250 m thick. The lower part of the Formation contains a 10 m to 70 m thick limestone unit that is grey, thin bedded, fine to medium grained and in places contains stromatolites. It is overlain by dark greenish grey thinly laminated argillites interbedded with grey to brown, thin bedded shale. A conglomerate within the formation contains clasts of Manki and Shahkot Formations. The Utch Khattak Formation is conformable with the underlying Shahkot Formation and overlying Shekhai Formation (Yeats et al. 1987, Hylland et al. 1988). The Utch Khattak and Shahkot Formations have been assigned a tentative Late Precambrian age by Yeats et al. (1987), because they are conformable with the underlying Manki Formation.

Shekhai Formation: It is composed largely of limestone and marble interbedded with lesser amounts of quartzite and shale. The limestone varies in colour from grey, light brown to pink. Locally it is oolitic. In the Gandghar Range patches of white brecciated marble are associated with igneous dykes (Hylland et al. 1988). The Shekhai Formation has been assigned a tentative Late Precambrian age.

Dakhner Formation: In the Attock–Cherat Range, the central fault block consists mainly of argillite, sandstone and subordinate limestone assigned to the Dakhner Formation (Fig. 5.2). This Formation has a minimum thickness of 1,000 m. The sandstone and argillite are grey to greenish-grey, thin-bedded to massive and ripple marked. According to Hussain et al. (1990) there is a faulted lens within the Dakhner Formation that has yielded Cenomanian fossils. The Dakhner Formation is unconformably overlain by Jurassic and Cretaceous sequence. Hussain (1984) and Yeats et al. (1987) have tentatively assigned a Late Precambrian age to the Dakhner Formation owing to lack of apparent correlation with fossiliferous Phanerozoic sequences, absence of fossils and absence of bioturbation in fine grained strata. According to them, the Dakhner Formation could be stratigraphically equivalent to the Manki Formation, differing only in degree of metamorphism.

Swat–Hazara Crystalline Nappe and Thrust Belt

North and northeast of the Peshawar depression along the Indus Suture zone, there is a narrow belt of metasedimentary rocks perched on the northern margins of the Indo-Pakistan crustal plate. This belt extends from the Afghan border northeastward through Swat, Indus Kohistan, Chilas, round the Nanga Parbat–Haramosh Massif into the Astor Valley and the Kashmir Himalayas (Lawrence et al. 1989). These metasediments are largely crystalline schists and paragneisses, which occupy discontinuous stratigraphically-distinct wedges and thrust slices in a number of large-scale crustal nappes (see Chapter 4). Between Swat and Kaghan Valleys there are six major nappes, namely the Swat Nappe, the Besham Nappe, the Hazara Nappe, the Banna Nappe and the Lower and Upper Kaghan Nappes (Treloar et al. 1989a). Each of these nappes constitutes a distinct tectonostratigraphic unit. The basal part of the Swat, Besham, Hazara and Lower Kaghan Nappes consist of metasediments believed to be of Precambrian age. These Precambrian complexes are unconformably overlain by variably metamorphosed cover sediments (Table 5.1). At a few places, the metasediments have been intruded by Early Paleozoic granites such as the Mansehra Granite which has yielded a Rb/Sr date of 516 m.y. (Le Fort et al. 1980). ^{40}Ar/^{39}Ar dating of the metasedimentary sequence near Besham (Besham Group) has given dates of 2,000 ± 6 m.y., 1,950 ± 3 m.y., and 1,865 ± 3 m.y. to 1,887 ± 5 m.y. (Baig et al. 1989b). New ^{40}Ar/^{39}Ar dating of some of these basement rocks has indicated five pre-Himalayan metamorphic events which occurred at 2,000 ± 6 m.y., 1,950 ± 3 m.y., 1,865 ± 3 m.y. to 1,887 ± 5 m.y., 650 M 2 m.y. and 466 ± 2 m.y. (Baig et al. 1989b). The meta-sedimentary sequence of this tectonic belt is briefly described in the following paragraphs.

Manglaur Formation: In the Swat area the Manglaur Crystalline Schist, at its type locality near Manglaur village, was identified as a Precambrian unit by Kazmi et al. (1984c). It is exposed from near the Afghan border through Swat to the Besham area (Lawrence et al. 1989, Treloar et al. 1989a, 1989b, DiPietro et al. 1991). The Manglaur Schist constitutes the Swat Nappe and crops out as tectonised and, at places, mylonitised non-calcareous granoblastic quartz-mica-garnet schist, which has been feldspathised and tourmalinised. There are three main lithologic varieties, namely quartz-feldspar schist, quartz, mica-kyanite schist and quartz-mica-garnet schist. Thin layers of graphitic and hornblende schists (para-amphibolite) are present (Kazmi et al. 1984). The Manglaur Schist is overlain unconformably by the Alpurai Group. The age of the Manglaur Schist is uncertain. Because of its resemblance with the Tanawal Formation and correlation of the Mansehra Granite with the Swat Granite Gneiss, Kazmi et al. (1984) and subsequent workers (Lawrence et al. 1989, Treloar 1989c, Williams 1989) have tentatively assigned a Precambrian age to the Manglaur Schist. DiPietro (1990) has referred to these schists as Manglaur Formation.

Besham Group: Westward near Alpurai, the Swat Nappe structurally overlies the Besham Nappe which consists largely of sequences of the Precambrian Besham Group (Fletcher et al. 1986, Treloar et al. 1989b). This group includes both granitic, biotite rich orthogneisses as well as a sequence of metasedimentary gneisses, schists, metasammites, marbles and amphibolites (see Chapters 4 and 6). North of Besham and below the Main Mantle Thrust (MMT), there are several hundred metres of thick quartz feldspathic mylonites and blastomylonites (Lawrence et al. 1983). Largely undeformed tourmaline granites intrude the gneisses and mylonites (Williams 1989).

The original relationships between the ortho- and para-gneisses of the Besham Group are not clear because most of the contacts are sheared. According to Treloar et al. (1989a) it is likely that the orthogneisses formed a basement for the paragneisses. The Besham gneisses are overlain by, and imbricated with, the covering metasediments of the Karora Group. ^{40}Ar/^{39}Ar dating of the Besham Group amphibolites has given ages of 2,000 ± 6 m.y., 1,950 ± 3 m.y. and 1,865 ± m.y. to 1,887 ± 5 m.y. (Baig et al. 1989b). The age of the Besham Group is therefore defined as Late Archean (?) to Middle Proterozoic.

Baig (1990) has divided the Besham Group in two units. The **Thakot Formation**, which is the lower unit, is largely composed of quartzo-feldspathic gneisses, graphitic schists and gneisses and subordinate amounts of calcareous and pelitic schists. It grades into the overlying **Pazang Formation** which is comprised of calcareous schists, banded quartzite, tremolite-diopside-bearing marbles and metapyroxenitic komatiites.

Karora Group: The Besham Group is overlain by the Karora Group and the contact is along an angular unconformity (La Fortune 1988, Treloar et al. 1989a). At the base there is a conglomerate bed, the **Amlo Conglomerate** of Baig 1990, which is followed by graphitic and psammitic phyllite, intraformational conglomerate, calc-pelite, banded to massive quartzite, dolomite and marble. The rocks are deformed and metamorphosed to greenschist facies. Sodic leucogranites and pegmatites intrude the Karora Group, some of which have been dated as Early to Late Paleozoic and on this basis according to Baig et al. (1989b) the Karora Group is Middle to Late Proterozoic.

Pogue et al.(1995) and DiPietro et al. (1996) identify the lower part of the Karora Group as **Karora Formation** and the upper part as the **Gandaf Formation**. Their Karora Formation comprises thick, dark graphitic schist, marble, and subordinate amounts of non-schistose metapsammite, and metaconglomerate. It is intruded by small plutons. It differs from the Gandaf Formation largely due to the absence of garnetiferous rocks. The Gandaf Formation consists of dark schists overlain and interbedded with graphitic slate, phyllite, schist, and marble. The garnet schist forms conspicuous stratigraphic horizons. This Formation also is characterised by abundant intrusive rocks. Its lower contact with the Karora Formation is transitional and placed below a unit of light coloured metapsammite and garnet schist (DiPietro et al. 1996).

Tanawal Formation: East of Besham the Hazara Nappe crops out and is separated from Besham Nappe by the Thakot shear zone. The Hazara Nappe consists largely of metaquartzites, metapsammites and garnet-mica schists of the Tanawal Formation (Calkins et al. 1975, Treloar et al. 1989a). This Formation crops out extensively in the Hazara District and unconformably overlies the Hazara Formation of Precambrian age. It is overlain unconformably by the Abbottabad Formation of Cambrian age. It is intruded by the Late Cambrian Mansehra Granite (Le Fort et al. 1980) and is placed in the Late Proterozoic (Williams 1989, Treloar et al. 1989a).

Hazara Formation: First described as "Salt Series of Hazara" by Middlemiss (1896), this Formation consists of light-grey, partly phyllitic slates with some intercalations of sandstone, limestone, gypsum and graphite. Based on the presence of evaporites, the epimetamorphic Hazara Formation has been correlated with the non-metamorphic Salt Range Formation by Latif (1973). Radiometric age determinations carried out by Crawford and Davies (1975) on two samples from the Hazara Formation, however, yielded values of 765 ± 20 m.y. and 950 ± 20 m.y., respectively, showing that the Hazara Formation is probably older than the Salt Range Formation.

Kaghan Group: West of Balakot, the Balakot shear zone separates the Hazara Nappe from the Lower Kaghan Nappe. The Lower Kaghan Nappe consists of a series of metapelites, schists, quartzites, marbles, metaconglomerates and pegmatites which have been called the Kaghan Group (Ghazanfar and Chaudhry 1986). These rocks were originally loosely correlated with the Salkhala Formation of Precambrian age (Calkins et al. 1975, Ghazanfar and Chaudhry 1985).

Sharda Group: In the northeastern part of the Kaghan Valley, a northwest trending thrust fault, the Batal–Luat Fault (MCT?), truncates the Kaghan Group. North of this fault, higher grade metamorphic rocks such as garnetiferous calc-pelitic gneisses, pelitic gneisses, graphitic gneisses, pelitic gneisses with kyanite and sillimanite, and marbles with sheet granites and amphibolites (Table 5.2) form open elongated basins and domes. These rocks have been named Sharda Group by Ghazanfar and Chaudhry 1986 and they consider it to be Middle Proterozoic to Archaean in age (Chaudhry and Ghazanfar 1987). Previously, in this region, this unit was referred to as the Salkhala Formation.

Table 5.2. Stratigraphic sequence in the Upper Kaghan region

	Chaudhry and Ghazanfar (1987)	**Greco et al. (1989, 1993) ****
↑ Sharda Group ↓	Burawai Gneisses* (Schists, marbles, paragneisses)	Burawai Formation (Paleomesozoic metasedimentary sequence).
	Dadar Granite Migmatite Complex Saiful Muluk Granite Gneiss	Amphibolites (=Permian Panjal Trap)
	Lulu Sar Gneisses*	Naran Formation (Metapelitic-metagreywacke gneisses =Lower Paleozoic basis of cover)
	Besal di Khari Granite Gneiss Besal Gneisses*	? – Granitic basement (Cambrian and older).

(*include amphibolites). ** This sequence has been assigned a much younger age by Greco et al; see section under "Carboniferous–Permian".

Salkhala Formation: In the Himalayan crystalline zone between Kashmir and Hazara area, metamorphic sediments of this Formation ("Salkhala Series" of Wadia 1930) are widely distributed. Type locality is the village of Salkhala on the Kishanganga River in Kashmir. The Formation is made up of slates, phyllites, schists and intercalated marble and quartzite beds as well as graphite layers. At the type locality in Kashmir, the Salkhala Formation is overlain by the Dogra Slate which is correlated with the Hazara Formation of Hazara area. Hence, also the Salkhala Formation is considered as Proterozoic.

Nanga Parbat Group: In the Nanga Parbat–Haramosh Massif (Gilgit District), a sequence of crystalline schist and gneiss has been referred to as the Nanga Parbat Group (Madin 1986). The Group consists of three lithostratigraphic units- the Shengus Gneiss, Iskere Gneiss and the Haramosh Schist. The **Shingus Gneiss** is structurally the lowest and includes fine-grained laminated, amphibolite-grade pelitic and psammitic paragneiss with subordinate amphibolite, and calc-silicate gneiss. It is at least 5 km thick. Its contact with the Iskere Gneiss is sheared and marked by sillimanite bearing mylonites. Treloar et al. (1991) consider the Shengus Gneiss as cover to the Iskere Gneiss and correlate it with the Tanawal Formation–Mansehra Granite sequence of Hazara region. Zeitler et al. (1989) have shown that Shengus Gneiss is interlayered with 500 Ma orthogneisses. The **Iskere Gneiss** is dominantly amphibolite-grade, coarse-grained, biotite gneiss, with subordinate calc-silicate gneiss, amphibolite and biotite schist. It has a minimum thickness of 8 km. The gneiss is a plutonic rock intruded into a sequence of metasediments. It is overlain by the **Haramosh Schist** which consists of amphibolite grade biotite schist and gneiss with marble, calc-silicate gneiss and amphibolite. The Iskere Gneiss has given a U/Pb isochron date of 1,800 m.y. and is considered to be Archean or Early Proterozoic (see Chapter 6).

Karakoram

Chikar Formation: A metasedimentary sequence intruded by pre-Ordovician Ishkarwaz Granite, is exposed in the upper Yarkhun Valley, south of Baroghil Pass in Chitral District (Le Fort et al. 1994). It probably represents a section of the crystalline basement of the Karakoram microplate and is similar to the Precambrian substratum of central Afghanistan. It is largely comprised of dark-coloured siltstones and quartzites, metamorphosed into hard spotted schists and massive hornfels. The metasediments are intruded by granite dykes, which increase in density towards the east, where the sequence grades into migmatitic gneisses and anatectic granite (Tongiorgi et al. 1994). Le Fort et al. (1994), who first recognised this sequence, have referred to it as pre-Ordovician meta-sedimentary rocks and migmatites, without giving it a name. Because it is a significant unit, we propose that it may be named Chikar Formation (after Chikar village).

PALEOZOIC

Paleozoic sedimentary rocks occur at several localities in Pakistan, but a complete Paleozoic sequence is lacking. Paleozoic rocks have been reported from northeast Balochistan, Khisor Range, Salt Range, Peshawar Basin, Hazara–Kashmir Syntaxial region, Swat, Besham, Hazara, Kaghan, Nanga Parbat–Haramosh Range and the Karakoram. The Paleozoic sequence is, however, best developed in the Peshawar Basin. A correlation of the Paleozoic sequences is shown in Table 5.1.

Cambrian

Salt Range

Cambrian sediments are well-exposed in Salt Range and Khisor Range; they have also been identified in Hazara area. In the Salt Range, the Cambrian sequences have been subdivided into four formations: Khewra Sandstone, Kussak Formation, Jutana Formation and Baghanwala Formation.

Khewra Sandstone: Previously known as the "Purple Sandstone Series" (Wynne 1878), the Khewra Sandstone (Shah 1977) overlies the Late Proterozoic Salt Range Formation without any apparent disconformity. The type locality is the Khewra Gorge in eastern Salt Range. The Khewra Sandstone is widely exposed in the Salt Range, and in the Khisor Range. It was also penetrated by drilling in the southern and eastern Kohat–Potwar foredeep, as well as in the Punjab Plains south of the Salt Range and east of the Sulaiman foredeep. The Khewra Formation consists mainly of reddish-brown to purple, thick-bedded to massive sandstones with few brown shale intercalations. The sandstone is characteristically cross-bedded, has abundant ripple marks and mud cracks, and, in places, exhibits convolute bedding. The thickness of the Khewra Sandstone is 150 m at the type locality in eastern Salt Range, and 80 m in the Khisor Range.

Apart from rare trace fossils interpreted as trilobite trails by Schindewolf and Seilacher (1955), the Formation is devoid of fossils. Because of its position between the Late Proterozoic Salt Range Formation and the fossiliferous Early (to Middle ?) Cambrian Kussak Formation, the Khewra Sandstone is thought to represent the basal part of the Lower Cambrian.

Kussak Formation: First described as "Obolus Beds" by Wynne (1878) and as "Neobolus Beds" by Wynne (1895), the Kussak Formation rests disconformably upon the Khewra Sandstone, marked by a widespread, thin conglomerate developed at the base of the Kussak Formation. Type locality is near Fort Kussak in the eastern Salt Range. The Kussak Formation is well-exposed in the Salt Range, between Jogi Tilla in the east and Chhidru in the west, as well as in the Khisor Range. It has also been encountered by wells drilled in southeastern Potwar area and in the Punjab Plains. The Formation consists mainly of grey, silty and sandy, glauconitic shales with some sandstone intercalations and few black shale layers. The thickness of the formation is 75 m at the type locality, 30 m in the Khisor Range, and locally more than 200 m in the southern Punjab Plains. The Kussak Formation contains brachiopods and trilobites, among them *Neobolus warthi* and *Redlichia noetlingi*, pointing to an Early Cambrian (Schindewolf and Seilacher 1955), or early Middle Cambrian age (Teichert 1964), respectively. The fauna, together with the glauconite content, indicates a marine depositional environment.

Jutana Formation: Previously named "Magnesian sandstone beds" by Fleming (1853) and "Jutana stage" by Noetling (1894), respectively, the Jutana Formation (Shah 1977) overlies the Kussak Formation conformably. Type locality is near the village Jutana in eastern Salt Range. The Formation has a distribution similar to the Kussak Formation in the Salt and Khisor Ranges, in the eastern and southern Potwar Basin, in the Mianwali Re-entrant and in the southern Punjab Plains. It is composed of cliff-forming, thick-bedded to massive, brownish-weathering, sandy dolomites and dolomitic sandstones with few shale intercalations. The thickness is 75 m–90 m in the eastern Salt Range, and 45 m in the

Khisor Range. The Formation has yielded some brachiopods, gastropods and trilobites, among them *Redlichia noetlingi* and *Pseudotheca subrugosa* (Teichert 1964), as well as *Cruziana sp.* The fossils indicate a late Early Cambrian to early Middle Cambrian age.

Baghanwala Formation: This Formation rests conformably on the Jutana Formation. It was first described as "Pseudomorphic Salt Crystal Zone" by Wynne (1878), and later as "Salt Pseudomorph Beds" by Holland (1926). Type locality is near Baghanwala Village in the eastern Salt Range. The distribution of the Formation in surface outcrops is practically identical with that of the Kussak and Jutana Formations. In the Potwar area however, only few erosional remnants of the Baghanwala Formation have been preserved below the Permian unconformity. The Formation is also present in the subsurface of the southern Punjab Plains. The Baghanwala Formation consists mainly of reddish-brown shales and platy to flaggy sandstones characterised by an abundance of salt pseudomorphs. Ripple marks and mud cracks are common. The thickness of the Formation is about 100 m at the type locality, but is commonly reduced by erosion in other parts of the Salt Range. Apart from some trace fossils, the Formation is largely devoid of faunal relics. Because of its conformable contact with the underlying Jutana Formation, it has tentatively been assigned to the Middle Cambrian. The sediments of the Formation were deposited in a lagoonal environment under arid conditions.

Khisor Formation: In the southern Khisor Range, the Jutana Formation is overlain conformably by the Khisor Formation (Shah 1977). It was named previously "Gypsiferous Series" by Gee (1945). Type locality is west of Saiyeduwala. The prevailing lithology is white to light-grey, partly dark-grey and laminated gypsum with dolomite layers and intercalations of dark-grey to black, silty, partly dolomitic shale. The thickness ranges from 100 m to 135 m. The Khisor Formation is barren of fossils. It is interpreted as an evaporitic time-equivalent of the Middle (?) Cambrian Baghanwala Formation.

Lower Hazara

In the southern Himalayan foothills of Hazara, a different Paleozoic sequence is exposed. It consists of the Abbottabad and Hazira Formations, to which a Cambrian age has been assigned (Table 5.3).

Abbottabad Formation: This Formation overlies the Precambrian Tanawal Formation unconformably. It was described by previous authors as "below the Trias" (Waagen and Wynne 1872) and "Infra-Trias" (Middlemiss 1896), respectively. Type locality is the Sirban Hill near Abbottabad. The Formation extends from the Tarbela area (Indus River) in the west through Abbottabad to Muzaffarbad and Balakot in the east and northeast. In the Abbottabad area, Latif (1970a, 1970b, 1974) described the Abbottabad Formation, for which he suggested the name "Abbottabad Group", as mainly consisting of thick-bedded sandstones, with some shale, siltstone and dolomite intercalations in its lower part ("**Kakul Formation**"). The basal unconformity is marked by a conglomerate ("**Tanakki Conglomerate**"). The upper part of the Abbottabad Group as defined by Latif is formed by a thick dolomite/limestone/marble sequence with few quartzitic sandstone layers ("**Sirban Formation**"). The marbles contain phosphorite which is being mined near Abbottabad (see Chapter 9). The lithology of the Abbottabad Formation shows great lateral variations. Calkins et al. (1968) reported the presence of phyllites as well as dolomite, quartzite and

Table 5.3. Precambrian and Cambrian sedimentary sequence in southern Hazara. Table shows lithostratigraphic units proposed by various authors.

Middle Miss 1896		Marks & Mohammad Ali 1962		Gardezi & Ghazanfar 1965		Latif 1974				Butt 1989	
	Volcanic Material Etc.		Triassic Rocks ?		Hazira Formation / Hematite Formation	Triassic ?	Tarnawi Formation		Hazira Member Galdanian Member		
Infra-Triassic Series	Upper Limestone	Abbottabad Formation	Unconformity Upper Dolomite Member	Abbottabad Group	Abbottabad Formation	Cambrian	Abbottabad Group	Sirban Formation		Shekhan Bandi Formation Hazira Formation	
								Kakul Formation	Mirpur Member	Abbottabad Formation	Mirpur Member
	Lower Sandstone and Shale		Upper Shale and Sandstone Member						Mahmdagali Member		Mahmdagali Member
			Lower Dolomite Member						Sangargali Member		Sangargali Member
			Lower Shale and Sandstone Member						Tannaki Member		Tannaki Memember
	Basal Conglomerate		Basal Conglomerate Member								
					Unconformity				Unconformity		Unconformity
	Tanol Series		Tanol Formation			Precambrian	Tanol Formation				Hazara Formation
	Slate Series		Hazara Slate Formation				Hazara Fm.		Upper Formation Langrial Limestone Middle Formation Miaranjani Lst. Lower Formation		

conglomerate. The thickness of the Formation is about 660 m at the type locality and 900 m elsewhere. The Formation is sparsely fossiliferous. *Hyolithes* spp. and *Hyolithellus* spp. indicate an (Early?) Cambrian age (Shah 1977).

Hazira Formation: This Formation overlies the Abbottabad Formation with an apparently conformable contact (Gardezi and Ghazanfar 1965). Type locality is the village of Hazira and the main area of distribution is southern Hazara. The Lower part of the Formation, also described as "**Haematite Formation**" or "**Galdanian Formation**" (Latif 1970a), is characteristically developed near Galdanian northeast of Abbottabad. It consists of purple quartzitic siltstones with some manganese oxide layers, followed by an alternation of nodular haematitic shale and sandstone, as well as quartz breccias. Locally, volcanic rocks have been observed within this part of the sequence. The upper part of Hazira Formation is mainly composed of grey and yellowish-brown, calcareous siltstone and sandstone with earthy concretions, and some quartzite beds. The basal portion is characteristically glauconitic, whereas the middle part is phosphatic (Latif 1970a). The maximum thickness of the Formation is 300 m. Fuchs and Mostler (1972) described a small fauna of Porifera, Hyolithids and Annelida indicative of an Early (?) Cambrian age. Rushton (1973) re-examined the Hyolithids and sponge spicules found by Latif in the glauconitic part of Hazira Formation and confirmed the Cambrian age.

Peshawar Basin and Attock–Cherat Ranges

Amber Formation: Cambrian rocks might also be present within the Paleozoic complex near Nowshera, Peshawar District. Pogue and Hussain (1986) and Hussain et al. (1989)

modified and revised the stratigraphy of the Nowshera area and suggested the name "Amber dolomite" for the oldest unit of this complex. On the basis of similar lithology they correlated it with the Cambrian Abbottabad Formation of Hazara. Hussain et al. (1990) have now referred to it as Amber Formation. This unit is mainly composed of brownish-grey dolomite with chert lenses and some shale layers. The upper part is more massive and contains stromatolitic laminations. Dolomite pebbles are in the basal conglomerate of the Ordovician Misri Banda Quartzite which overlies the Amber Formation unconformably. The thickness of this Formation is more than 425 m (base not exposed).

Darwaza Formation: In the Attock Range, according to Hussain et al. (1990) the oldest formation exposed on its southern slopes, consists of cream-coloured limestone with dolomite overlain by maroon argillite. It has been named Darwaza Formation (Hussain et al. 1990). Its lower contact is not exposed. Its upper contact is gradational with the overlying Hisartang Formation (Hussain et al. 1990). The Darwaza Formation is un-fossiliferous and has been tentatively correlated with the Amber Formation (Hussain et al. 1990).

Ordovician–Silurian

Ordovician to Silurian sequences are largely confined to the Peshawar Basin and the northern sedimentary belt of the Karakoram (Table 5.1, Figs. 4.21 and 4.40).

Peshawar Basin

Ordovician rocks have been identified in the Paleozoic complex of the Nowshera area. In this region the Ordovician-Silurian sequence is comprised of Misri Banda Quartzite, Hisartang Formation and Panjpir Formations. **Misri Banda Quartzite** (Stauffer 1968a, Pogue and Hussain 1986, Hussain et al. 1989, 1990) contains the trace fossil *Cruziana rugosa* indicating an Early to Middle Ordovician age. The Formation rests unconformably upon the Cambrian (?) Amber Formation. It consists chiefly of grey to pinkish, cross-bedded, feldspathic quartzite with subordinate argillite. The quartzite is cross-bedded, displays ripple marks and worm burrows. The base is locally formed by a layer of conglomerate. The upper third of the sequence contains limestone layers with crinoid fragments, as well as thin beds of dark-grey to black phyllite. The Formation is 175 m thick.

Hisartang Formation: In the Attock–Cherat Range a series of unfossiliferous quartzites, with argillite beds in the middle, has been named Hisartang Formation by Hussain et al. (1989, 1990). The quartzite is white to grey and fine-grained and contains impressions of worm burrows. The argillite is dark-grey to black and laminated. This Formation has a conformable gradational contact with the underlying Darwaza Formation and is overlain conformably by the (Devonian ?) Inzari Limestone. Owing to the lithologic similarities the Hisartang Formation is correlated with the Early to Middle Ordovician Misri Banda Quartzite. Its minimum thickness is 650 m.

Panjpir Formation: The Misri Banda Quartzite is overlain unconformably by the Panjpir Formation (Pogue and Hussain 1986, Hussain et al. 1989, 1990). The type locality is the Panjpir village in Mardan District. At the base there is a conglomerate with clasts of quartzite dolomite, limestone and argillite. The Formation is largely comprised of slate and phyllite with intercalations of limestone and quartzite. It is up to 1,075 m thick. The limestone in the lower part contains crinoids, pelecypods, cephalopods and Ludlovian conodonts (Hussain et al. 1990). Nautiloid fragments have been found in limestones in the upper part of Panjpir Formation. A crinoidal limestone at the top has yielded conodonts of Late Silurian (Pridolian) age (Talent and Mawson 1979). The age of the Formation has been interpreted as Middle Ordovician (?) to Late Silurian by Pogue and Hussain (1980).

Landikotal Formation: In Khyber Agency, near the Afghan border, a thick series of phyllites and slates with subordinate limestone, dolomite and quartzite, intruded by basic dykes and sills, is exposed for which Stauffer (1968b) introduced the name "Landikotal Formation". Outcrops of the Formation extend from Jamrud across the border into Afghanistan. Type locality is Landi Kotal (Khyber Pass). Because of the structural complexity, the thickness of 3,300 m quoted by Stauffer appears doubtful; the figure of 1,600 m cited by Shah et al. (1970) might be more realistic. The Landikotal Formation has not yielded any fossils. Part of the Formation has been correlated with the "Lower Formation" of Afghanistan which contains conodonts (Shah et al. 1970) indicative of a Late Ordovician to Late Silurian (Middle Ludlovian) age. The upper part of Landikotal Formation might correlate with the upper part of the Panjpir Formation of Nowshera.

Karakoram

Baroghil Formation: Ordovician sedimentary rocks occur in the upper Yarkhun Valley, south of Baroghil Pass, in Chitral District and have been named Baroghil Formation by Tahirkheli (1982). Le Fort et al. (1994) have referred to the Lower Ordovician to Permian sequence in this region as Baroghil sediments whereas Gaetani et al. (1995) have used the name Yarkhun Formation for the Lower Ordovician to Silurian sequence of this region. According to the National Stratigraphic Code, Tahirkheli's Baroghil Formation takes precedence and we therefore retain this name. This Formation is comprised of siliceous slates, flaggy quartzites, thin bedded dolomitic limestones and siliceous dolomites. This sequence unconformably overlies the Chikar Formation and the Ishkarwaz Granite (Le Fort et al. 1994). A one metre thick conglomerate with grey-green chert clasts occurs at the base, followed by slates, ferruginous subarkoses interbedded with slates, dolomites with Middle to Late Ordovician conodonts (Talent et al. 1981), crinoid and bryozoan fragments. The shaly beds of this sequence have yielded abundant Early Ordovician (Arenigian) acritarchs. This palynological assemblage is divided into two groups, one below and the other above the ferruginous subarkosic zone. The lower group has yielded 34 species, though the more characteristic ones are *Arbusculidium filamentosum, coryphidium bohemicum, C. minutum, C.miladae, Striatotheca principalis parva* and *Striatotheca rarirrugulate,* (Tongiorgi et al. 1994, Le Fort et al. 1994). The second assemblage consists of 22 species though it may be characterised by the common occurrence of *Striatotheca microrugulate, Tongzia* sp, *Vogtlandia hoffmanensis* and *Peteinosphaeridium* sp.

The conodont fauna from this sequence comprises *Rhynchognathodus divicate, Drepanodus homocurvatus, D. Suberectus (?), Amorphognathus* sp, *Cordylodus* sp, *Dichognathus* sp, *Oistodus* sp and *Teraprionodus* sp (Talent et al. 1979).

The Baroghil palynoflora belongs to the cold-water peri-Gondwana acritarch bioprovince fringing the northern Gondwana, which is also referred to as the Mediterranean Province. According to Tongiorgi et al. (1994) and LeFort et al. (1994), biogeographical and geological comparisons confirm that before accretion to the Eurasian continent, the Karakoram microplate was located along the northern margin of Gondwanaland. According to Tahirkheli (1982) an angular unconformity marks the upper boundary of the Baroghil

Formation and it is unconformably overlain by the Devonian sequence (Fig. 5.3). Talent et al. (1982), however, consider this discordance to be a fault contact.

Devonian

Devonian sequences have been identified in the Peshawar Basin and in the Chitral region of the Karakoram (Table 5.1, Fig. 4.21 and 4.40).

Peshawar Basin

Nowshera Formation: At Nowshera, the Late Silurian Panjpir Formation is conformably overlain by the Nowshera Formation (Stauffer 1968, redefined by Pogue and Hussain 1986, Hussain et al. 1989, 1990). The type locality is 3.5 km north of Nowshera. The outcrops of the Formation near Nowshera have been described as reef complex by Teichert and Stauffer (1965). The core of the reef is formed by limestone rich in corals, stromatoporoids, cephalopods, gastropods and brachiopods, and is surrounded by limestones containing brecciated reef material, and sparsely fossiliferous dolomites. The Nowshera Formation as defined by Pogue and Hussain (1986) includes a lower fossiliferous limestone/dolomite unit (**Nowshera Formation** of Stauffer 1968a), a middle unit of carbonate-cemented sandstone (**Misri Banda Quartzite** of Stauffer), and an upper limestone/dolomite unit (**Pir Sabak Formation** of Shah 1977), doubtfully dated as Early Carboniferous by Latif (1970a). The total thickness of the Nowshera Formation is quoted as 595 m by Pogue and Hussain (1986). Conodont studies by Molloy (in Talent and Mawson 1979) indicate an Early Devonian age of the Nowshera Formation. Corals found in the upper part (Pir Sabak Formation) are characteristic and indicate a pre-Late Devonian age (Shah et al. 1970).

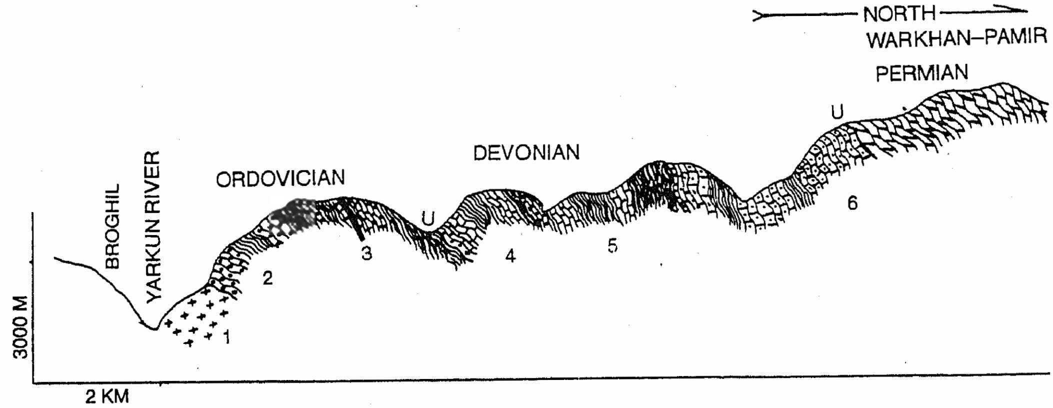


Fig. 5.3. Geological section across Yarkhun River, south of Baroghil Pass. 1. Hornblende granite (Ishkarwaz), 2. Flaggy quarzites, slates, thin dolomitic limestones (Yarkhun Fm), 3. Siliceous dolomitic bed with Ordovician conodonts, 4. Angular unconformity, 5. Slates, dolomitic limestone, quartzite (Chilmarabad Fm ?), 6. Dolomitic limestone with slaty partings (Permian ?), U. Unconformity (modified from Tahirkheli 1982).

Khyber Carbonate Complex: In the Khyber Agency, the Ordovician/Silurian (?) Landikotal Formation is overlain, with a faulted contact, by a series of shales, quartzites, limestones and dolomites described by Shah et al. (1970) as "Khyber undifferentiated carbonate complex". The series extends from Ghundi Sar in the south to Shahid Mena in the north. It is highly deformed, particularly in the northern part, and it is dissected by basic intrusions. Some of the limestone beds are fossiliferous and contain bryozoans, and crinoid remains. Shah et al. (1970) correlated them with the Nowshera Formation and, hence, assigned an early Devonian age to the Khyber undifferentiated carbonate complex.

Shagai Limestone: The Khyber carbonate complex is in tectonic contact with the Shagai Limestone (Stauffer 1968b) which is exposed between the eastern end of Khyber Pass, Shagai Fort, Ali Masjid and Tauda Mela. Type locality is 1 km northeast of Shagai. The unit consists of grey to black, partly brown limestone which generally is thin to medium-bedded in the lower part and thick-bedded to massive in the upper part. The limestone is locally dolomitic and highly fractured. The top of the unit is formed by thinly laminated limestone and sheared shale. Apart from rare, unidentifiable brachiopod remains, the Shagai Limestone is unfossiliferous. In view of its conformable contact with the overlying Late Devonian Ali Masjid Formation, it is considered as Silurian (?) to Early Devonian (Shah et al. 1970).

Ali Masjid Formation (Stauffer 1968b): This Formation extends all over the Khyber Agency. Type locality is the village of Ali Masjid (Khyber Pass). The Formation is characteristically composed of red-coloured shale, alternating with siltstone, sandstone, quartzite and limestone. Repeated breaks in sedimentation are indicated by thin conglomerates and lateritic beds. The total thickness of the Formation is 220 m at the type locality. West of the type section, some highly fossiliferous beds were found by Shah (1969) which yielded a brachiopod and coral fauna indicative of a Late Devonian age. Shah (1977) correlated the Ali Masjid Formation with the Haji Gak Formation of Afghanistan.

Inzari Limestone (Hussain et al. 1990): This Limestone is thin-bedded, yellowish to greenish-grey and crystalline. Dedrites and stylolites are common. It conformably overlies the Hisartang Formation. Its upper contact is faulted. No fossils have been found. Hussain et al. (1990) considered it as time-equivalent with the Early Devonian Nowshera Formation.

Swat

Jobra Formation: In the Swat area, about 5 km southeast of Ilam, discontinuous lenses of calc-silicate-bearing marble unconformably rest on the Swat Granite Gneiss and the marble is unconformably overlain by Alpurai Group. This marble unit has been named Jobra Formation by Dipietro (1990); its age is as yet not known. From its superposition between the Swat Granite Gneiss and the Alpurai Group it may be inferred to be Paleozoic. On the basis of its stratigraphic position and its calc-silicate mineralogy, Dipietro (1990) correlates it with the Nowshera Formation.

Karakoram

Devonian sedimentary rocks crop out extensively in the western region of the Karakoram. However the stratigraphy of this region is largely based on scattered reconnaissance mapping and remains incoherent and confused, compounded by the fact that various authors have coined different formation names for overlapping lithostratigraphic units.

Talent et al. (1979) collected several Devonian fossils from a number of localities in Chitral which include graptolites, (*Monograptus* cf *uniformis*), brachiopods (e.g., *Cyrtina sp, Dalejina hybrida, Leptostrophia sp, Nucleospira sp, Stegerhynchus*), Corals (species of *Rhizocorallium, Favosites, Fenestella*) and Conodonts (including *Belodella resima, Icriodus huddlei, Neopriomodus excavatus, Ozarkodina denckmanni, Polygnathus lenzi, Spathognathodus steinhornensis*). This fauna is largely Devonian. Middle to Late Devonian fossils (corals and conodonts) have been collected from Kuragh Spur and Shogram (Reed 1922, Schouppé 1965, Sartenaer 1965, Vandercammen 1965, Gaetani 1967, Talent et al. 1979). The Devonian fossils from Chitral have been described without tangible description of lithostratigraphic sections or units and therefore presently they are not very helpful in sorting out the stratigraphic correlation problem in Chitral, particularly lateral changes in lithofacies, and correlation of the lithostratigraphic units overlying and underlying the Shogram Formation (Table 5.1). In this section we summarise the presently available information.

Charun Quartzite: The Charun Quartzite (Stauffer 1975) occurs at the base of the Devonian succession and is comprised of white, medium-grained quartzites which are up to 100 m thick. Talent et al. (1979) have reported Early Devonian fossils from this unit. According to Tahirkheli (1982) the Charun Quartzite is unconformably overlain by the remaining Devonian sequence.

Chilmarabad Formation: A Devonian sedimentary sequence dominated by slates, dolomitic limestone and dolomites unconformably overlies the Baroghil Formation (Figs. 5.3 and 5.4) in the Baroghil Pass region (Tahirkheli 1982) and has been referred to as the Chilmarabad Formation by Le Fort et al. (1994) and Gaetani et al. (1995). It is overlain by the Late Devonian Shogram Formation (Gaetani et al. 1995).

Shogram Formation: This Formation is about 800 m thick near the type locality (Shogram village) and consists of a massive crinoidal dolomite in the lower part, overlain by bedded dolomites, fossiliferous limestone (with corals) and thin shales in the middle, and massive cross-bedded, fine to medium-grained quartzites and limestone with brachiopods in the upper part (Desio 1966, Tahirkheli 1982). Calkins et al. (1981) refer to the Shogram Formation as "Devonian Unit" and include in it the Charun Quartzite of Stauffer (1975). According to them the Devonian sequence is exposed in an inverted anticline, thrust southeastward over the Cretaceous Reshun Formation.

It appears that the Shogram Formation goes through rapid lateral and vertical facies changes, a feature which may have caused much of the prevailing confusion in local stratigraphy. Only a few kilometres southwest of Shogram, above the Reshun Thrust, and in the same tectonostratigraphic position as the Shogram Formation, Leake et al. (1989) have described and mapped a sequence of green schists, micritic limestone, cherts, dolomite, carbonate phyllites, sandstones and breccias, named **Sewakht Formation**. They correlated it with Shogram Formation. In their geological map of the region southwest of Shogram Calkins et al. (1981) apparently include the Sewakht Formation of Leake and the Shogram Formation in their Sarikole Shale.

Reed (1922) recorded several Devonian fossils from the type locality of Shogram. Calkins et al. (1981) collected Middle Devonian to Late Devonian rugose corals near Pasti. The Shogram Formation is therefore Devonian in age. According to Tahirkheli (1982) it has a faulted contact with the overlying shales, phyllites and quartzites (Sarikole Shales of Hayden 1915; and Lun Shales of Desio 1966). Pudsey et al. (1985) mention that this contact is faulted at some places and conformable at others. According to Calkins et al. (1981), Tahirkheli (1982) and Pudsey (1985), the Sarikole/Lun Shales range in age from Devonian to Permian or even up to Jurassic (mainly due to reported occurrence of Devonian fossils in shaly sequence adjacent to the dolomite and quartzite of the Shogram Formation). Desio (1975), Stauffer (1975) and Calkins et al. (1981), however, consider the Sarikole/Lun

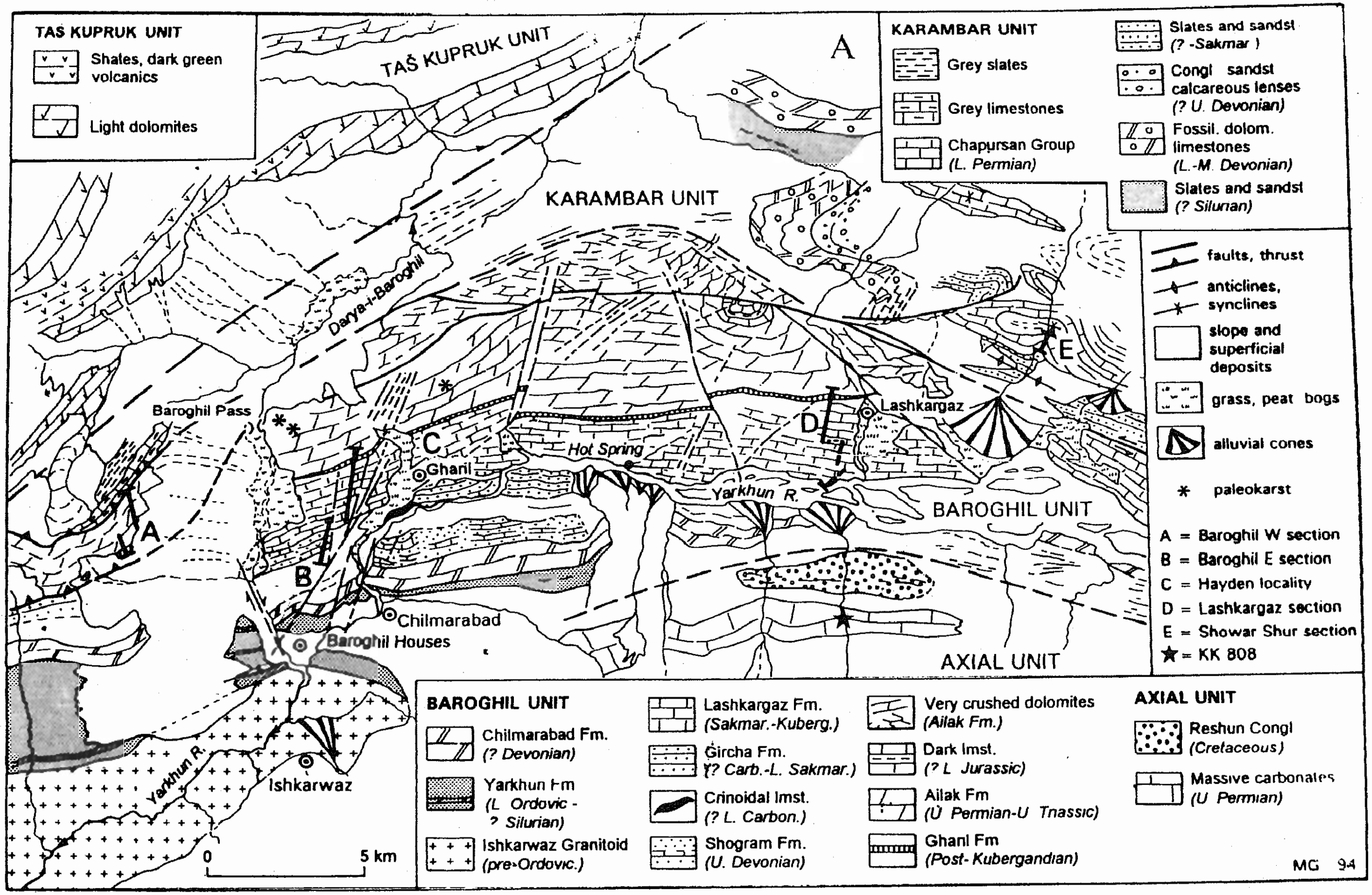


Fig. 5.4-A. Geological map of Upper Yarkhun Valley and Baroghil Pass (from Gaetani et al. 1995).

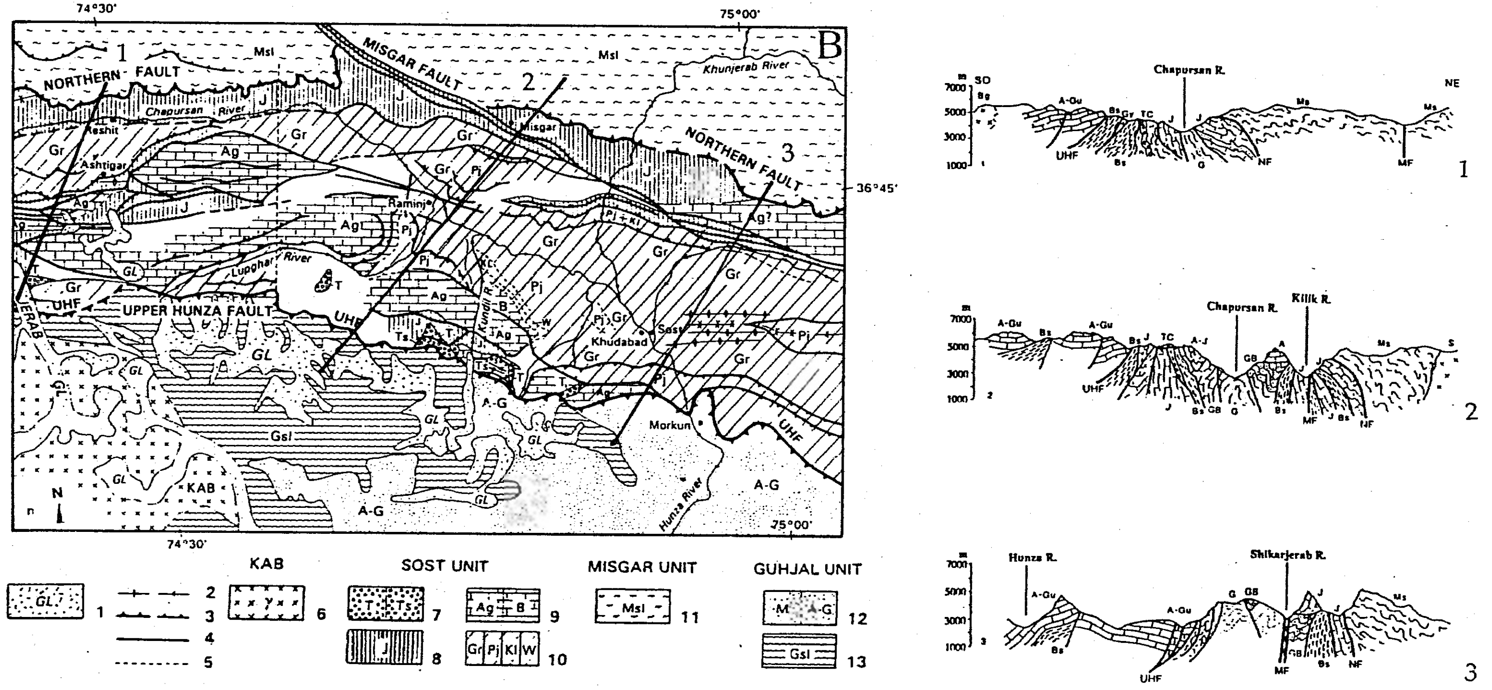


Fig. 5.4–B. Geological map of Chapursan region of the Northern sedimentary belt of the Karakoram (from Zanchi 1993).
(1) glaciers; (2) fold axes; (3) major and minor thrust planes; (4) strike-slip faults; (5) stratigraphic contacts: (6) Karakoram Axial Batholith; (7) Cretaceous: Tupop Fm. (T), Late Cretaceous marls (Ts); (8) Jurassic sediments undifferentiated (J);
(9) Triassic: Aghil Fm. (Ag), Borom Fm. (B); (10) Permian: Gircha Fm. (Gr), Panjshah Fm. (Pj), Kundil Fm. (Kl), Wirokhun Fm. (W); (11) Misgar Slates (Msl); (12) marbles (M), Aghil-Guhjal dolomitic limestones (A-G); (13) black slates of the Guhjal Unit (Gsl).
1, 2, 3. Cross sections through the Sost Unit,Chapursan Valley. A–Gu, Aghil Formation and Guhjal Dolomites; Bg, Batura granitoids; Black slates of the Sost Unit, mostly Mesozoic; G, Gircha Formation; GB, Gircha to Borom Formation sequence; Gy, Gypsum; J, Jurassic; MF, Misgar Fault; Ms, Misgar Slates; NF, Northern Fault; S, Gyraf "Syenite"; TC, Tupop Conglomerate; UHF, Upper Hunza Fault. Vertical and Horizontal scales are equal.

Shales to be Carboniferous to Permian in age.

Carboniferous–Permian

Carboniferous to Permian sedimentary rocks have been reported from northeastern Balochistan, Khisor Range, Salt Range, Peshawar Basin, Hazara–Kashmir Syntaxis, Swat, Hazara, Kaghan and the Karakoram. The Carboniferous sequence is however missing at most localities except Nowshera, Swat and the Karakoram (Table 5.1).

Northern Balochistan

Permian sedimentary rocks have been reported from parts of Sulaiman and Kirthar Ranges (Vredenburg 1904, 1909a, Williams 1959, HSC 1960, Sokolov and Shah 1965). Near Wulgai, Ghazaband and Kalat, dark-grey to black, crystalline limestone, partly associated with slate, have yielded brachiopods such as *Athyris, Dielasma, Marginifera* and *Notothyris,* as well as fusulinids, crinoids, stromatopores and algae (Shah 1977). As discussed in Chapter 4, the Permian fossils actually occur in large blocks of olistoliths in a Mesozoic sedimentary melange. We suspect that other occurrences of Permian rocks may be of a similar nature.

Salt Range and Khisor–Marwat Ranges

The Salt Range in northern Punjab is one of the classical areas in the world where a practically complete Permian sequence is developed and perfectly exposed (Table 5.1). The Permian is , here, subdivided into two Groups: the predominantly clastic **Nilawahan Group** and the mainly calcareous **Zaluch Group**. The boundary between the Upper Permian and the Lower Triassic has been the subject of intensive studies (Kummel and Teichert 1966, 1970).

Nilawahan Group: It consists of four formations: Tobra Formation, Dandot Formation, Warcha Sandstone, and Sardhai Formation.

The **Tobra Formation** (Shah 1977) was formerly correlated with the "Talchir Boulder Beds" (Gee, in Pascoe 1959) and later described as "Salt Range Boulder Beds" by Teichert (1967). The Formation rests unconformably upon different Cambrian formations or the Salt Range Formation, respectively. Type locality is the village of Tobra, north of Khewra, in the eastern Salt Range. The Formation is exposed throughout the Salt Range as well as in the Khisor Range. It was encountered by wells drilled in the Kohat–Potwar area, Mianwali Re-entrant and Marwat Range, as well as along the shallow eastern flank of Sulaiman foredeep. In the eastern Salt Range, the Tobra Formation consists mainly of polymict conglomerates with pebbles and boulders of igneous, metamorphic and sedimentary rocks. Based on the observation of polished and scratched boulders, Teichert (1967) interpreted the Tobra conglomerate as a tillite. The thickness of the Formation is 20 m at the type locality. In the western Salt Range, the Tobra Formation is developed in a distinctly different facies. Dark-grey to black, diamictic mudstones with interspersed clasts of sand-grain size to boulder size prevail. The facies is interpreted as fluvioglacial. The maximum thickness (120 m–130 m) was observed in Zaluch Nala. Thin coal beds are locally developed at the top of Tobra Formation, few metres below the base of Warchha Sandstone. The Tobra Formation contains ostracods and fresh-water bivalves (Reed 1936), as well as floral remains including *Glossopteris* and *Gangamopteris*. On the basis of *Striatopodocarpites* and *Protohaploxypinus,* Balme (in Teichert 1967) assigned an Early Permian age to the Formation.

The Tobra Formation is overlain conformably by the **Dandot Formation** (Shah 1977), previously known as "Olive Series", "Eurydesma beds" and "Conularia beds" (Wynne 1878), lower part of "Speckled sandstone" (Waagen 1879), or "Dandot Group" (Noetling 1901). Type locality is the village of Dandot, north of Khewra, in the eastern Salt Range. The Formation is well represented in the eastern and central Salt Range, but is not developed in the western Salt Range and Khisor Range.

In eastern Salt Range, the Dandot Formation mainly consists of dark greenish-grey, splintery shale and siltstone with intercalated sandstone, whereas in the central Salt Range greenish-grey to black, carbonaceous shales with sand flasers alternate with cross-bedded sandstones. The Formation contains a rich fauna of brachiopods *(Discina, Martiniopsis, Chonetes),* bivalves (*Eurydesma*), gastropods, pteropods (*Conularia*), bryozoans and ostracods, as well as spores. On the basis of its faunal content and its gradational contact with the underlying Tobra Formation, the Dandot Formation has been dated as Early Permian (Teichert 1967).

The depositional environment is marine in eastern Salt Range and becomes shallower toward the west (tidal-flat facies in central Salt Range). The thickness of the Dandot Formation is 45 m–50 m in the east (Makrach Nala) and decreases toward the central part of the Salt Range.

The **Warchha Sandstone** (Hussain 1967) rests conformably upon the Dandot Formation. It represents the middle part of the "Speckled sandstone" of Waagen (1879) and the lower part of the "Warchha Group" of Noetling (1901). Type locality is the Warchha Nala in west-central Salt Range. The Warchha Sandstone is widely exposed in the Salt Range and the Khisor Range. The sandstone is generally thick-bedded to massive, reddish-brown, cross-bedded, medium to coarse-grained and arkosic. Intercalated purple to dark-grey shale layers reach a thickness of several metres each. In the western Salt Range, reddish-brown to dark-brown sandy shale with white or reddish sandstone layers prevail. In Khajji Wahan, thin shaly coal beds are developed at the base of the Formation. The Warchha Sandstone is unfossiliferous. It is considered as Early Permian because of its position between the fossiliferous Early Permian Dandot and Sardhai Formations. The sediments of the Formation are interpreted as fluviatile, deposited in extended alluvial flats. The thickness of the Warchha Sandstone reaches 150 m–165 m in the Salt Range and 120 m in the Khisor Range. In the Kohat–Potwar foredeep, the thickness generally increases from the south-east to the northwest.

The Warchha Sandstone has a transitional contact with the overlying **Sardhai Formation** (Gee in Shah 1977) which is an equivalent of the "Lavender Clays" of Wynne (1878) and of the upper part of the "Warchha Group" of Noetling (1901). Type locality is the Sardhai (Sohal) Nala in the eastern Salt Range. The Formation has an areal distribution similar to the Warchha Sandstone. The prevailing lithology in the eastern and central Salt Range is bluish-grey, purple or reddish claystone which become dark-violet to black toward the western Salt Range. Plant remain and fish scales have occasionally been found. In the Khisor Range, the claystone contains layers of argillaceous, fossiliferous limestone which have yielded bryozoans and brachiopods (e.g., *Anastomopora* sp., *Fenestella* sp., *Athyris, Spirifer*) indicative of an Early Permian age (Hussain 1967). The palaeoenvironment is interpreted as mainly terrestrial, partly lagoonal, with marine incursions which become more frequent toward the west. The thickness of the Sardhai Formation is 40 m at the type section and 60 m in the western

Salt Range (Zaluch Nala), increasing toward the northern and western parts of the Kohat–Potwar area.

Zaluch Group: It is subdivided into three formations distinguished from each other by differences in the proportion of limestone. These are the Amb Formation, Wargal Limestone and Chhidru Formation.

The **Amb Formation** (Teichert 1966), first described as "Lower Productus Limestone" by Waagen (1879) and subsequently as "Amb Sandstone Beds" by the same author (1889, 1891), overlies the Sardhai Formation disconformably. The type section is 5 km southwest of the village of Amb in west-central Salt Range. The Zaluch Nala and Chhidru Nala in the western Salt Range are additional reference sections. Exposures of the Formation extends from the central Salt Range in the east to Khisor Range in the west. In eastern Salt Range and southeastern Kohat–Potwar area, the Amb Formation has been truncated by the Paleocene transgression. The prevailing lithology is a highly fossiliferous, calcareous sandstone alternating with sandy limestone and dark-grey, locally black and carbonaceous shale. Thin coal beds are developed in the vicinity of the Amb village. The Amb Formation contains abundant floral remains, including *Glossopteris* and *Gangamopteris* (Balme 1970), and a rich fauna including foraminifera, bryozoans, brachiopods, pelecypods, gastropods, cephalopods and ostracods (Waagen 1879, 1889, 1891, Dunbar 1933, Reed 1941, Pascoe 1959, Teichert 1966, Kummel and Teichert 1970, Pakistan--Japanese Research Group 1985). The brachiopods include different species of *Orthotichia, Cleiothyridina, Neochonetes, Derbyia, Dictyoclostus, Marginifera, Spirifer, Neospirifer, Spirigerella, Strophalosia, and Dielasma.* The fusulinids *Monodiexodina kattaensis* and *Codonofusiella laxa* indicate an Early Permian (Late Artinskian) age. The rich fauna of the calcareous sandstone and limestone proves a shallow to very shallow marine paleoenvironment, probably with intermittent periods of paralic or lacustrine conditions during which the carbonaceous shale and coal were deposited. The Formation is as much as 80 m thick in the Salt Range and 40 m–50 m in the Khisor Range.

The Amb Formation is disconformably overlain by the **Wargal Limestone** (Teichert 1966), equivalent of the "Middle Productus Limestone" of Waagen (1879) and the "Wirgal Group" of Noetling (1901). The Formation is widely exposed in the central and the western Salt Range and in the Khisor Range. The Formation is mainly formed by grey, medium or thick-bedded to massive, partly sandy limestone and dolomite with few, thin intercalations of dark-grey to black shale. The Wargal Limestone is highly fossiliferous. It contains an abundance of brachiopods, trilobites, pelecypods, gastropods, ammonoids, nautiloids, echinoids, corals, bryozoans, sponges, foraminifera, ostracods, conodonts and fish remains, as well as algae and spores (Waagen 1879-1891, Reed 1944, Rao and Verma 1953, Grant 1966, Teichert 1966, Kummel and Teichert 1970, Pakistani–Japanese Research Group 1985). Among the brachiopods are *Enteletes, Derbyia, Waagenites, Waagenoconcha, Richthofenia, Oldhamina, Linoproductus, Spirigerella, Costiferina, Chonetella, Cleiothyridina, Phricodothyris, Notothyris, Hemiptychina, Terebratuloidea, Kiangsiella, Uncinunellina,* and *Callispirina.* The trilobites are represented by *Kathwaia capitorosa* and *Ditomopyge fatmii,* the ammonoids include *Xenodiscus* and *Pseudogastrioceras.* Corals are rare to common, and include *Wentzelella, Wentzelellites, Iranophyllum, Sinopora* and *Michelinia.* The fusulinids include different species of *Reichelina* as well as *Codonofusiella, Nanlingella, Nankinella, Sphaerulina, Chusenella* and *Schubertella*. The conodonts are represented by *Iranognathus, Hindeodus* and *Gondolella*.

On the basis of its faunal content, the Wargal Limestone has been dated as Late Permian (late Murgabian to early Dzhulfian). The palaeoenvironment is interpreted as generally shallow marine, with the exception of few strata which might have been deposited under deep-water conditions (Pakistani–Japanese Research Group 1985). The thickness of Wargal Limestone is 180 m–200 m in the Salt Range and 150m in the Khisor Range.

The **Chhidru Formation** (Dunbar 1933) which represents the "Upper Productus Limestone" of Waagen (1879) and the "Chhidru Group" of Noetling (1901), respectively, overlies the Wargal Limestone with a conformable, transitional contact. Type section is Chhidru Nala in the western Salt Range. The Formation is exposed in the central and western Salt Range as well as in the Surghar Range and the Khisor Range. In the southern and eastern parts of the Kohat–Potwar area, it is missing under the Paleocene unconformity. It consists of dark-grey, sandy shales at the base, overlain by calcareous sandstone and sandy limestone. The top of the Formation is formed by a characteristic white sandstone layer as much as 5 m thick. The total thickness of the Formation ranges from 75 m to 85 m.

The Chhidru Formation contains abundant brachiopods and gastropods and subordinate microfossils. The fauna has been described in detail by Kummel and Teichert (1970) and the Pakistani–Japanese Research Group (1985). It includes the following genera:

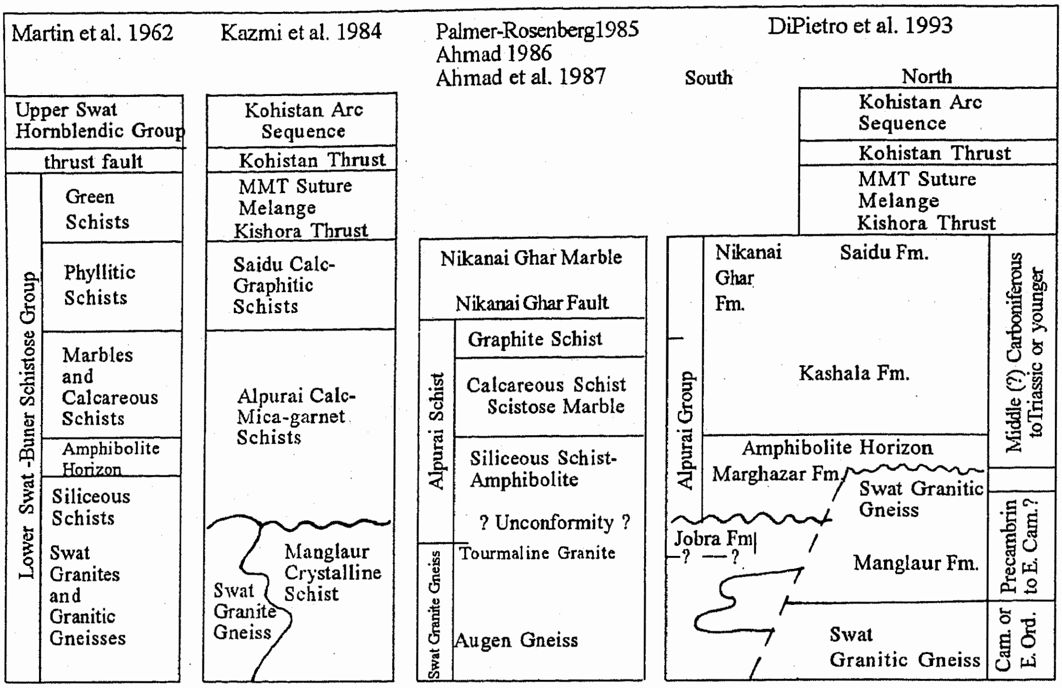
Brachiopods: *Aulosteges, Callispirina, Chonetella, Cleiothyridina, Crurithyris, Derbyia, Dielasma,Enteletes, Hemiptychina, Hustedia, Kiangsiella, Linoproductus, Lyttonia, Marginifera, Martinia, Megasteges, Neospirifer, Oldhamina, Orthotichia, Richthofenia, Spinomarginifera, Spiriferella, Spirigerella, Strophalosia, Waagenites, Waagenoconcha, Whitspakia.*
Gastropods: *Bellerophon, Euphemites.*
Pelecypods: *Entolium, Neoschizodus, Permophorus, Palaeolima, Schizodus, Crytorostra.*
Ammonoids: *Cyclolobus, Eumedlicottia, Episagoceras, Stacheoceras, Xenodiscus.*
Bryozoans: *Dybowskiella, Hexagonella, Geinitzella, Stenopora, Fenestella, Polypora.*
Fusulinids: *Codonofusiella, Reichelina, Nanlingella, Nankinella* in association with *Colaniella* and other smaller foraminifera.

Stratigraphic analyses indicate that the Chhidru Formation is most probably of Late Dzhulfian age, and that the uppermost part of the Permian (Changhsingian, Dorashamian) is missing within the Salt Range section (Pakistani–Japanese Research Group 1985). The abundance of terrigenous clastic material in Chhidru Formation indicates a nearshore environment and/or an uplift of the source area. The white sandstone forming the top of the Formation was subjected to subaerial erosion (Kummel and Teichert 1970).

Peshawar Basin

Khyber Limestone: In the Peshawar Basin and in the Khyber Agency, a number of formations are observed to which an undifferentiated Carboniferous to Permian age has been assigned. In the Khyber Agency, the Upper Devonian Ali Masjid Formation is overlain, conformably by the Khyber Limestone (Stauffer 1968b, Griesbach 1892, Hayden 1898). Type locality is the village of Ali Masjid (Khyber Pass). It consists mainly of thick-bedded limestones, marbles and massive dolomites, locally with minor shale intercalations. The middle part of the Formation contains some limonite-rich arenaceous beds which

Table 5.4. Tectonostratigraphic sequence in Lower Swat (from DiPietro et al. 1993).



might indicate disconformity (Shah et al. 1970). To the north and west of the type section, the upper part of the Formation is intruded by basic dykes and sills. At Tauda Mela and Misri Khel, the Khyber Limestone is intercalated with shale and is highly fossiliferous. The fauna collected from the upper part of the Formation includes *Pseudovermiporella sp.*, *Nodosaria sp.*, *Geinitzella sp.*, *Glomospirella sp.*, *Robuloides sp.*, *Palaeotextularia sp.*, and *Frondilina sp.*, and indicates a Permian age (Shah et al. 1970).

Jafar Kandao Formation: In the hill ranges northeast of Nowshera in the Rustam area, a sequence of argillites with subordinate limestone, quartzite and conglomerate overlies the Nowshera Formation. It has been named Jafar Kandao Formation by Pogue and Hussain (1986). The basal unconformity is marked by discontinuous conglomerate bed. The Formation is overlain by greenschist. According to Pogue et al. (1992), latest Devonian to Early Mississippian and Late Pennsylvanian conodonts have been found in the limestone lenses in this Formation. It is thus Carboniferous in age.

Swat

Alpurai Group: Northeast of Nowshera in the Swat area, several hundred metres of unfossiliferous siliceous schist, calcareous quartz-mica-garnet schist, marble and para-amphibolite overlie unconformably the Swat Granite Gneiss. This sequence was named Alpurai Schists by Kazmi et al. (1984c, 1986) and according to them, these rocks could have occurred anytime during Palaeozoic to Mesozoic. This unit forms the cover of the Precambrian Manglaur Schist and is a tectonostratigraphic component of the Swat nappe (Treloar et al. 1989a, 1989b). The Alpurai Schists extend over a large area (Ahmad et al.

1987, Lawrence et al. 1989). DiPietro (1990, 1991) has correlated a number of other lithostratigraphic units with the Alpurai Schists, and has referred to them as Alpurai Group. His Alpurai Group includes the Marghazar, Kashala, Saidu and Nikanai Ghar Formations (Table 5.4).

Marghazar Formation: This Formation consists of garnetiferous schist, amphibolite, hornblende schist, psammitic schist and phlogopite marble. It overlies the Manglaur Formation and Swat Granite Gneiss unconformably (c.f., Alpurai Schist of Kazmi et al.). The Marghazar Formation is overlain by the calcareous schists of the Kashala Formation followed by the Saidu Schist and Nikanai Ghar Formations. The Kashala Formation has yielded Late Triassic (Carnian) conodonts (Pogue et al. 1992) and, therefore, at least the sequence above the Marghazar Formation of DiPietro (1990, 1991) is Mesozoic in age. The Marghazar Formation is tentatively considered to be of Carboniferous age (Pogue et al. 1992).

Hazara

Banna Formation: East of Besham, near Banna village, the Banna Nappe is mostly a series of marble, slate, graphitic schist and chlorite schist which have been referred to as Banna Formation (Tahirkheli 1979, Treloar et al. 1989a, 1989b). This sequence occurs in the immediate footwall of MMT and is characterised by brittle deformation features. It is separated from the underlying sillimanite gneisses of the Hazara Nappe by the extensional Banna Thrust (Fig. 4.36).

Hazara–Kashmir Syntaxis

Panjal Formation: Around the apex of Hazara–Kashmir Syntaxis, sediments and volcanic rocks are exposed which were described as Panjal System by Lydekker (1878), and Volcanic Greenstone and Agglomerate Slate by Middlemiss (1910). Calkins et al. (1969) subdivided these rocks into the Panjal Formation, formed by moderately metamorphic lava flows and tuffs ("greenstones"), and the Agglomerate Slate, consisting of black, Carboniferous shale, slate and phyllite as well as quartzose, agglomeratic sandstone. Calkins et al. (1969) have, however, applied the name Panjal Formation to include both the units and have not discussed their contact relations or superposition. They are considered as of Carboniferous to Permian age, based on the determination of rare fossils by Holland (1926).

Kaghan

Farther eastward, in the upper Kaghan Valley, calcareous garnet-kyanite-bearing schist, marble and amphibolite of the **Sharda Group** (Ghazanfar et al. 1986) are exposed south of the MMT and form a thrust slice between the north dipping Batal Thrust and the MMT (Treloar 1989b). These rocks are similar to the Alpurai Schists. According to Treloar (1989a), it is likely that the cover sediments of the Sharda Group, the Banna Formation and the Alpurai Schists may correlate. As mentioned earlier, Greco et al. (1989), however, divided the Sharda Group of Ghazanfar and Chaudhry into a granitic basement complex (Cambrian) which is unconformably overlain by Naran Formation, which they described as "Lower Paleozoic basis of the cover" (Table 5.2). The latter is followed by Burawai Formation which has been described as "Paleomesozoic metasedimentary cover sequence" by Greco and Spencer (1993).

Naran Formation (Greco et al. 1993): It is largely comprised of metapelitic-metagreywacke gneisses followed by thick amphibolitic layers and diorite dykes. On petrographic, geochemical and field evidence Greco and Spencer (1993) correlate the amphibolites with the Permian Panjal Trap and describe the overlying paragneisses as a "distinct second lithotype of basement" and which are the "metamorphic, magmatic and sedimentary products of a Late Pan-African orogenic event" and form the pre-Permian basement on which rocks of the Alpine sequence were deposited.

Burawai Formation (Greco et al. 1993): It forms the cover sequence above the Naran Formation and is largely comprised of garnetiferous calc-pelites, marbles and subordinate amounts of quartzites. These are interlayered with amphibolites (basaltic flows) which range in thickness from thin bands to thick beds.

Karakoram

The sedimentary rocks of the Karakoram are comprised of Paleozoic to Mesozoic metasediments. Which largely occur in the form of stacks of thrust sheets and are not so fossiliferous. The stratigraphy and palaeontology of this region is yet poorly known and is mainly based on reconnaissance mapping of scattered localities. Stratigraphical correlation across the Range is therefore fraught with difficulties and at best it is very tentative.

The Tirich Mir Zone

Lun Shales: The Karakoram contains a series of metasediments, believed to be largely Permo-Carboniferous in age. Hayden (1915) introduced the name "Sarikol Slate" for a sequence of slates extending from the Chinese border through Chitral to eastern Afghanistan. Stauffer (1975) and Calkins et al. (1981) have used the same name whereas Desio (1959, 1966, 1975), Tahirkheli (1982) and Pudsey et al. (1985) refer to these rocks as Lun Shales. This unit is comprised of a thick sequence of grey, splintery, slaty shales and slates with quartzites, dolomites and limestones and overlies the Shogram Formation. It covers a large area between Tirich and Reshun Faults in NW Chitral (Fig. 4.40). According to Pudsey et al. (1985) however, the Lun Shales extend westward across the Tirich Fault up to the Afghan border and have been intruded by granitic plutons. Towards the Tirich Mir pluton the metamorphic grade increases to garnet-staurolite schists with granite veins. The quartzites are thin-bedded to massive, cross-bedded and bioturbated. The limestone are commonly grey, thin to thick bedded or massive, micritic, and at places contain bryozoa, corals, brachiopods, crinoids, fenestellid ectoprocts, algae and Permian fusulines (Talent et al. 1979). The dolomites are largely interbedded with terrigenous succession in units tens of hundreds of metres thick (Pudsey et al. 1985).

The upper part of the Lun Shales has been apparently truncated by the Tirich Fault. Its lower contact with the underlying Shogram Formation has not yet been demarçated or described by any worker. Devonian fossils have been reported from strata believed to be part of Lun Shales (Calkins et al. 1981, Tahirkheli 1982, Pudsey 1985). Species *Bariophyton* and Carboniferous ferns occur in the highest beds of the Kuragh section (Talent et al. 1979).

A thick scarp-forming fusulinid limestone near Parpish village has been named **Parpish Limestone** (Pascoe 1923, Tahirkheli 1982). It is apparently a part of the Lun Shale sequence. On the left bank of the Mastuj Valley, between Tormal Gol and Zait, a thick massive and black dolomitic limestone with a 6 m thick pisolitic ironstone, containing Lower Triassic conodonts has been named **Zait Limestone** (Tahirkheli 1982). It overlies Early Devonian sandstones and occurs within the locality mapped as Shogram Formation.

About 30 km north of Zait and not far from the Tirich Fault, Middle or Late Jurassic ammonites have been collected from Rosh Gol (Talent et al. 1979). This region is apparently comprised of the upper part of the Lun Shale sequence which may therefore extend from Devonian ? or Carboniferous to Late Jurassic (Calkins et al. 1981, Pudsey et al. 1985). Pudsey et al. (1985) suspect an overlap in age between the Devonian carbonate unit (Shogram Formation) and Lun Shales, the former representing a marine shelf environment and the latter deposited in deeper environments.

Leake et al. (1989) have described a monotonous sequence of light green phyllites, east of Tirich Fault, as **Lutkho Formation**. He considers it as a probable subdivision of the Lun Shales.

Arkari Formation: In Chitral District, the region west of the Tirich Fault is largely comprised of metasediments which have been metamorphosed to at least lower amphibolite facies and have been named Arkari Formation (Leake et al. 1989). Earlier Pudsey et al. (1985) had included it in their Lun Shales and Calkins et al. (1981) had referred to it as "Jurassic to Devonian rocks". Leake et al. (1989) mention that the Arkari may be a subdivision of the Lun Shales. It is largely comprised of mica schist which is interlayered with feldspar-rich schist and, to a lesser extent, with graphitic schist. Grey, finely foliated phyllite, micaceous quartzite up to 25 m thick and calcsilicate quartzite interbedded with schists are common. At places tourmaline schists or gneisses and tourmalinite also occur in this formation. Within the Arkari Formation there are many extensive marble zones up to 1 km thick and several kilometres long. The marble is grey to brown with a coarse-grained mosaic of calcite and some muscovite and quartzite. Thin layers containing concentrations of muscovite, phlogopite and chlorite give it a banded appearance. The marble and calcsilicate quartzite commonly contain scheelite (Leake et al. 1989).

The Arkari Formation is unfossiliferous and if we accept it as a metamorphosed facies and an upper member of the Lun Shale and consider the Belemnite found north of Tirich Mir pluton (Pascoe 1923), its age may be Permian to Jurassic.

Southern Sedimentary Belt

Darkot Group: In the Mastuj, Yasin, Ishkuman and Hunza areas, north of the Main Karakoram Thrust (MKT), a thick sequence of slate, limestone, quartzite, conglomerate, schist, marble, gneiss and volcanics is exposed and has been named Darkot Group by Ivanac et al. (1956). The Group forms two east-west trending, long belts, (see Chapter 4). The Darkot Group has yielded bryozoans, forams, brachiopods, crinoids, corals and gastropods from a few widely scattered localities and has been assigned a Permo-Carboniferous age by Ivanac et al. (1956).

Dumordo Formation: The Darkot Group has been divided into a number of separate units by various workers. Stauffer (1968c) named the rocks exposed south of the Karakoram Granodiorite in Hunza Valley as Baltit Group, whereas Desio (1963, 1964) had already named them Dumordo Formation. The latter name takes priority according to the Stratigraphic Code of Pakistan. This Formation consists largely of garnet-staurolite schist, garnet-mica schist, garnet amphibolite, crystalline marble and micaceous quartzite with scattered layers and lenses of biotite gneiss. Its contact with Karakoram Granodiorite is partly gradational and partly intrusive. No fossils have been found in the Dumordo Formation. Stauffer (1968c) has tentatively placed it in the Permo-Carboniferous whereas Tahirkheli (1982) correlated this Group with his Chitral Slates and placed it in the Precambrian-Lower Paleozoic.

Chalt Formation: In the Hunza Valley the Dumordo Formation is exposed along an overturned anticline and is faulted against the Chalt Formation (Desio 1963; 1964, Desio et al. 1972, Stauffer 1968c, Gansser 1980). The Chalt Formation consists of a dark-grey quartz-biotite schist with subordinate quartzite, marble and conglomerate. Near Sandi village, about 80 km west of Chalt, poorly preserved bryozoans, *Fenestella* and *Rhombopora cf., Lepidodendroides* Meek were found, along with a large, crushed productid and crushed coral specimen (Ivanac 1956, Stauffer 1968c). On this evidence the Chalt Formation has been assigned a Permo-Carboniferous age. Searle (1991) has referred to the Chalt Formation as Ganchen Formation, name earlier given by Desio (1963) to a metasedimentary sequence overlying the Dumordo Formation in the Braldu Valley.

Karakoram Metamorphic Complex: Metamorphic rocks south of the Karakoram Batholith, those described above and others (Hushe Gneiss, Ashkole Amphibolite, Ganchen Formation, Panmah Ultramafic and Dassu Gneiss) have been grouped as Karakoram Metamorphic Complex by Searle (1991) because there is as yet insufficient information and no age control to sort out their relative stratigraphic position (see Chapter 7).

Ganchen Formation: Named after the Ganchen Peak (6,462 m) by Desio (1963, 1964, 1979), this Formation crops out north of the Shigar Valley (Skardu District) and is largely comprised of metapelites and less common gneisses and amphibolites. Thick beds of crystalline limestone and dolomite are included. The pelitic assemblages vary in metamorphic grade and in general are characterised by kyanite-staurolite-garnet-biotite-muscovite-plagioclase-quartz assemblage (Searle 1991). Though originally defined and mapped by Desio (1964) in the region north of Shigar Valley, where it was shown to be in contact with the Dumordo Formation, Searle (1991) has extended this Formation to include the Chalt Formation in Hunza.

Rocks shown as Ganchen Formation in Searle's map northeast of Shigar Valley, have been named and mapped as **Daltumbore Formation** by Hanson (1989) who has described them as interbedded layers of slate and marble. The metamorphic grade increases northwestward and the slates are metamorphosed to garnet-biotite schists. Previously Hanson's Daltumbore Formation was divided into three units by Desio (1963, 1964) and Zanettin (1964), namely the **Lugma slates,** the **Daltumbore mica-schist** and **Askore amphibolite**. Brookfield and Gupta (1984) reported Permian brachiopods, crinoids and bryozoans in the limestone beds, and, according to Hanson (1989) Mississippian conodonts and horn corals of probable Devonian age have been also found. The Daltumbore Formation is thus believed to be Permo-Carboniferous.

Askore Amphibolite: Searle (1991) has described thin bands of amphibolite within marbles (Dumordo) interlayered with pelites south of Aliabad in Hunza Valley. This unit is associated with minor amounts of quartzites, greenschists, agglomeratic slates and tuffs. This Askore Amphibolite is not to be confused with the Askore amphibolite mapped by

Desio (1964) on the eastern and western margins of the Nanga Parbat–Haramosh Massif and the Shigar Valley, and which is now obsolete and abandoned.

Ashkole Amphibolite (Searle 1991): This unit comprises dark-coloured amphibolites interbedded with marbles, schists and gneisses and is exposed near Ashkole village, Mango Gusar and Panmah and Braldu Valleys. The amphibolites contain hornblende-biotite-plagioclase-quartz ± garnet ± clinopyroxene ± epidote and they represent metamorphosed volcanic or intrusive rocks associated with a limestone and shale sequence.

Panmah Ultramafic (Searle 1991): This unit is comprised of a melange which consists of blocks of gabbro, metabasalt, chert and ultramafic rocks in a shaly matrix. This tectonic ultramafic-mafic melange occurs in discontinuous bands in a narrow zone south of the Karakoram Batholith, bounded by thrust and shear zones within the Dumordo and Ganchen Formations. According to Searle (1991) it may represent a Mesozoic suture zone with an ophiolite complex which has been obliterated by post-collision ductile shearing and folding.

Hushe Complex: A sequence orthogneisses associated with thin bands of marble and arenaceous metasediments crop out in the Hushe Valley (north of Khapalu village) along the hanging wall of the MKT east and north of Shigar town in Skardu District and has been named Hushe Complex by Searle et al. (1986, 1989). It is largely comprised of hornblendite, hornblende diorite, amphibolite, biotite granodiorite-monzogranite. Rocks from this complex have given a U-Pb zircon age of 145–150 Ma and K-Ar hornblende ages of 208 ± 8 Ma and 163 ± 7 Ma. Desio (1964) had earlier mapped these rocks with his Baumaharel Schist, Askore Amphibolite and Ganchen Formation at different localities.

Dassu Gneiss: Near Dassu village, north of Shigar Valley, a series of orthogneisses with intercalations of garnetiferous mica-schists have been named Dassu Gneiss. These rocks as well as the other units of the Karakoram Metamorphic Complex mentioned above have been discussed in Chapter 7 also.

Northern Sedimentary Belt

Along the northern margin of the Karakoram Batholith, the upper Hunza Valley is traversed by the NW-SE trending Misgar and upper Hunza Thrust Faults, which divide the region into three tectonostratigraphic units namely Guhjal, Sost and Misgar (Gaetani et al. 1993, Zanchi 1993). The Guhjal Unit is comprised of Paleozoic metacarbonates (Guhjal Dolomites) and black terrigenous slates (Pasu Slates). Northward it has been thrust over the Sost Unit (Fig. 5.4) which comprises an antiformal stack of Permian to Cretaceous sediments. The Misgar Unit structurally overlies the Sost Unit and is comprised of monotonous dark slates (Misgar Slates). The Permian-Carboniferous sequence in these tectonostratigraphic units, as well as their eastward extension into the Baroghil Pass region, has been briefly described below.

Pasu Slate: North of the Karakoram Granodiorite a thick sequence of dark-grey and black slates with intercalations of dark-grey limestone and yellowish grey quartzite has been named Passu Slate (Schneider 1975, Desio 1963). The contact of the slate with the Karakoram Granodiorite is sharp and intrusive (Stauffer 1968c); however, according to Tahirkheli (1982), it is a tectonic contact. Based on bryozoans, forams and brachiopods found near Khaibar village, the Passu Slate is assigned a Permo-Carboniferous age (Ivanac et al. 1956, Stauffer 1968c).

Guhjal Dolomite: North of Passu village, crystalline dolomites, quartzites and dark slates, named Guhjal Dolomite by McMahon (1900) and Desio (1963, 1964), are faulted into the Passu Slate. This sequence is largely comprised of crystalline limestone and dolomites with intercalations of dark-brown slates and quartzites. In the Passu region this unit has a transitional lower contact with Passu Slate. The transitional lithofacies comprises 20–30 m thick sequence of interbedded slates and marly limestone. A fusulinid assemblage dominated by *Parafusulina sp.,* has been recently reported from the marly limestones of this transitional facies, suggesting a Middle to Late Permian age (Gaetani et al. 1995). Earlier this Formation was considered to be of Triassic age (Desio 1964), due to the presence of megalodontids and colonial scleractinians.

Gircha Formation: The earliest rocks exposed in the Sost tectonostratigraphic unit consist of thick-bedded to massive sandstone with slate intercalations. These units are overlain by dark argillites with thin limestone-dolomite intercalations. This sequence was named Gircha Formation by Desio (1963, 1964). The sandstone are arkosic to quartz-arenites with abundant rock fragments (largely igneous rocks). Dark pelites and thick cross-laminated channelised sandstone suggest a fluviatile or deltaic setting and deposition in rapidly subsiding rift troughs (Gaetani et al. 1990).

In the Ashtigar region of the Chapursan Valley, this Formation is comprised of a lower shaly unit interbedded with fine-grained subarkose containing Early Permian brachiopods. The middle part of the sequence comprises fine-grained, cross-laminated sandstone with mudclasts interbedded with dark grey siltstones. The upper part consists of thin black shales overlain by cross-laminate, medium-grained, white quartzose sandstone. According to Gaetani et al. (1995), the Gircha Formation continues westward into the Baroghil area of Chitral District, where it is thinner and is largely comprised of interbedded siltstones and sandstone, coarse grained sandstones with mudclasts, and medium-grained cross-bedded, feldspathic quartzarenites.

According to Tahirkheli (1982) the calcareous part of the sequence contains brachiopods, corals, bryozoans and foraminifera of Early Permian age. Gaetani et al. (1995) have reported brachiopods (*Lyonia sp., Rhynchopora sp., Trigonotreta Stokesi, T. lyonsensis, Spirelytha petaliformis, Punctospirifer afghanus*) and bivalves (*Dellopecten sp., Leiopteria sp, Eurydesma sp*) from the middle part of the Formation in the Chapursan Valley. This fauna suggests an Early Permian age for the middle part of the sequence, with the possibility that the lower (unfossiliferous) part of the exposed sequence may represent part of the Carboniferous (Gaetani et al. 1995). The base of the Formation is not exposed. Its upper boundary in Gircha–Chapursan area is transitional from a sandstone shale sequence to the overlying sandstone siltstone-limestone sequence of the Chapursan Group. Gaetani et al. (1995) consider Pasu Slates as being a local facies of the Gircha Formation and also include in it the Shimshal Slates of Casnedi et al. (1985) and Kilik Formation of Desio (1963).

Chapursan Group: In the Shimshal and Chapursan Valley of upper Hunza, the Gircha Formation is overlain by a sequence of Middle to Late Permian marine sandstones, siltstones and limestone. This sequence was earlier named Panjshah Formation by Gaetani et al. (1990a) but now they have divided this sequence into two units Lupghar and Panjshah Formations which crop our in the Upper Hunza region, and the Lashkargaz Formation which is exposed in the Baroghil region (Table 5.5).

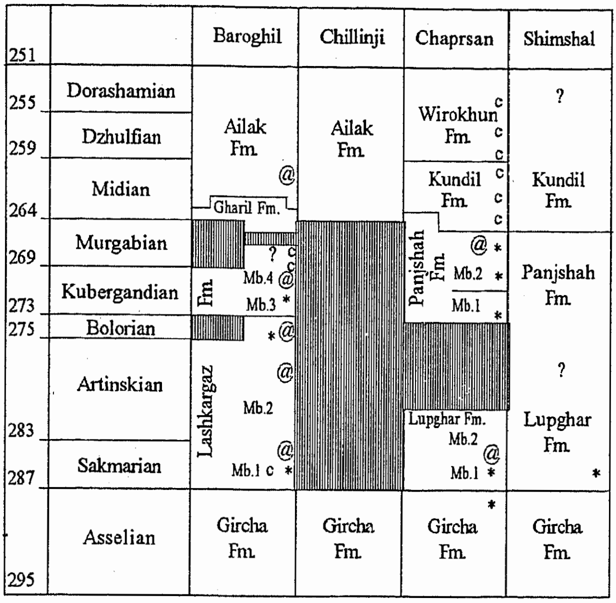


Table 5.5. Permian stratigraphic sequence in the Northern sedimentary belt of the Karakoram (from Gaetani et al. 1995). Time scale after Ross et al. (1994). In this time scale the span attributed to the Sakmarian is possibly too short. * = brachiopod occurrence; @ = fusulinid or other foraminifer occurrence; c = conodont occurrence.

Lupghar Formation (Gaetani et al. 1995): This Formation is about 300–380 m thick and is comprised of interbedded shales mudstones and sandstones with phosphatic nodules in the lower part, followed by thick bedded bioclastic and oolitic limestone and dolomite. The lower terrigenous sequence contains brachiopods (*Permochonetes pamiricus, Reticulatia sp, Globiella cf. rossiae, Costatumulus irwinensis, Cleiothyridina ailakensis, Cleiothyridina sp, Cyrtella cf. nagmargensis, Hunzina tenuisulcata, Trigonotreta paucicostulata*) whereas the upper carbonate sequence is characterised by a fusulinid assemblage (*Pseudofusulina plena cf. psharti, P. cf. karapetovi, P. tumidiscula, P. incompta, P. cf. sedujachensis, P. cf. callosa, P. cf. granuliformis, Eoparafusulina sp., E. pamiriensis*). The Lupghar Formation is conformably overlain by the Panjshah Formation.

Panjshah Formation (Gaetani et a. 1990a, 1995): This unit is about 230–260 m thick and contains interbedded dark shales, siltstones, and cross-laminated feldspathic quartzarenites in the lower part and fossiliferous dark grey limestone, marls interbedded with marly limestone and dark marls interbedded with dark-grey limestones in the upper part. The lower part of the Formation contains abundant *callytharrella sinensis*, along with *Derbyia grandis, Orthothetina convergens* and *Costiferina sp*. The upper part bears a rich fauna comprised of foraminifera (*Neofusulinella sp., Chitralina undulata, Globivalvulina sp., Dekerella sp., Langella sp., Climacammina sp., Geinitzina sp.*), brachiopods (*Stenoscisma armenica, Chapursania tatianae, Retimarginifera sp., R. goetaii; Magniplicatina sp., Martinia sp., Martiniopsis sp.*), corals (*Duplocaninia sp., Ufimia hunzensis, Paracaninia sp*) as well as echinoderms and bryozoans (Gaetani et al. 1995). This fossil assemblage suggests that the Panjshah Formation ranges from Kubergandian to Murgabian. It is conformably overlain by the Late Permian Kundil Formation.

Lashkargaz Formation: A thick sequence of shales with subordinate sandstones and limestones, and with abundant late Early to Middle Permian fossils is exposed in the Baroghil region of the Upper Yarkhun Valley and has been named Lashkargaz Formation

(Gaetani et al. 1995). Its base is not exposed, having been covered by the stream-bed deposits of the Yarkhun River, which separates it from the underlying outcrops of the Gircha Formation. The Lashkargaz Formation is up to 1,000 m thick and in ascending order it is comprised of siltstones containing phosphatic nodules, brachiopods and conodonts; oncoidal limestones, shales and dolomites with fusulinids and other fossils; sandstones, shales and marly limestones; and grey cherty limestones and dolomites characterised by fusulinids, corals, brachiopods, crinoids and bivalves.

Some of the significant fossils found in this Formation include: Foraminifera–*Pseudofusilina, Pseudoendothyra, Chalaroschwagerina, Pamirina, Misellina parvicostata;* conodonts– *Adetognathus paralautus, Sweetognathus aff. whitei, Gondolella cf. idahoensis, Anchignathodus sp.* and *Iranognathus sp.;* corals–*Protomichelinia sinopora? cf. syrinx* and *Yatsengia hangchowensis;* bivalves–*Girtypecten sp., Etheripecten sp., Permopecten sp.,* brachiopods– *Hunzina electa, Derbia cf. baroghilensis* (Reed), *Cleiothyridina ailakensis* (Reed), *Neochonetes costellata, N. naroghilensis* (Reed), *Retimarginifera praelecta* (Reed). According to Gaetani et al. (1995) this fauna suggests that the Lashkargaz Formation ranges from Sakmarian to Kubergandian and that it is a lateral facies variant of the Lupghar–Panjshah Formations which cropout only about 90 km to the east in the Chapursan Valley (Table 5.5).

Gharil Formation (Gaetani et al. 1995): In the upper Yarkhun Valley, this Formation overlies the Lashkargaz Formation. In the Baroghil region it lies on a deeply scoured surface of the Lashkargaz Formation and is comprised of sandstones, poorly sorted microconglomerates, and dark red hematitic cross-bedded sandstones (ironstone). Eastward, in the Lashkargaz area, the Formation largely consists of extensively burrowed, grey siltstones, very fine-grained sandstones and marls, capped by conglomerates and quartz-bearing dolostones (Gaetani et al. 1995). Age-diagnostic fossils have not been found in this Formation, though on the basis of its stratigraphic position, Gaetani et al. (1995) place it between Kubergandian and Midian.

Ailak Formation (Gaetani et al. 1995): This Formation consists of about 1,000 m thick massive dolomites and limestones exposed along the continental divide in the Baroghil region. It overlies the Gharil Formation. Its eastward extension is uncertain though according to Gaetani et al (1995) it may be the lateral equivalent of the Kundil and Wirokhun Formations of the Chapursan Valley (Table 5.5). The few fossils *(Paraglobivalvulina? sp., Dagmarita chanakiensis* Reitlinger, *Langella sp.*) found in this Formation indicate a Late Permian age (Gaetani et al. 1995).

Kundil Formation (Gaetani et al. 1990a): This Formation conformably overlies the Panjshah Formation and is comprised of a 40–50 m thick lower unit consisting of thick bedded cherty calcarenites with Middle Permian fossils (*Gondolella idahoensis, Cancellina, Praesumatrina, Minojapanella elongata*); a middle unit made up of 70 m thick grey, thick bedded, cherty limestone with late Middle Permian conodont *(Sweetognathus hanzhongensis)* and an upper unit comprising 50 m of thin-bedded greyish white, nodular limestone containing *Gondollella bitteri* and *G. rosenkrantzi.* Metabreccia up to 100 m thick occurs between the middle and upper units of this Formation.

Wirokhun Formation (Zanchi 1993, Gaetani et al. 1995): This Formation overlies the Kundil Formation. It is of limited extent and has been recognised only on the ridges between the Borom and Kundil Valleys (tributaries of Chapursan). It is comprised of dark shales and marls interbedded with grey mudstone, thin bedded cherty limestone, tuff and black splintery shales and marls. This sequence contains abundant conodonts (which include *Gondollela orientalis, G. subcarinata changxingensis, Hindeodus minuteus, Iranognathus cf. unicostatus* and *Neospathodus dieneri*). It conformably underlies the Lower Triassic Borom Formation and according to Gaetani et al. (1995) its age ranges from Late Permian (Dzhulfian) to Early Triassic (Dienerian).

Misgar Slate (Desio 1963, 1964): The Misgar tectonostratigraphic unit of Gaetani et al. (1993) and Zanchi (1993) has been thrust southward on the Sost Unit (Fig. 5.4, Table 5.1). It is largely made up of a monotonous series of dark grey slate and phyllite with siliceous and calcareous intercalations and is exposed in the uppermost part of the Hunza Valley. This Formation is about 3,500 m thick and is commonly intruded by sills and dykes of dolerite, gabbro, pegmatite, aplite and quartz-syenite (Tahirkheli 1982). No fossils have been so far reported from this Formation and it is tentatively referred to the Paleozoic.

Singhie Shales: A thick sequence of well-cleaved black shales, corresponding to the Pasu Slates of Upper Hunza, are exposed north of the Karakoram Batholith in the Shaksgam–Baltoro region of eastern Karakoram and in the Baltoro region. It has been named Singhie Shales by Desio (1936). At different localities they have been given various names (Shimshal Shale, Baltoro Shale, Rimo Shale, Sarpo Laggo Slate, Doksam sequence) by other authors. In the Baltoro region, the black shales are interbedded with grey limestone and sandstones and their metamorphic grade increases northward, where they contain chlorite, biotite, garnet and andalusite. No fossils have been yet reported from these shales and a Carboniferous age is inferred because they underlie the Permian Shaksgam Formation.

Shaksgam Formation: Named after the Shaksgam Valley, north of K2, by Auden (1938) and Desio (1963), this Formation is comprised of 600 m to 1,000 m thick fossiliferous limestone containing Permian brachiopods, lamellibranchs, corals, bryozoans, crinoids and foraminifera *(Parafusilina sp., Paraschwagerina sp.,* and *Neoschwagerina sp.)*. The limestones are thick-bedded, recrystallised and interbedded by quartz diorites and lamprophyre dykes in the vicinity of Baltoro and Abruzzi Glaciers. According to Desio (1980), thin-bedded siliceous limestones with cherty nodules occur at the tip of the Formation and stratigraphically it is overlain by Triassic Urdok Conglomerates which laterally grade into Chikchiri Shales.

Gaetani et al. (1990b) have however described the Permian section in the Shaksgam region differently. They mention a basal (Lower Permian ?) alluvial to deltaic sequence of 150 m thick, dark siltstones, fine sandstones and coarser litharenites topped by marine sandstones containing fragmented brachiopods and bryozoans. This unit is overlain by the Shaksgam Formation of Desio (1980) which according to Gaetani et al. (1990b), may be divided into three members. The lower member comprises fossiliferous crinoidal or bioclastic limestone containing foraminifera (specially *Archaeodiscidae*), algae *(Epimastopora, Tubiphytes)*, brachiopods *(Enteletes)*, corals *(Lophophyllum, Allotropichisma, Paracaninia)*, conodonts *(Merillian aff. oertii)*, and bryozoans. The middle member is made up of cross-bedded sandstone rich in fusulinids *(Parafusulina)*. The upper member is comprised of thick bedded grey nodular limestone and marls with small foraminifera and brachiopods. This sequence of the Shaksgam Formation ranges from Lower to Middle Permian.

The Cherty limestone sequence described by Desio (1980) at the top of Shaksgam

Formation has been named Staghar Cherty limestone by Gaetani et al. (1990b) and it contains spong spicules, radiolaria, redeposited fusulinids and a Late Permian conodont assemblage *(Gondolella bitteri, Godolella liangshanensis).*

MESOZOIC

Mesozoic sedimentary rocks (Table 5.6) crop out extensively in the Kirthar–Sulaiman region and the Kohat–Potwar–Salt Range region. They also form significant components of the tectonostratigraphic sequences in the Himalayan thrust-and-fold belt, the Kohistan Island arc and the Karakoram (see Chapters 4 and 6). In the Lower Indus Basin (Kirthar–Sulaiman region), the Mesozoic rocks are largely marine, calcareous and argillaceous and several thousand metres thick (Fig 5.5). Northward, in the Upper Indus Basin (Kohat–Potwar–Salt Range) they decrease in thickness to about one thousand metre and include a substantial amount of terrestrial deposits. In the Chagai area of Balochistan as well as in the Kohistan Island Arc Complex, the Mesozoic sedimentary sequence consists largely of Cretaceous volcanic and sedimentary rocks. The Himalayan thrust-and-fold belt which is mainly a series of disjointed, overlapping slices of thrusts and nappes, contains Mesozoic metasediments locally interbedded with volcanics or covered with obducted masses of ophiolites and melanges. The Mesozoic metasediments form the cover of the basement rocks. Further northwards in the Karakoram, Mesozoic metasediments crop out in the form of extensive east-west trending arcuate thrust blocks, imbricated with earlier metasedimentary sequences.

Triassic

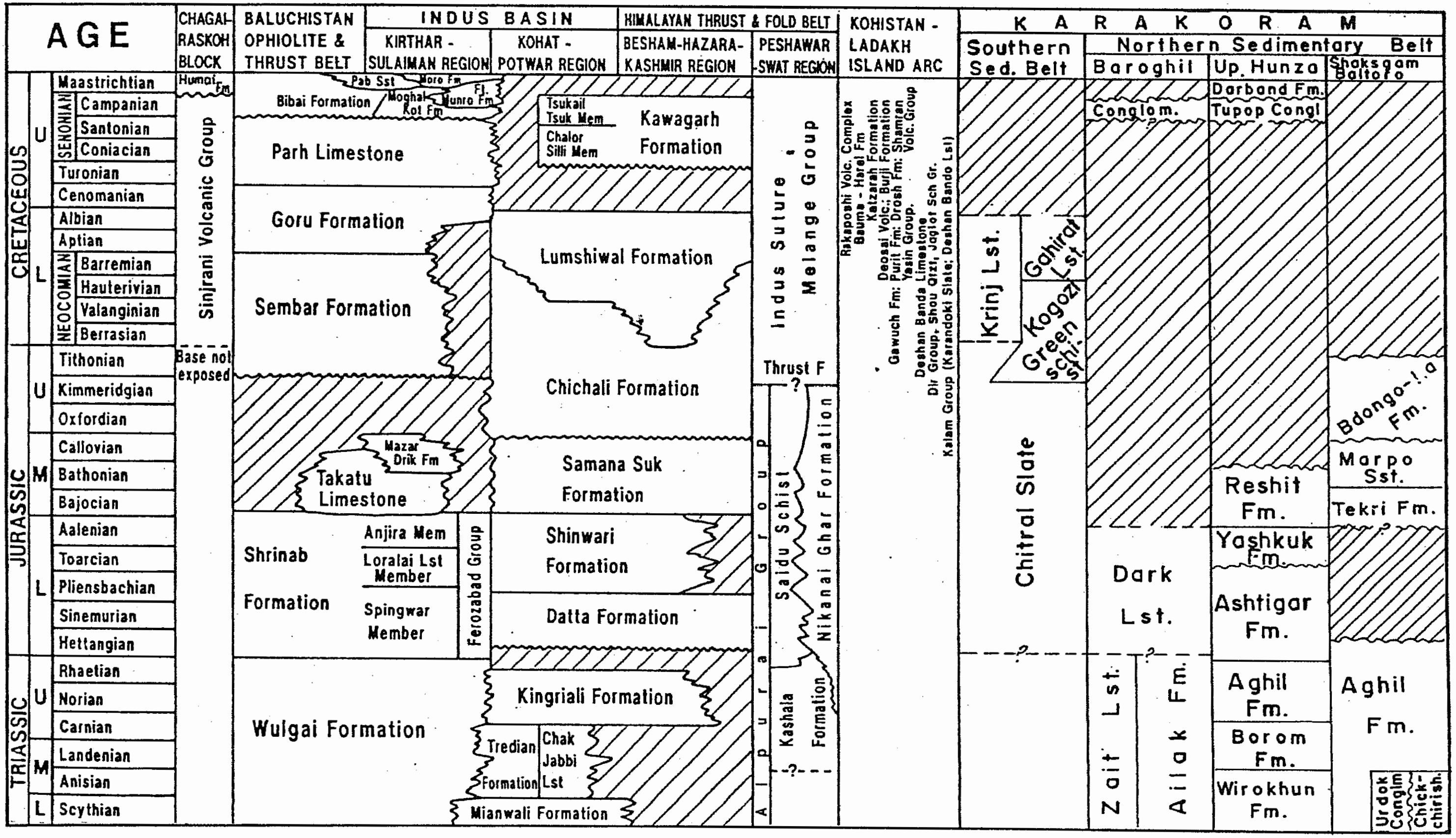
The Triassic sedimentary sequence in Pakistan is rather restricted in its thickness and extent (Table 5.6). Triassic rocks assigned to the Wulgai Formation are present in tectonised blocks in the Kirthar–Sulaiman region in Balochistan and the Himalayan thrust-and-and-fold belt. Triassic rocks belonging to the Mianwali Formation, Tredian Formation, Chak Jabbi Limestone and Kingriali Formation are exposed in the Kohat–Potwar–Salt Range region of the Upper Indus Basin. The Himalayan thrust belt in the Swat–Mardan region contains outcrops of the Kashala and Nikanai Ghar Formations which are the Triassic components of the Alpurai Group. East of this region, in the Hazara area, the Triassic sequence is missing and the Jurassic rocks unconformably overlie the Palaeozoic sequence. In the Chitral area of the Karakorams, the Zait Limestone is exposed in the Mastuj Valley.

In the upper Hunza Valley, the Triassic sequence comprises the Borom and Aghil Formations and farther eastward, in the Shaksgam–Baltoro region of the Karakorams, Triassic Urdok Conglomerate, Chikchiri Shale and Aghil Limestone are exposed.

Sulaiman–Kirthar Fold Belts

Wulgai Formation (Williams 1959): This Formation is exposed in the Balochistan ophiolite-and-thrust belt in Zhob, Kalat and Khuzdar regions. It has been encountered also in wells drilled in the Lower Indus Basin. It consists of indurated dark grey mudstone and shale with intercalations of thin limestone and calcareous mudstone and sandstone. Previously these rocks were described under different names– Alozai Group (Wulgai section near Muslimbagh), Shirinab Formation (Khuzdar section) and Winder Group (Lasbela area) by Hunting Survey Corporation (1960). The thickness of the Wulgai Formation in the type section at Wulgai is estimated at 1,180 m (Williams 1959), whereas according

Table 5.6. Correlation of the Mesozoic sequences.



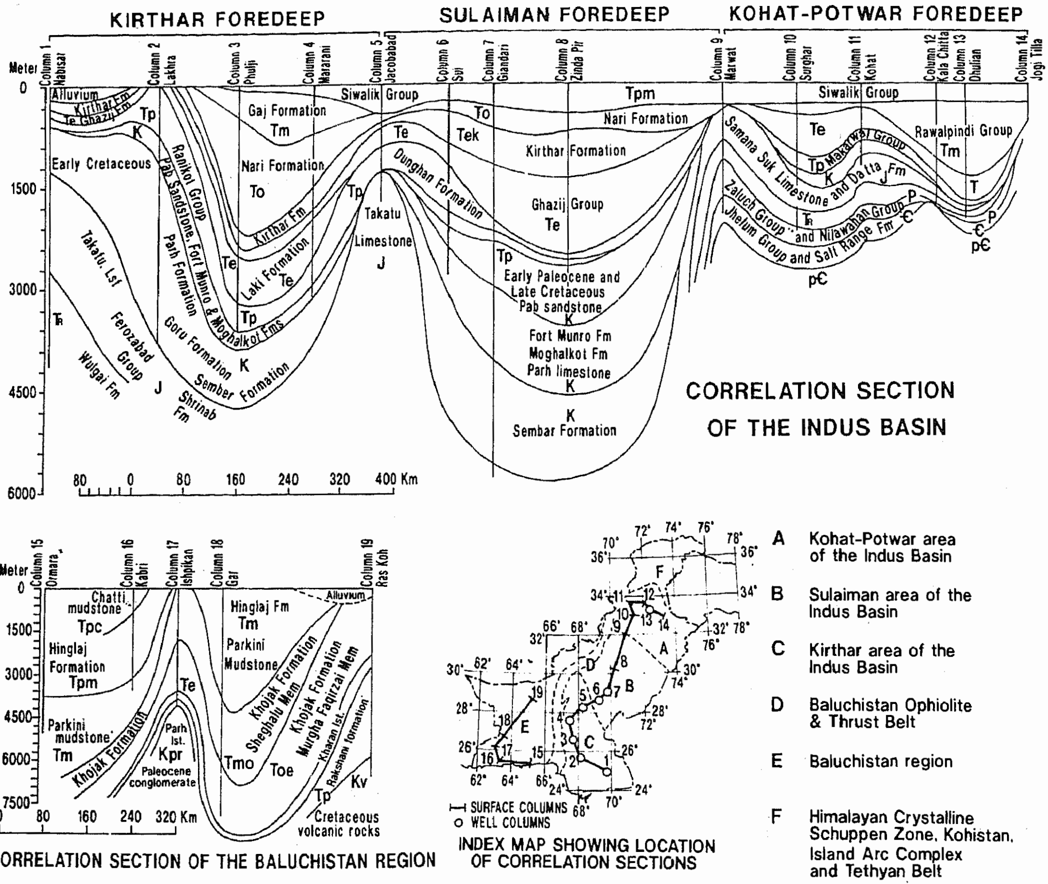


Fig. 5.5. Correlation of the sedimemtary sequences of the Indus Basin and western Balochistan. (Modified from Bakr et al. 1964).

to Sokolov and Shah (1965) the total thickness is 985 m. The lower contact of the Wulgai Formation with Permian beds is tectonically disturbed in the Wulgai area but it is conformable in the Shirinab section (HSC 1960, Sokolov and Shah 1965). Its upper contact with the Jurassic sequence is transitional (Fatmi 1977). The Formation is sparsely fossiliferous. *Columbites sp.*, has been reported from its lower part whereas *Halorites sp.*, *Jovites sp.*, *Pararcestes sp.*, *Arietoceltites sp.*, *Halobia sp.* and *Monotis salinaria* have been obtained from the upper part. According to Fatmi (1977) these fossils indicate an Early to Late Triassic age (see Chapter 4 also).

Himalayan Fold-and-Thrust Belt

Mianwali Formation: The rocks included in the Mianwali Formation (Kummel et al. 1966) were previously known as "Mianwali Series", "Ceratite beds" and with other names (Gee 1945, Pascoe 1959, Waagen 1879, 1895). This Formation consists mainly of marl, limestone, sandstone, siltstone and dolomite. It is exposed in the Khisor Range, Surghar Range and the western part of the Salt Range. Farther eastward it wedges out as a result of erosion that preceded Tertiary deposition. This Formation has been divided into

Kathwai Member, **Mittiwali Member** and **Narmia Member** (Kummel et al. 1966).

The Mianwali Formation overlies the Chhidru Formation of Permian age (Gee 1989) conformably, though a para-unconformity is indicated on palaeontological and palynological grounds (Kummel et al. 1970). It is overlain by the Tredian Formation in the Salt Range and by the Chak Jabbi Limestone in the Kalachitta Range. According to Fatmi (1977) ammonoid fossils are abundant, the more significant ones being *Owenites*, *Anakashmirites*, *Meekoceras*, *Ophiceras*, *Glyptophiceras*, *Proptychites*, *Gyronites*, *Kymatites*, *Kingites* and *Ambites*. The brachiopods include *Lingula*, *Linoproductus*, *Spirigerella derbyi*, *Orthothetina*, *Enteletes*, *Orthotichia* and *Martinia* (Kummel and Teichert 1966). Forams, ostracods, crinoids and conodonts are also found in the Mianwali Formation (Iqbal et al. 1980); it has been assigned an Early Triassic (Scythian) age.

Tredian Formation: It consists of a nonmarine sequence of thin- to thick-bedded, variegated, micaceous sandstone with ripple marks and slump structures, interbedded with shale (**Landa Member** of Kummel 1966) in the lower part which is 19 m–29 m thick. The upper part of the Formation contains massive to thick-bedded sandstone (38 m –59 m), interbedded with thin dolomite beds in the upper part (the **Khatkiara Member** of Danilchik et al. 1967). Formerly this Formation was known as "Kingriali Sandstones" (Gee 1945) but later Gee (in Kummel 1966) renamed it as the Tredian Formation. It is widely exposed in the Khisor Range and the Salt Range. The Tredian contains spores, pollens and wood fragments; the significant ones are *Aratrisporites paenulatus*, *Calamospora*, *Cyciogranisporites arenosus*, *Falcisporites stabilis* and *Platysaccus queenslandi* (Balme 1970). This Formation overlies the Early Triassic Mianwali Formation conformably and grades into the overlying Kingriali Formation. The Tredian Formation is therefore regarded as Middle Triassic in age.

Chak Jabbi Limestone: In the Kalachitta Range and in parts of Hazara District, the Mianwali Formation is overlain conformably by thin- to medium-bedded, grey, unfossiliferous sublithographic limestone – the Chak Jabbi Limestone (Fatmi 1972). The Chak Jabbi Limestone grades into the overlying Kingriali Formation. Its stratigraphic position is thus similar to the Tredian Formation. The Chak Jabbi Limestone is considered as Middle Jurassic in age.

Kingriali Formation: Previously known as "Kingriali Dolomite" (Gee 1945), this Formation crops out in the Trans-Indus Range, Salt Range, Kalachitta, Kohat and in Southern Hazara. It contains thin- to thick-bedded or massive, grey dolomite and dolomitic limestone with dolomitic shale and marl. It is 76 m–106 m thick. Fossils are rare though some brachiopods, bivalves and crinoids have been reported (Fatmi 1977). The lower contact of the Kingriali Formation is transitional with the Mianwali Formation or Chak Jabba Formation and it has a disconformable contact with the overlying Jurassic Datta Formation. It is therefore believed to be Late Triassic in age.

Kashala Formation: In the Mardan–Swat region of the Himalayan fold-and-thrust belt, the Triassic sequence is mainly included in the Kashala and Nikanai Ghar Formations which form a part of the Alpurai Group of DiPietro (1990). The Kashala Formation consists of calcareous-schist, garnet-actinolite schist and schistose marble with massive white to grey marble. The Formation is about 4,000 m thick and has a sharp contact with the underlying amphibolites of the Marghazar Formation. Marble beds near the middle of

Kashala Formation have yielded Upper Triassic (Carnian) conodont– *Neogondolella polygnathiformis angusta* (Pogue et al. 1992).

Saidu Schist: South of the MMT, in the vicinity of Saidu and elsewhere in the Swat area, and in thrust contact with melanges of the Indus Suture zone, there is a thick sequence of grey to dark-grey calcareous and pelitic schist which has been named Saidu Schist by Kazmi et al. (1986). Most of this schist is graphitic though chlorite schist is locally present. The metamorphic grade is upper greenschist to lower amphibolite. The Saidu Schist is included in the Alpurai Group of DiPietro (1990, 1991) and overlies the garnetiferous schist and marble of the Kashala Formation (Table 5.4). About 30 km south of Saidu, the Saidu Schist is replaced by the Nikanai Ghar Formation which overlies the Kashala Formation in this locality. Like the Nikanai Ghar Formation the Saidu Schists may also be Upper Triassic to Jurassic in age.

Nikanai Ghar Formation (Palmer-Rosenberg 1985; Ahmad et al. 1987): It consists of white to grey, thick-bedded to massive crystalline marble and dolomitic marble with subordinate thin beds of calcareous schist, graphitic schist and quartzite. The marble beds contain poorly preserved fossils. Fossils collected by Palmer-Rosenberg have been identified as "some type of colonial metazoans" Ordovician or younger in age, whereas presence of a "questionable *palaeoniscoid* fish teeth" (Late Devonian to Jurassic) has been reported by Pogue et al. (1992). The Nikanai Ghar Formation has a fault contact with the underlying Kashala Formation and is tentatively assigned a Late Triassic to Jurassic age (Pogue et al. 1992).

Karakoram

Zait Limestone: Triassic rocks have been reported from Chitral. A thick sequence of massive black dolomitic limestone exposed in the Mastuj Valley has been named Zait Limestone by Talent et al. 1979. According to Tahirkheli (1982) this limestone has an extensive distribution and overlies Early Devonian sandstone. Conodonts (*Anchignathodus typicalis, Anclignathodus isarcicus, Lonchodina triassica, L.inflata, Prionodian mulleri* and *Ozarkhdina sp.*) have been reported from this Limestone and it is Triassic in age (Talent et al. 1979, Tahirkheli 1982). Also see Lun Shale.

Borom Formation: In the Chapursan Valley of upper Hunza, a series of platy and dark limestone with ribbons of black chert overlie the Permian Kundil Formation and has been named Borom Formation (Gaetani et al. 1990a). The upper part of the sequence contains a carbonate breccia with black limestone pebbles. This Formation contains bivalves, ostracods and radiolarians. Some of the fossils obtained from it include *Gondollela regale, G. foliata inclinata, Metapolygnathus diebeli,* and *Daonella indica* (Gaetani et al. 1990a). This assemblage indicates a Middle Triassic (Ladinian) age.

Aghil Formation: Desio (1964) described this shallow water, marine sequence of grey and white limestone and dolomite with Megalodon and Dicerocardium from the Baltoro Region and named it after Aghil Pass. Above the Baltoro and Abruzzi Glaciers, the limestones of the Aghil Formation form some of the highest peaks of the Karakoram–Hidden Peak (8,068 m), Gasherbrum Peaks II (8,035 m), III (7,925 m) and IV (7,925 m). This Formation is also exposed in the Sost Unit of Upper Hunza (Table 4.8, Fig. 5.4). In this region the Formation consists of thick-bedded or massive limestone and dolomite. The limestone is black and is comprised of wakestones and mudstones. The dolomite is whitish and cyclothemic with large megalodons and stromatolithic layers. Other fossils include *Aulotortus sp, Ophtalmidium sp, Miliolipora sp, Thaumatoporella sp, Multispira, Involutina, Glomospira* and *Glomospirella*. The upper part of the Formation is rich in *Lithiotis* (Gaetani et al. 1993, 1990a).The Aghil Formation ranges from Late Triassic to Early Jurassic.

Urdok Conglomerate: In the Baltoro region, the Aghil Formation is underlain by a conspicuous multicoloured conglomerate described as Urdok Conglomerate by Desio (1963). It overlies the Permian Shaksgam Formation (Table 4.8) and according to Searle (1991) crops out along the upper Abruzzi Glacier near the base of Gasherbrum peaks IV, V and VI, in the Urdok Glacier basin, Siachen area and along Shaksgam Valley. The Conglomerate is up to 500 m thick and includes well rounded clasts of limestone, marble, chert and dolomite with little matrix and lack of granite and volcanic clasts. It is characterised by sediment fill and cross-bedding. Because it is sandwiched between the Triassic Aghil and Permian Shaksgam Formation its age is likely to be Early Triassic (Searle 1991).

Gaetani et al. (1990b) however consider the Urdok Conglomerate to be Cretaceous because it contains sandstone clasts which in their opinion are similar to the Jurassic Marpo Sandstones and clasts of foliated gneisses which they think are similar to K2–Sarpo Laggo Gneisses of Middle Cretaceous age. If this assumption is correct, the Urdok Conglomerate would be younger than the Marpo Sandstone and the Cretaceous gneisses and may be Late Cretaceous in age.

Chikchi-ri Shales: A sequence of black, brown, green and sometimes red arenaceous shales and greenschists with thin beds of limestone, tuff and lava flows have been described and mapped by Desio (1963, 1964) in the Chikchi Range and the valley of the Singhie Glacier. These Shales overlie the Permian Shaksgam Formation and occur below the Late Triassic Aghil Limestone. According to Desio (1979) these Shales are exposed over a large area of eastern Karakoram, from Shaksgam Valley and Karakoram Pass to the Shyok Valley. Several fossils have been found in these Shales. The Chikchi-ri Shales and Urdok Conglomerates pass laterally into one another.

Gaetani et al. (1990b) have described a Lower Triassic sequence of 30 m thick marls and shales overlain by 50 m thick bedded limestone with conodont *(Neospathodus homeri)* and foraminifera *(Meandrospira pusilla)*. According to them this sequence “evolves into a huge polygenetic brecciated body, including boulders from thin-bedded Triassic limestone”. This unnamed lithostratigraphic sequence of Gaetani et al. (1990b) is probably another lateral facies of the Urdok Conglomerate/Chikchi-ri sequence mentioned earlier.

Jurassic

The Jurassic sedimentary rocks largely constitute a thick (820 m–3,000 m) sequence of marine pericratonic shelf deposits consisting of limestone, shale and sandstone with subordinate dolomite and ferruginous beds. They form a part of the platform cover in the entire Indus Basin (Fig. 5.5). In the Kirthar–Sulaiman region, the Jurassic outcrops are largely restricted to the anticlinal cores whereas in the Balochistan ophiolite-and-thrust belt and the Himalayan fold-and-thrust belt, they form extensive thrust blocks or sheets. In the Karakoram the Jurassic metasediments occur in antiformal thrust stacks or thin thrust slices.

Indus Platform and East Balochistan Fold-and-Thrust Belt

In the lower part of the Indus platform, the Sulaiman–Kirthar fold belt and the Balochistan ophiolite-and-thrust belt, the Jurassic rocks attain their maximum thickness (3,000 m) and they consist mainly of marine limestones and shales (Shirinab Formation and its equivalent Ferozabad Group, Takatu Formation and Mazar Drik Formation). There is an apparently gradual transition from a dominant shale lithology of the underlying Triassic rocks to the thin bedded intercalations of limestone and shale of the Jurassics (Williams 1959, HSC 1960).

Shirinab Formation: In the lower Indus Basin this Formation is probably the earliest Jurassic exposed rock unit, ranging in thickness from about 1,500 m to 3,000 m (HSC 1960, Fatmi 1977). It consists of interbedded limestone and shale. According to Fatmi (1977), it is transitional downwards and grades into the shaly Triassic Wulgai Formation. In fact the "Shirinab Formation" of HSC (1960) includes undifferentiated Jurassic, Triassic and even Permian rocks. The Shirinab Formation, as presently recognised, is Early Jurassic in age and is exposed in the Kalat, Quetta, Zhob and Loralai Districts of Balochistan. Equivalent rocks in the Khuzdar District have now been renamed **Ferozabad Group** (Fatmi et al. 1986).

The Shirinab Formation has been subdivided into Spingwar Member, Loralai Member and the Anjira Member (Williams 1959). The basal **Spingwar Member** includes a sequence of interbedded grey to black crystalline limestone about 600 m to 1,800 m thick, which is oolitic and shelly in places and contains calcareous shale with sandstone interbeds in its lower part. The type section is at Spingwar, 35 km northwest of Loralai. According to Williams (1959), Spingwar is considered to be Liassic in age on the basis of ammonites identified by the British Museum from the type locality.

The **Loralai Limestone** overlies the Spingwar Member conformably and is dominantly thin- to medium-bedded, dark grey to black, hard, crystalline limestone. The type section is at Zamarai Tangi. The Loralai Limestone is overlain conformably by the Anjira Member, in the area south of Kalat whereas in the locality of the Zamarai Tangi it is overlain by Cretaceous rocks with an angular unconformity. It is considered Liassic in age owing to the fact that Toarcian to Bajocian fossils have been found in the overlying Anjira Member, whereas the underlying Spingwar contains Liassic ammonites.

The **Anjira member** consists of dark grey, thin, hard, porcelaneous to sub-lithographic limestone and interbedded softer argillaceous limestone and splintery mudstone. Its thickness ranges from about 100 m to 400 m and its outcrops are apparently restricted to the Kalat–Khuzdar area. The type section is located about 12 km east of Anjira. The ammonites collected from the Anjira Member range in age from Toarcian to Bajocian (Williams 1959). This unit is overlain disconformably by the Upper Jurassic to Lower Cretaceous Sember Formation or the Lower Cretaceous Goru Formation.

Fatmi et al. (1986) have revised the stratigraphic nomenclature of the Jurassic sequence in the vicinity of Khuzdar and have introduced the name Ferozabad Group which was previously designated by Fatmi (1977) as the Shirinab Formation. The name "Zidi" was previously used by HSC (1960), but discarded by Fatmi (1977) and the National Stratigraphic Committee.

Ferozabad Group: The main difference between the Shirinab and the Ferozabad lithostratigraphic units is that the lower part of the Ferozabad Group contains a greater influx of clastic sediments and consists largely of grey to brown, micritic limestone, sandy limestone and sandstone with shale interbeds. More than 248 m of these sediments are exposed in the type section 17 km WNW of Khuzdar. The upper sandstone contains poorly preserved molluscan shells, but it appears that no diagnostic fossils have yet been found in the lower unit. According to Fatmi et al. (1986) the lower age limit of the Ferozabad Group is uncertain and doubtfully may be Triassic to Early Jurassic.

The lower unit of the Ferozabad Group is transitionally overlain by the Middle unit, which consists of thick to massive, grey, mottled, oolitic, coquinoid or micritic limestone with abundant bioturbated beds, and marly limestone intercalations. This unit is about 387 m thick in the Ferozabad section and has yielded abundant fragments of *Pecten weyla, Girvillia*, *Isocrines* and *Spiriferina sp.*

The Middle unit grades into the upper unit, which is about 300 m thick and consists of brown to grey, thick-bedded limestone with marl and shale. *Spiriferina, Terrebratula, Montlivaltia,* and Lower Toarcian to Lower Bajocian fossils like *Bouleiceras, Protogrammoceras, Dactylioceratids, Cenoceras, Hammotoceras, Phymatoceras, Phylsiogrammoceras, Graphoceras, Shaeroceras, Chandroceras, Hammotoceras* and *Belemnopsis* have been identified (Fatmi et al. 1986).

A paraconformity separates the Ferozabad and overlying Sembar Formation. The Ferozabad Group (and its sub-units) have been correlated with the Shirinab Formation.

Takatu Limestone: In the Kalat, Quetta, Sibi and Loralai Districts a massive to thick-bedded dark- to light-grey or cream coloured, sub-lithographic to oolitic limestone 750 m to 1,800 m thick conformably and gradationally overlies the Shirinab Formation. At places the limestone is reefal or biohermal. It was named Chiltan Limestone by HSC (1960) and the name was adapted by Fatmi (1977). Following Williams (1959) it has now been renamed Takatu Limestone (Shah 1987). The type section is along Dara Manda south of Bostan. Only poorly preserved or unidentifiable fossils have been found in this Formation and owing to its stratigraphic position between the Lower Jurassic Loralai Member of the Shirinab Formation and the overlying Callovian to Bathonian Mazar Drik Formation, the Takatu Limestone is considered Middle Jurassic in age (Fatmi 1977).

Mazar Drik Formation: In the Sibi, Kalat and Khuzdar Districts, the Takatu Limestone is overlain conformably by the Mazar Drik Formation (Arkell 1956), which consists mainly of interbedded grey limestone and dark shale. The type area is Mazar Drik in Marri Hills where the Formation is about 30 m thick. A number of ammonites, including *Macrocephalites, Dolikephalites, Indocephalites, Pleurocephalites, Indosphinctes, Paralicidia, Bomburites, Choffatia, Bullatimorphites bullatus, Clydoniceras* have been found in this Formation (Arkell 1956, Fatmi 1969, 1977) and it is Callovian to Late Bathonian in age. The Mazar Drik Formation is overlain disconformably by the Sembar Formation.

Upper Indus Platform and the Himalayan Fold-and-Thrust Belt

In the northern part of the Indus Platform, the Salt Range and the Trans-Indus Ranges, Kohat–Potwar Plateau and the Kalachitta–Margalla fold-and-thrust belt, the Jurassic sequence is comprised of the Datta, Shinwari and Samana Suk Formations, and the Lower part of Chichali Formation. Compared with the lower part of the basin, the Jurassic sequence in the Upper Indus Basin is much thinner (820 m) and consists mainly of arenaceous and argillaceous sediments of continental origin that grade upward into marine calcareous and argillaceous rocks (Fatmi 1977). Here, the Jurassic unconformably overlies the Triassic Kingriali Formation and there is an Upper Bathonian - Middle Callovian disconformity within the Jurassic sequence (Tables 4.3 and 5.6).

Datta Formation (Danilchik 1961): Previously known as "Variegated stage" (Gee 1945), this Formation contains the earliest Jurassic rocks in the region and consists largely of variegated sandstone, shale, siltstone and mudstone, at places interbedded with fire clay. The type section is in Datta Nala in the Surghar Range. Its thickness ranges from 150 m to 400 m, though northwards in Hazara it is reduced to 10 m (Fatmi 1977). The Datta Formation overlies the Kingriali Formation unconformably. In the Hazara area it rests unconformably on Precambrian, Paleozoic or Triassic rocks. The Datta Formation is believed to be Early Jurassic inasmuch as it underlies the Toarcian Shinwari Formation.

Shinwari Formation (Fatmi 1977): This Formation has a transitional contact with the underlying Datta Formation and consists mainly of thin-bedded grey limestone, nodular marl, shale and sandstone. Current bedding and ripple marks are present. Its thickness varies from 12 m in Kalachitta to over 400 m in Samana Range. It has a transitional contact with the overlying Samana Suk Formation. Fossils including *Bouleiceras, Terebratula, Spiriferina, Montivaltia, Pholadomya, Zeilleria, Pecten,* and *Lima* have been reported from the Shinwari Formation and it is believed to be Early Jurassic (Toarcian) to Middle Jurassic in age.

Samana Suk Formation (Davies 1930): It includes medium to thick-bedded, grey, oolitic, and at places, shelly or dolomitic limestone with interbedded marl and calcareous shale. It contains bivalves, gastropods, crinoids, brachiopod and ammonites: *Reineckeia ancepto, Choffatia sp., Obtusicostites sp., Hubertoceras sp., and Kinkeliniceras sp.* (Fatmi 1977). The Formation is thus Middle Jurassic in age. Its thickness ranges from about 66 m in Chichali Pass to 366 m. It is disconformably overlain by the Chichali Formation.

Kohistan Magmatic Arc

Kalam Group: In the Kohistan magmatic arc terrain of Dir and Swat Districts a sequence of slates, limestones and quartzites intruded by diorites have been described and named Kalam Group by Matshushita et al. (1965). The limestone, is thin-bedded, fine-textured and 35–45 m thick (Tahirkheli 1982). It has yielded *Rhabdophyllia sp.,* (Karig 1981), on the basis of which this Group may be assigned Middle Jurassic to Cretaceous age (see section on Cretaceous).

Karakoram

Unfossiliferous metasediments in the Karakoram believed to be of Jurassic age, occur in Chitral whereas a fossiliferous Jurassic sequence has been reported from Upper Hunza and Shaksgam region in the Northern sedimentary belt.

Chitral Slate: In the Chitral region of the Tethyan Belt, a 3–4 km thick sequence of monotonous, soft, dark-grey shale with thin ash beds and subordinate fine sandstone is exposed west of Chitral and has been named Chitral Slate (Desio 1959, 1966, 1975, Calkins et al. 1981, Tahirkheli 1982, Pudsey et al. 1985). According to Pudsey the slate is largely without bedding but it shows a complex structure. The sandstone may form as much as one third of the sequence and it is largely in the form of grey, fine-grained, thin- to medium-bedded quartzite though isolated fine-grained greywackes are also present sporadically. Parallel lamination and graded bedding are common, slump structures are present and there is no bioturbation. No fossils have yet been found in the Chitral Slate and its base is not exposed.

In the Kunar River Valley, southwest of Chitral town, the Chitral Slate forms an anticline, the northwestern limb of which is overlain by the Cretaceous Krinj Limestone. Its southeastern limb is overlain by the Koghozi Greenschist, which is conformably overlain by the unfossiliferous Gahirat Limestone (Desio 1955, 1966, 1975, Pudsey et al. 1985).

The **Koghozi Greenschist** comprises fine-grained green chlorite-epidote-quartz schists interbedded with subordinate, thin acidic layers with abundant quartz. Bedding is not very clear and the cleavage tends to anastomose (Pudsey et al. 1985).

The **Gahirat Limestone** is grey, massive, and contains coarsely crystalline marble as much as 3 km thick. Southwest of Chitral, a zone of grey micaceous phyllite is seen in the middle of the limestone. According to Pudsey et al. (1985) the phyllite is either infolded or introduced by faults.

Being unfossiliferous, there is no direct evidence for the age of the above sequence of Chitral Slate, or of the Koghozi Schist and the Gahirat Limestone. Tipper (in Fermor 1922) reported a *Spirifer* and *Dielasma* from Chitral Slate, which indicated a probable Permian age. Desio (1959, 1966, 197) placed these Formations in the Triassic. Calkins et al. (1981) have the Chitral Slate interpreted as of Cretaceous age. Pudsey et al. (1985) have placed the Chitral Slate and Koghozi Greenschist in the Jurassic and the Gahirat Limestone in the Cretaceous, though they concede that the Chitral Slate may range from Cretaceous down to Permian.

Tahirkheli (1982) on the other hand has correlated the Chitral Slate with the Baltit Group and the two units laterally merge without any tectonic or stratigraphic break. According to him Chitral Slate underlies the Lower Devonian sequence in the Mastuj Valley and across the border in Afghanistan. He has therefore proposed a Precambrian to Early Paleozoic age for the Chitral Slate.

Aghil Limestone: This Formation crops out in the Upper Hunza and Shaksgam–Baltoro region and ranges from Upper Triassic to Early Jurassic. It has been described earlier under the section on Triassic.

Ashtigar Formation: In the upper Hunza region (Fig. 5.6), the Aghil Limestone is conformably overlain by the Ashtigar Formation (Gaetani et al. 1993). The lower boundary of the Ashtigar is sharp and shows transition from dark laminated Aghil Limestone to black shales intercalated with calcarenites and volcanic debris. The Ashtigar Formation is largely comprised of a terrigenous sequence of thick, dark grey marly shales, interbedded with fine-grained turbidites. Its upper boundary with the overlying Yashkuk Formation is erosional and is characterised by a 5 m thick conglomerate with abundant clasts of Aghil Limestone. Age data from the Ashtigar Formation is lacking, though fragments of *Trigonia*-type bivalves have been reported. In the northern part of Chapursan Valley the Ashtigar is missing (Fig. 5.6) and is apparently replaced by fine-grained, dark grey packstones/ wackestones (unit 2 A of Gaetani et al. 1993). The fossils in this unit comprise nannoflora (*Mitrolithus jansae, Lotharingius huffii* and *Schizosphaerella punctulata)*, brachiopods (*Zeilleria sp, Spiriferina sp)*, crinoids and foraminifera. This fauna points to a Liassic age.

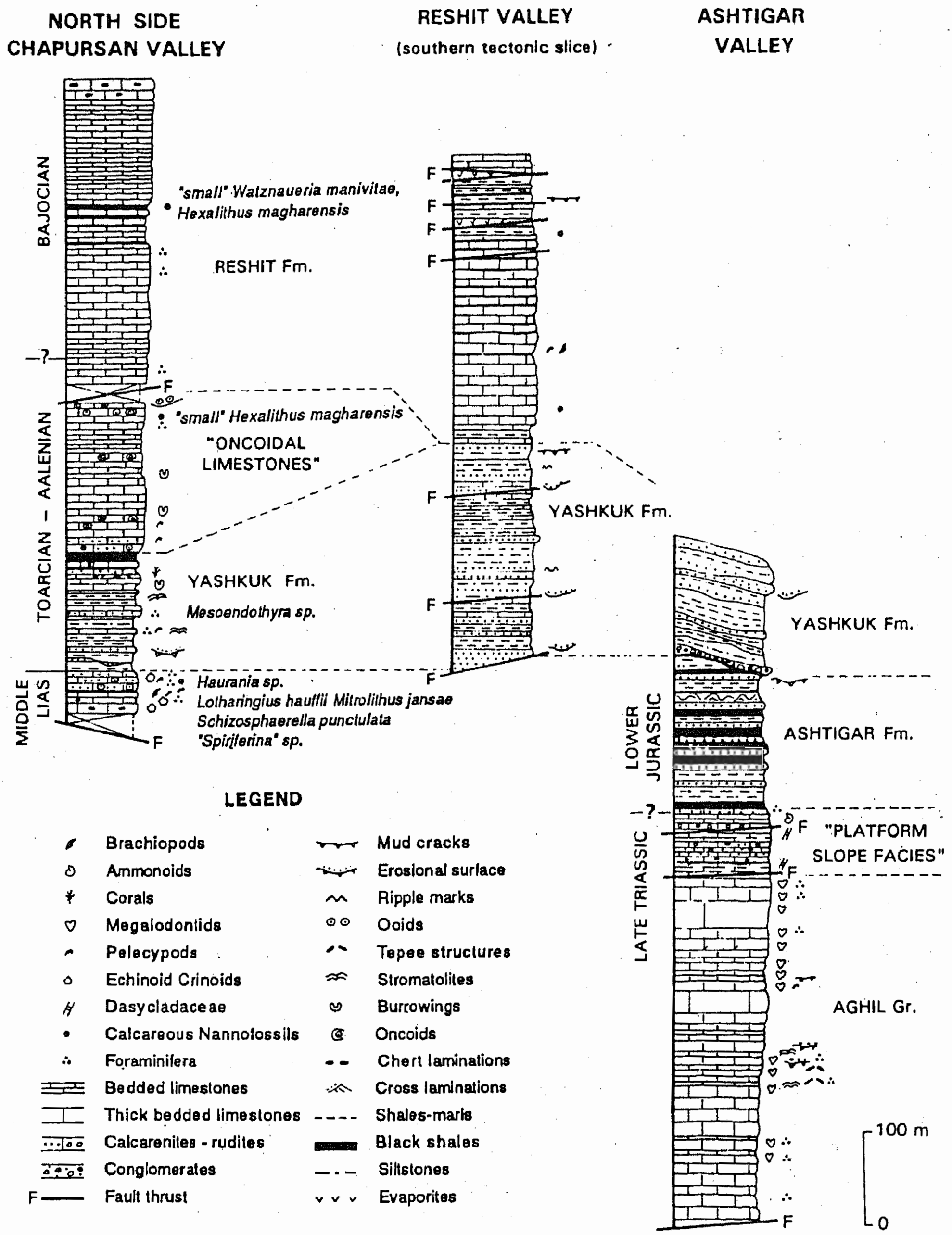


Fig. 5.6. Stratigraphic columns showing the Jurassic sequence in the Chapursan Valley, Upper Hunza (from Gaetani et al. 1995).

Yashkuk Formation (Gaetani et al. 1990a): This Formation overlies the scoured upper beds of the Ashtigar Formation. The lower part of the Yashkuk Formation consists of fine-grained, reddish, cross-laminated, ripple marked sandstones overlain by very fine-grained red to green sandstones with raindrop prints, ripples and mud cracks. This lower sequence is interpreted as a meander-bed deposit. It is overlain by pelites, dark marly limestone, oolitic calcarenites and oncoidal and cherty limestones. This upper carbonate unit suggests, marine transgression and deposition in open lagoon. Gaetani et al. (1990a, 1993) have reported benthic foraminifera *(Trocholina, Lenticulina, Saccocoma, Trochammina, Ammodiscus, Frondicularia, Involutina, Ammobaculites)* and algae *(Solenoporacea and Dasycladacea)*, nannofossils *(Schizosphaerella punctulata, Hexalithus magharensis,* and *Watznaueria manivitae)*, ostracods and blue-green algae from this Formation. According to Gaetani et al. 1990a), the Yashkuk Formation ranges in age from Middle to Upper Liassic.

Reshit Formation: In the Chapursan Valley, the Middle Jurassic sequence is intensely faulted (Fig. 5.4) which renders stratigraphic correlation difficult. A monotonous, 300 m thick sequence of dark grey, cherty mudstones and wackestones overlies the Yashkuk Formation with a fault contact (Fig. 5.6). This Formation contains foraminifera, ostracods, crinoids, corals, echinoids, brachiopods and ammonites (Gaetani et al. 1990a). Identified fossils include *Lenticulina, Spirillina, Trocholina, Protopeneroplis, Saccocoma* and *Kallirhynchia* and a late Middle Jurassic age in indicated.

Tekri Formation: In the Shaksgam region of eastern Karakoram a sequence of grey mudstone/wackestones with rare dark chert, oncolitic limestones and bioclastic packstones has been named Tekri Formation by Gaetani et al. (1990b). It has a gradational contact with the underlying Aghil Limestone and contains sponges, gastropods, and ostreids. The upper part contains foraminifera *(Lenticulina, Protopeneroplis, Haurania, Mesendothyra, Dentolina)* which indicate a Middle Jurassic age.

Marpo Sandstone: (Gaetani et al. 1990b): This unit overlies the Tekri Formation in the Aghil Pass region and Shaksgam Valley and consists of 20–100 m thick red sandstones and siltstones. The basal part of the sequence comprises ostracod bearing grey marls interbedded with red siltstones overlain by polymict conglomerates. The upper part of the sequence is largely comprised of red shaly sandstone. Marpo Sandstone has been placed in the Middle Jurassic by Gaetani et al. (1990b).

Bango-La Formation: Named by Desio and Fantini (1960), this Formation crops out in the Shaksgam Valley and it is comprised of marly sandstones and dark-coloured limestones. It contains corals and ammonites and has been assigned an Upper Jurassic age by Desio and Fantini (1960) and Gaetani et al. (1990b). According to the latter authors, this Formation conformably overlies the Marpo Sandstone.

Cretaceous

Cretaceous sedimentary rocks are exposed in the Indus Basin, the Himalayan fold-and-thrust belt, the Kohistan magmatic arc, the Karakoram and in the Chagai magmatic arc. At many localities the Cretaceous sequence contains volcanic rocks, obducted masses of melanges and ophiolites and igneous intrusions.

In the Kirthar–Sulaiman region of the Lower Indus Basin, except for local disconformities, there is a complete sequence of the Cretaceous ranging from Late Tithonian through Neocomian to Maestrichtian. This sequence comprises over 2,000 m of fossiliferous marine shale, carbonate, and clastic sediments (Sembar and Goru Formations, Parh Limestone, Mughal Kot, Fort Munro, Bibai and Moro Formations and Pab Sandstone). In the Balochistan ophiolite-and-thrust belt, the sequence consists largely of thrust blocks and sheets which include extensive sheets of volcanic rocks and obducted masses of melanges and ophiolites.

In the Kohat–Potwar region of the Upper Indus Basin, the lower part of the sequence of Cretaceous age consists of marine sandstone and shale (Chichali and Lumshiwal Formations) with limestone (Kawagarh Formation) in the Upper Cretaceous. In this sequence there are disconformities; strata of Cenomanian, Turonian and Maestrichtian age are missing (Table 5.6).

In the northern areas of Pakistan, the Kohistan Complex (see Chapter 6) contains a thick Cretaceous sequence of volcanics and metasediments (Dir and Kalam Groups, Gwach, Purit and Drosh Formations, Shamran Volcanic Group, Rakaposhi Volcanic Complex, Burji, Katzarah and Bauma–Harel Formations). Farther northward the metamorphosed and highly tectonised and imbricated rock sequence of the Karakoram fold-and-thrust belt includes a series of Cretaceous crystalline marble, phyllite, slate and shale (Gahirat Limestone, Krinj Limestone, Reshun Formation). The Northern sedimentary belt of the Karakoram contains a highly tectonised Cretaceous sedimentary sequence of limited extent comprised of Tupop and Darband Formations. A relatively thick sequence of volcanic and sedimentary rocks also crops out in the Chagai–Ras Koh region of Balochistan (Sinjrani Volcanic Group and Humai Formation).

Lower Indus Platform and East Balochistan Fold-and-Thrust Belt

Sembar Formation (Williams 1959): This is the lowermost unit of the Cretaceous sequence in the Kirthar–Sulaiman region, consisting of black shale interbedded with siltstone and nodular, argillaceous limestone. The shale and siltstone are commonly glauconitic. The Formation is 133 m thick in type area (Sembar Pass) and 262 m in the Mughal Kot section. It has a gradational contact with the overlying Goru Formation though at places an unconformity has been reported by Williams (1959). The fossils most commonly found in the Sembar Formation are belemnites *Hibolithes pistilliformis, H. subfusiformis,* and *Duvalia sp.* It is mainly Neocomian in age. According to Fatmi (1977) it may extend to the Late Jurassic..

Goru Formation (Williams 1959): This Formation is composed of interbedded limestone, shale and siltstone. The lower part is more shaly and consists of very thin-bedded, light-coloured limestone interbedded with thin to irregularly bedded, calcareous, hard, splintery grey to olive-green shale. The upper part is largely thin-bedded, light-coloured porcellaneous limestone with subordinate shale. It grades into the overlying Parh Limestone. At the type locality (Goru village) the Formation is 536 m thick. Northwards in the Quetta area the thickness is reduced to about 70 m whereas southwards, in some of the wells drilled, 1,200 m to 2,700 m of Goru sediments have been encountered.

The Goru Formation contains belemnites *(Hibolithes sp.)* and foraminifera (*Globigerinelloides algeriana, G. breggiensis, G. caseyi, Ticinella roberti, Gavelinella lorneiana, Rotalipora ticinensis, R. appennenica, R. brotzeni, R.reichli, Praeglobotruncana stephani* and *Planomalina buxtorfi* (Fritz and Khan 1967). It is Cretaceous in age (Albian ? to Cenomanian ?; Williams 1959).

Parh Limestone (Blanford 1879, Vredenburg 1909, Williams 1959): This is a very uniform, distinct and persistent rock formation and is exposed extensively in the Kirthar–Sulaiman region. It is thin-bedded, light-grey, white or cream coloured with a persistent pink, purple to maroon coloured band of interbeds of variegated shales and marls (Photo. 29).

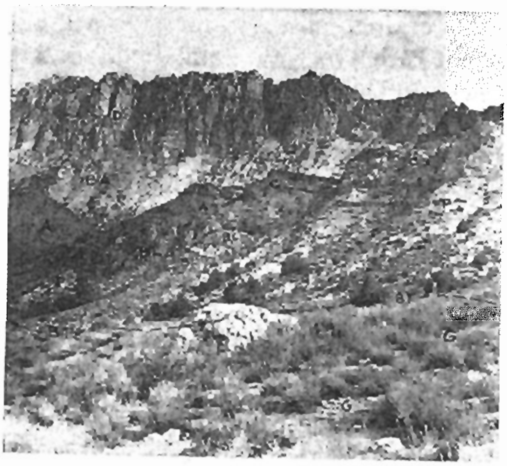


Photo. 29. The Bibai Nappe seen east of Kach. The Parh Limestone overlies the Ghazij shales along the Bibai Thrust. The Bibai volcanics are interfingered with Parh Limestone along the contact. The Dungan Limestone forms the ridge at the top. A-Bibai agglomerate BA-Bibai ash and mudstone, BL-Bibai lava flows, BT-Bibai Thrust, C-Bibai conglomerate D-Dungan Limestone, G-Ghazij shales, Parh Limestone (Photo: *A. H. Kazmi*).

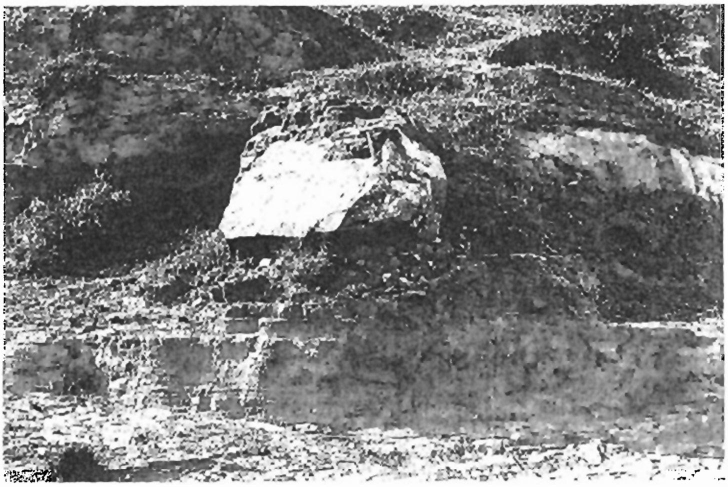


Photo. 30. Quaternary deposits of the Potwar Plateau near Lawrencepur. The upper terrace comprises Potwar Loess, underlain by coarse sandstones. At the contact of the two there is a boulder bed with an erratic– a huge block of granite– seen in the central part of the photo. (Photo: *A. H. Kazmi*).

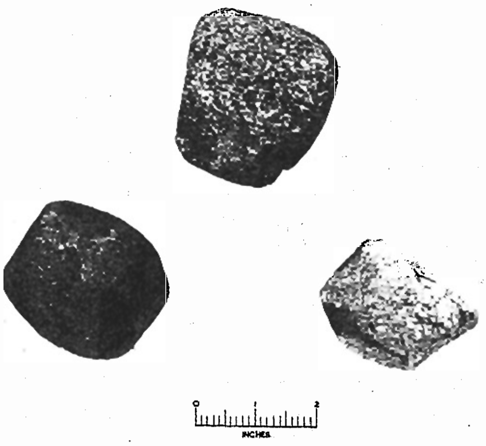


Photo. 31. Facelted cobbles from Tandojam Formation. (Photo: *A. H. Kazmi*).

The limestone is hard, lithographic to porcellaneous or argillaceous, platy to slabby with a characteristic conchoidal fracture. In the type area in the Parh Range, the Parh Limestone is 268 m thick though elsewhere its thickness ranges from 300 to 600 m. It has a conformable and gradational contact with the underlying Goru Formation and is overlain conformably by the Mughal Kot Formation. However, at many localities there is a disconformity at the top (Kazmi 1955, 1979, 1988, Williams 1959). East of Lasbela, the Parh Limestone overlies the Cretaceous Porali agglomerate (DeJong et al. 1979). In the Ziarat–Kaghan area, northwest of Ziarat, the Parh Formation is interlayered with, and overlain by, basalts and tuffs of the Bibai Formation of Late Cretaceous age (Kazmi 1955, 1979, 1988).

The Parh Limestone contains many species of *Globotruncana* (Kazmi 1955, Gigon 1962) and is considered Upper Cretaceous (Barremian to Campanian by Kazmi 1955, 1979; Senonian by Williams 1959).

Mughal Kot Formation (Williams 1959, Fatmi 1977): This Formation consists of grey calcareous shale and mudstone with intercalations of arkosic sandstone and grey argillaceous limestone. In the Sulaiman area, the sandstone is well developed whereas in the Kirthar area the Formation is largely grey, silty, calcareous shale. In one of the oil wells (Dabbo Creek) near Karachi, basalts are present whereas near Kahan village in the Ziarat area the Formation locally contains a thick sequence of conglomerate with boulders and pebbles of basalt (Kahan Conglomerate Member of Williams 1959; Bibai Formation of Kazmi 1955, 1979, 1988). The thickness of the Formation ranges from about 150 m to 1,170 m. The type section is 2 to 5 km west of Mughal Kot Post.

In most areas the Mughal Kot Formation overlies the Parh Limestone unconformably though in the type section and the Karachi region, according to Fatmi (1977), the contact is apparently conformable. *Omphalocyclus sp.*, *Orbitoides sp.*, and *Siderolites sp.* have been reported from this Formation (Williams 1959, Marks 1962) and it is assigned a Campanian to Early Maestrichtian age.

Bibai Formation: In the Ziarat area the sedimentary sequence equivalent to the Mughal Kot Formation has been named Bibai Formation (Kazmi 1955, 1979, 1984). The lower part of the Formation contains basaltic lava flows (at places altered to laterite) and tuffs, locally interlayered with a thick sequence of volcanic boulder conglomerate (Kahan Conglomerate Member of Williams 1959). The upper part consists of mudstones and ash beds with subordinate sandstones (Photo. 29). This Formation is about 30 m to 610 m thick and lies unconformably on the eroded surface of the Parh Limestone; at places its basal part is interlayered with wedges of the Parh Limestone (Fig. 4.17). The Bibai Formation grades into the overlying Dungan Formation. According to Kazmi (1955, 1988) the interlayered Parh Limestone contains *Globotrunca lapparenti* and *G.Linnei* which indicate a Campanian age of the lower part. Sawada et al. (1994) have reported a 71.4 ± 3.4 m.y. age for the volcanics in the lower part of the Bibai. In the upper part of the sequence the following fossils have been reported by Kazmi (1955) near Kach: Foraminifera:– *Omphalocyclus macropora, Orbitoides sp.;* Coelenterata: *Cyclolites sp.;* Gastropoda: *Turritella, Acteonella sp., Ovula expansa* (D Archiac and Heime), *Nerinea quettensis* (Noetling), *Crithium buddha* (Noetling), *Trochus lartetianus* (Leymerie); *Volutillithes dubia* (Noetling), *Volutonomorpha sp., Polinices* (?) Monfort, *Volutilithes sp., Planorbis sp., Conus sp.;* Pelecypoda: *Vola quinqueangularis* (Noetling), *Nucula sp., Spondylus sp., Exogyra aff. pyrenaica;* Echinodermata: *Hemipneustes aff. leymeri;* Cephalopoda:

Schaphites, Turrilites, Baculites binodosus (Noetling). This fauna indicates an Early to Middle Maestrichtian age for the upper part of the Bibai Formation (Kazmi (1955).

Allemann (1979), and DeJong and Subhani (1979) correlated the Porali Agglomerates and Bela Volcanics with the volcanics of Bibai Formation. Other sedimentary-cum-volcanic sequences similar to the Bibai (Bad Kachu Fm., Thar Fm., Gidar Dhor Gr.) have been described by HSC (1960) from the Balochistan ophiolite-and-thrust belt and these range in age from Paleocene to Eocene (Fig 4.14). Siddiqui et al. (1994) have proposed that the Bibai Volcanics are related to a hotspot because of the resemblance of major and trace elements of average MORB and average alkali basalts of Bibai, Hawaii, Ascension and other hotspot related volcanics. (see Chapter 6).

Fort Munro Formation: In the Sulaiman area, a dark-grey, hard, thick-bedded limestone, at places with marly and shaly interbeds, overlies the Mughal Kot Formation conformably and is known as the Fort Munro Formation (Williams 1959, Fatmi 1977). In the Kirthar area, this limestone is grey, brown to cream coloured, thin-bedded in the upper part, and thick-bedded to massive and reefoid in the lower part. Its thickness ranges from 44 m to 248 m and the type locality is in the western flank of Fort Munro anticline. In the Sulaiman area it is overlain conformably and transitionally by the Pab Sandstone or by the Moro Formation. In the Quetta area it is overlain unconformably by the Dungan Formation. The Fort Munro Formation contains *Omphalocyclus macropora, Orbitoides sp., Siderolites sp.* and *Actinosiphon punjabensis* and is considered Late Campanian to Early Maestrichtian in age (Fatmi 1977).

Pab Sandstone (Vredenburg 1909a, Williams 1959): This sandstone is exposed in the Kirthar–Sulaiman region. It rests conformably on the Fort Munro Formation, but in some localities overlies the Parh Limestone unconformably (Shah 1987). It consists mainly of white, cream or brown thick-bedded to massive, cross-bedded, medium- to coarse-grained quartzose, sandstone with intercalations of subordinate argillaceous limestone and shale. The type section is west of Wirahab Nai in the Pab Range where it is 490 m thick, though its thickness ranges from 240 m (Mughal Kot) to 1,000 m (Pab Range). It is missing in parts of Marri Bugti Hills and Quetta–Kalat area.

The Pab Sandstone is overlain conformably by the Moro Formation but at places it is overlain unconformably by the Paleocene Khadro, Rakhshani or Dungan Formations (Fatmi 1977). On the basis of Maestrichtian foraminifera the Pab Sandstone has been assigned a Late Cretaceous age (Vredenburg 1909a, HSC 1960).

Moro Formation (HSC 1960): It is exposed in the Kirthar–Sulaiman region and rests conformably and gradationally on the Pab Sandstone. Where the Pab Sandstone is not developed, the Moro Formation overlies the Fort Munro Formation conformably or rests disconformably on the Parh Limestone. It generally consists of a lower limestone unit, at places with volcanic conglomerate, a middle marly and shaly unit with subordinate sandstone, and an upper limestone unit. It is overlain conformably by the Dungan Formation although at places the Khadro Formation rests on it disconformably (Fatmi 1977). Foraminifera like *Globotruncana aff. G.linnei, Lituola sp., Omphalocyclus macropora, Orbitella media, Orbitoides sp.,* and *Siderolites sp.,* indicate that the Moro Formation is Maestrichtian in age. It is considered as a lateral facies of the Pab Sandstone, and is essentially the same as the Bibai Formation described earlier by Kazmi (1955).

Upper Indus Platform and the Himalayan Fold-and-Thrust Belt

Chichali Formation: In the Kohat–Potwar region the Cretaceous sequence includes the Chichali, Lumshiwal and Kawagarh Formations. The Chichali Formation (Danilchik 1961, Gee 1945) rests disconformably on the Samana Suk Formation and is mostly dark-green glauconitic sandstone with grey silty glauconitic shale in the lower part. The type section is in Chichali Pass. It crops out also in the Kalachitta and Hazara area. Its thickness ranges from 12 m to 70 m. Cephalopods such as *Perisphinctes, Mayaites, Aspidoceras, Physodoceras, Katroliceras, Pachysphinctes, Belemnopsis gerardi, Hibolithes, Pulacosphinctoides, Hildoglochioceras, Provalanginites, Blanfordiceras, Spiticeras multiforme, Subthurmannia, Neocomites, Distoloceras, Olcostephanus* and *Lyticeras*, have been found in this Formation and it has been assigned a Late Oxfordian to Neocomian age. According to Fatmi (1977), the Chichali Formation is mainly of Late Jurassic age in the Hazara area.

Lumshiwal Formation (Gee 1945, Fatmi 1973): It rests gradationally and conformably on the Chichali Formation. The type section is in the Lumshiwal Nala. In the type locality and the Trans-Indus Ranges, this Formation is mostly grey, thick-bedded to massive, current-bedded feldspathic and ferruginous sandstone, but contains silty, or sandy glauconitic shale towards the base. Northward and eastward, the Formation grades into a mostly marine sequence of sandstone, siltstone and shelly limestone. Its thickness ranges from 38 m to 194 m. Abundant moulds of brachiopods, bivalves, gastropods, ammonoids, belemnites and echinoids occur in this Formation in the Samana Range (Cox 1930a). Ammonoids include *Ammonitoceras, Ailoceras, Douvilleiceras mammilatum, Oxytropidoceras sp., Desmoceras sp., Cleoniceras sp., Branchoceras sp.*, and *Lemunoceras sp.* (Spath 1930, Fatmi 1968). Accordingly, the Formation has been assigned Aptian to Early Albian age in Kohat area, Neocomian to Middle Albian in Kalachitta and southern Hazara and Tithonian to Middle Albian in northern Hazara (Fatmi 1977).

Kawagarh Formation (Day in Fatmi 1977, Fatmi 1973): This Formation rests disconformably on the Lumshiwal Formation. In the type locality (Kawagarh Hills) and adjoining areas, the Formation consists mainly of dark marl, calcareous shale and nodular, argillaceous limestone. In eastern Kohat the lower part contains dolomitic limestone. Farther westward it consists of grey lithographic limestone **(Tsukail Tsuk Member)** and thin- to medium-bedded limestone with marl and shale **(Chalor Silli Member)**. Its thickness ranges from 40 m to 200 m, but the Formation is missing in the Salt Range and Trans-Indus Range. Its upper contact with the Paleocene Hangu Formation is disconformable. It contains *Glabotruncana lapparenti, G.fornicata, G.concavata carinata, G.elevata calcarata, Heterohelix reussi, H. globocarinate, H.globulosa, Pseudotextularia elegans, Rugoglobigerina rugosa* and *Globorotalites multisepta* (Latif 1970a, 1970c). The age of the Kawagarh Formation is Late Cretaceous (Coniacian to Campanian).

Kohistan Magmatic Arc

The Kohistan magmatic arc terrain contains a complex metasedimentary and metavolcanic sequence of Mesozoic age (Fig. 4.39) which has been described summarily by Bard et al. (1980) as follows:-

- an upper detrital series (Yasin Group of Ivanac et al. 1956) - Lower Cretaceous.
- a volcanic calc-alkaline series (Utror Volcanics of Jan et al. 1971) and

- a metasedimentary oceanic series (Kalam Group of Matsushita et al. 1965).

This sequence is exposed south of the Main Karakoram Thrust (MKT) which denotes the suture between the Kohistan arc terrain (see Chapter 6) and Eurasian Plate in the region extending from Dir and Chitral, through Gilgit, to Skardu and Ladakh. Many workers have studied and mapped this sequence in different segments of the arc, and several separate names have been proposed for the sequence. An extremely rugged terrain, a complicated schuppen structure and a varying degree of metamorphism has further complicated the definition and correlation of the tectonostratigraphy of the region. However, a summary of the Cretaceous stratigraphy of this area (from Dir and Chitral in the west to Skardu in the east) is given below.

Kalam Group (Matsushita et al. 1965, Tahirkheli 1979): This Group crops out at several places between Dir and Drosh and extends northeastward. This Group is about 800 m thick, includes the **Karandoki Slate** at the base, overlain by **Deshan Banda Limestone**, and **Shon Quartzite**. The Deshan Banda Limestone clasts in the overlying Utror volcanics contain *Rhabdophyllia sp.* of coral and the Kalam Group has been assigned a Middle Jurassic to Cretaceous age (Tahirkheli 1980).

Dir Group (Tahirkheli 1979): The Kalam Group is overlain by the Dir Group of rocks consisting of the **Baraul Banda Slate** with fossiliferous limestone interbeds (containing *Actinocyclina, Discocyclina and Nummulities atacicus*) which is in turn overlain by the **Utror Volcanics**. The Dir Group is faulted over the Kalam Group (Tahirkheli 1979). Bard et al. (1980) have described this sequence south of Drosh and south of MMT as Kalam-type metagreywacke series with shales, cherts, marbles and some oceanic gabbros and pillowed metabasalts interlayered with Late-kinematic heterogenous quartz-diorite and overlain by calcalkaline lava flows (mainly andesites and dacites = **Utror Series**). They state that these rocks underlie a thrust plate of Drosh–Yasin red and green series with basal conglomerates, shales and fossiliferous Aptian-Albian limestones containing Lower Cretaceous fossils which they refer to as the detrital upper series. They have shown this sequence to continue eastward to Gilgit and Ladakh in the form of an arcuate belt.

Yasin Group: A part of the above sequence, between the Shandur Pass and Yasin, has been referred to as the Yasin Group (Ivanac et al. 1956), which consists of lavas, tuffs and agglomerates containing lenses and beds of fossiliferous, massive and shaly limestone of Cretaceous age. This limestone yielded a Barremian to Aptian fauna including *Eugyra cf. E.neocomiensis* DeFrom, *Calamophylia cf. C.gracilis* Blainy, *Thecosmilia* sp., *Isastrea cf. I.regularis* DeFrom, *Montastrea sp., Horiopleura haydeni* Doub, *Horiopleura* cf. *H.haydeni* Doub, *Horiopleura cf. H.Lamberti* (Mun-Chal), *Nerinea cf. N.Coquandi* D' Orb. and *Ptygmatis sp.*, (Brunnschweiler in Ivanac et al. 1956). The volcanic rocks underlying and locally interbedded with Yasin Group have been variously named **Greenstone Complex**, **Turmik Formation**, **Rakaposhi Volcanic Group**, and **Chalt Volcanic Group** (see Chapter 6). Pudsey et al. (1985) have named the Yasin Group rocks as the **Shamran Volcanic Group**. According to them it consists of porphyritic andesites and tuff intercalated with shale. Locally there are pillow lavas chert bands, volcanic breccia and agglomerates. Shamran Volcanics are overlain by 500 m of volcanic-lithic sandstone, shale and micritic limestone with *orbitolinas* and *rudistids*.

Between Drosh and Shamran, south of MKT, Pudsey et al. (1985) mapped and described the same sequence (Kalam-Utror-Yasin Groups) which they divided into three units as follows:-

Gawuch Formation: The type section of this Formation is in Gawuch Gol where it is comprised mainly of green phyllites and limestone, having a maximum thickness of about 2 km. Towards the base the phyllites are intruded by diorite. It is overlain by the Purit Formation.

Purit Formation: The type section is in Drosh Gol, SE of Kawash. This Formation consists of red shales with subordinate sandstones and conglomerates. The shale is calcareous and with abundant calcite veins; the sandstone is lithic to feldspathic; and the conglomerate is polymict with clasts of andesite, limestone and volcanic rocks. This Formation is about 1 km thick and is overlain by the Drosh Formation.

Drosh Formation: Its type section is in Drosh Gol, where it consists of thickly bedded, epidotised, vesicular andesite interbedded with thin red shales. It is about 1.5 km thick and it is faulted against the northern suture melange (Shyok Suture).

The geological map of Pudsey et al. (1985) shows a fairly wide development and extent of the Drosh Formation and Shamran Volcanic Group, covering areas which had been earlier mapped and described as Greenstone Complex by Ivanac et al. 1956.

Burji Formation: The upper metasedimentary, volcanic and detrital sequence of the Kohistan magmatic arc (Bard et al. 1980), lying immediately south of MKT, may be traced eastward into Skardu–Ladakh area. Here a sequence of phyllites, slates, limestone and chlorite-epidote green-schists, several kilometres thick, is exposed south of the Indus River and has been named Burji Formation (Desio 1978). The limestone contains Upper Cretaceous fossils.

Deosai Volcanics: The Burji Formation is interbedded with Deosai Volcanics (Wadia 1937, Tahirkheli 1982), consisting mainly of andesite, rhyodacite and rhyolite. Northward and upward in the sequence, the Burji Formation has faulted contact with the Katzarah Formation.

Katzarah Formation (Desio 1963): This Formation consists of high-grade metasediments (amphibolite facies or higher) and crops out extensively in the Indus Valley and the lower Shigar Valley in the Skardu region. The Katzarah Formation is about 1,500 m thick and includes hornfels, biotite schist, sillimanite-schist, gneisses and amphibolites. Desio (1964) has subdivided this Formation into **Tsordas Gneiss**, **Skoyo Gneiss**, **Askore Amphibolite** and **Katzarah Schist**. The Formation is intruded by pegmatite veins and dykes and sills of basic to acidic igneous rocks. The Katzarah Formation has been intruded by the **Ladakh Intrusives** (Frank et al. 1977, Ivanac et al. 19556, Tahirkheli 1982).

Bauma–Harel Formation: The Katzarah Formation is overlain by the Bauma–Harel Formation (Desio 1963), which consists largely of volcaniclastic metasediments that have been metamorphosed to cholorite-epidote greenschists. The greenschists are interbedded with slate, phyllite, minor carbonate layers and conglomerate that has clasts of shale, marble and greenschist (Hanson 1989). Northward, the outcrops are truncated by the MKT and covered by the thrust block of Daltumbore Formation.

Jaglot Schist Group (Treloar et al. 1996): A thick sequence of metasediments and interbedded metavolcanics underlies the volcanics in the Sai Nala and near Jaglot, south of Gilgit. T. Khan et al. (1994, 1995, 1996) have described this sequence as **Jaglot Group.** The lower par of the sequence comprises schists and paragneisses of the Gilgit Formation

(Jurassic?) which are overlain by the Ganshu-Confluence Volcanics (Early Cretaceous). Slates, marbles and volcanics of the Thelichi Formation form the upper part of the sequence (see Chapter 6 also).

Earlier Ivanac et al. (1956) mapped the metavolcanic sequence of the Kohistan magmatic arc Complex south of MKT between Chitral and Gilgit and named it **Greenstone Complex**. They described it as an assemblage of lava, tuff, agglomerate, meta-gneiss, quartzite, limestone and calc-silicate rocks and following Wadia (1935) assigned these rocks to the Triassic. As may be seen from the description of these volcanic rocks by other authors (Matsushita et al. 1965, Tahirkheli 1979, 1982, Bard et al. 1980, Pudsey et al. 1985), this rock sequence is now believed to be largely of Cretaceous age.

Following Bard et al. (1980), Tahirkheli (1982) has proposed simplification of stratigraphic nomenclature for the Late Mesozoic metavolcano-sedimentary sequence in the Kohistan magmatic arc and has used the name **Rakaposhi Volcanic Complex** for the Greenstone Complex of Ivanac et al. (1956). He apparently includes in it the Shamran Volcanic Group of Pudsey et al. (1985). According to Tahirkheli (1982), *Globotrancana* has been found in the Rakaposhi Volcanic Group in the Tissar section upstream of Shigar and *Thaminasteria matshushitai* near Shamran in Ghizar Valley, which indicate a Cretaceous age for this Group. Tahirkheli (1982) also has extended Yasin Group from Skardu through Chalt, Ishkuman, Yasin to Chitral which implies that the Bauma–Harel Formation of Desio (1963), the upper part of the Greenstone Complex of Ivanac et al. (1956), and the Purit and Drosh Formations of Pudsey et al. (1985) may be included in the Yasin Group.

Karakoram

Krinj Limestone: In the Chitral area, thick platform-type Cretaceous carbonate sequence, exposed in the Lutkho Valley east of Shogor, has been named Krinj Limestone (Desio 1959, Tahirkheli 1982). The limestone overlies the Chitral Slate and interfingers tectonically with the Reshun Formation (Pascoe 1959). The Krinj Limestone, which is about 2.5 km thick, is a sequence of grey, cream to white, micritic to finely recrystallised, massive to well bedded limestone (Pudsey et al. 1985). The limestone contains *Orbitolina discordia* and *Hippurites* (Desio 1959).

Reshun Formation: Near Reshun, in the Mastuj Valley, a sequence of thick conglomerate overlain by micritic limestone and maroon and red shales known as Reshun Formation (Hayden 1915) forms an extensive outcrop. It is 140 m to 1,500 m thick. Its upper and lower contacts are faulted; on one side it rests against the Chitral Slate and on the other against Devonian carbonate rocks (Tahirkheli 1982). However SW of Reshun, the Formation overlies the Chitral Slates with a sharp angular unconformity (Pudsey et al. 1985). At Reshun the conglomerate bed is 350 m to 400 m thick and has clasts of limestone (44%), volcanic rocks (32%) and quartzite (16%). Because of its stratigraphic position and inclusion of *Orbitolina* and *rudist*-bearing limestone clasts, the Reshun Formation has been given a Late Cretaceous age (Hayden 1915, Calkins 1981, Tahirkheli 1982, Pudsey et al. 1985). Desio (1959, 1966, 1975), however, has considered it as a Tertiary sequence. Tahirkheli (1982) has divided the Formation into two units. He has named the lower psammites as the **Awi Conglomerate** and the overlying micritic limestone and maroon shales as **Reshun Shales**.

Tupop Conglomerate: A thick lithic conglomerate overlies Jurassic–Early Cretaceous sequence with a spectacular angular unconformity, in the Kundil and Borom tributaries of

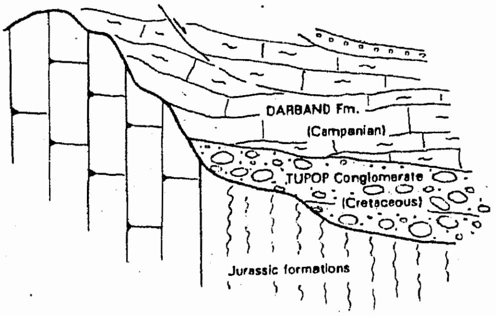


Fig. 5.7. Section showing the spectacular onlap of the Cretaceous sequence over the Jurassic in Upper Kundil Valley, Upper Hunza region (from Gaetani et al. 1993).

the lower Chapursan Valley in upper Hunza (Fig. 5.7). It has been named Tupop Conglomerate by Gaetani et al. (1993). It contains clasts of limestone, sandstone, siltstone, shale, quartzite, and felsic volcanic rock derived from the underlying Permian to Jurassic units. The Tupop Conglomerate lacks age-diagnostic fossils, though it contains abundant reworked echinoderm fragments, Permian fusulinids, and limestone clasts with algae, bivalves or gastropods. A Cretaceous age for the Tupop Conglomerate is inferred as it is unconformably bracketed between a Jurassic to Early Cretaceous and a Late Cretaceous sequence (Fig. 5.7).

Darband Formation (Gaetani et al. 1993): The Tupop Conglomerate is overlain with an angular unconformity by the 300 m thick Darband Formation which largely consists of red and grey nodular limestones, interbedded with conglomerates at the top (Fig. 5.7). Inderminate fragments of *inoceramids* and possible *globotruncanids* have been reported. The Formation however, has been assigned a Late Cretaceous age (Campanian to Early Maestrichtian) on the basis of coccoliths-*Ceratolithoides verbekii* Perch-Nielsen and *calculites obscurus* (Gaetani et al. 1993).

Chagai Magmatic Arc

Sinjrani Volcanic Group: Cretaceous volcano-sedimentary rocks are extensively exposed also in the Chagai Magmatic Arc (see Chapter 6). The oldest rock sequence exposed in this region is the Sinjrani Volcanic Group (HSC 1960). This Group consists mainly of agglomerate, volcanic conglomerate, tuff and lava with subordinate shale, sandstone and limestone. These rocks are 900 m to 1,200 m thick. Poorly preserved foraminifera, corals *(Mesophyllum)* and algae have been found and the Group has been assigned to the Cretaceous. Sinjrani-type rocks occur in the Ras Koh Range also where they have been named **Kuchakki Volcanic Group** (HSC 160). See Chapter 6.

Humai Formation (HSC 1960): This Formation unconformably overlies the Sinjrani Volcanic Group south of Chagai Hills; elsewhere this contact is disconformable as indicated by a basal conglomerate. The lithology of this Formation is very variable and it consists of grey to purple shale, sandstone, siltstone, thin-bedded limestone, volcanic conglomerate and massive reefoid limestone with *Hippurites sp., H. loftusi, Lapeirousia sp., Monopleuridea, Lepidorbitoides socialis, Orbitella media, Orbitoides sp., Omphalocyclus sp.,* and *Balculogypsinoides sp.,* (Fatmi 1977). The Humai Formation is Maestrichtian in age.

CENOZOIC

The Cenozoic sedimentary sequence contains an exceptionally good record of the geodynamic processes which were responsible for the extinction and demise of the Tethys Ocean during this Era. This sequence records the effects of the Indo-Pakistan plate collision with Eurasian Plate, namely the Himalayan Orogeny, emplacement of ophiolite and melange sequences, widespread volcanism and magmatism, and change from Paleogene marine sequence to littoral and terrestrial molasse-type deposition during Neogene. It also records the evolution of vertebrate faunas. The present day configuration of the country thus evolved during this Era.

The Cenozoic sedimentary sequence consists of a largely marine Palaeogene sequence covered by dominantly terrestrial Neogene series of rock formations in the Indus Basin, and an entirely marine flysch-type sequence in the Balochistan region. At most localities the Cenozoic sequence has an accumulated thickness of over 1,500 m (Fig. 5.5). A widespread hiatus marks the close of the Mesozoic Era and consequently, the Cenozoic-Mesozoic boundary is variable from place to place. In the Sulaiman fold belt and in the Balochistan ophiolite-and-thrust belt, the Cenozoic-Mesozoic contact changes from an angular unconformity between Paleocene and older units (down to the Jurassic) to a disconformity and overlap in the Kohat-Potwar-Salt Range region (Raza et al. 1989; Gee 1989). In the Lower Indus Basin, the contact between the Paleocene and the Cretaceous is disconformable (Williams 1959) though according to the HSC (1960) it is transitional at places.

The various lithostratigraphic units that comprise the Cenozoic sedimentary sequence are shown on Table 5.7.

Paleogene

Chagai Magmatic Arc

Rakhshani Formation: The Paleogene sequence in Chagai rests conformably on the Humai Formation of Cretaceous age; in the Ras Koh Range it conformably rests on the Cretaceous Kuchakki Volcanic Group. The lower part of the sequence is called the Rakhshani Formation (HSC 1960; Cheema 1977) which consists of green to grey, medium to coarse-grained sandstone with lava flows, tuffs and agglomerates. Subordinate grey to black argillaceous limestone beds in the lower part of the Formation contain foraminifera, bivalves, gastropods, corals and algae. The foraminifera include *Alveolina vredenburgi, Flosculina globosa, Miscellanea miscella,* and *Saudia labyrinthica.* The Rakhshani Formation is Paleocene in age.

Kharan Formation: In the Kharan area, Rakhshani Formation is conformably overlain by the Kharan Formation (HSC 1960, Cheema 1977; type section near Jalwar) which consists mainly of grey, thin- to thick-bedded, argillaceous, fossiliferous and reefoid limestone. It contains intercalations of thin, grey to brown calcareous shale and fine- to medium-grained calcareous sandstone. Its thickness ranges from 90 to 600 m. Some of the larger foraminifera found in the Formation are *Assilina granulosa, A. exponens, Flosculina globosa, Fasciolites oblonga, F. subpyrenaica, Discocyclina ranikotensis,* and *Orbitolites complanatus*. The age of the Formation is Early to Middle Eocene. The Kharan Formation is overlain conformably and transitionally by the Khojak Formation of Late Eocene to Oligocene age.

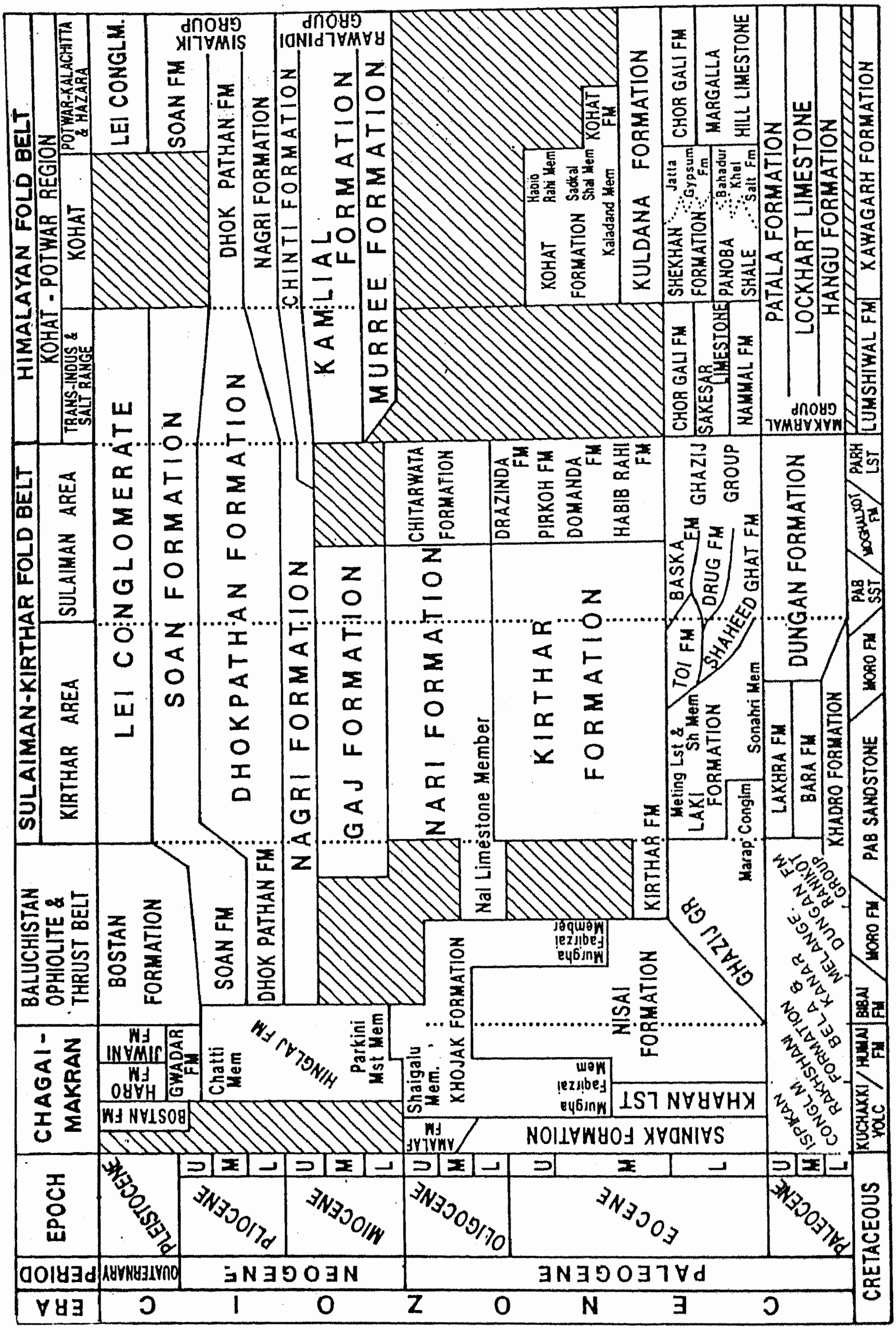


Table 5.7. Correlation of the Cenozoic sequences.

Saindak Formation: In the Chagai–Ras Koh area the Saindak Formation (HSC 1960, Cheema 1977) rests conformably on the Rakhshani Formation. It consists of interbedded shale, sandstone, limestone and volcanic rocks. The shale is green to brown or maroon, sandy or calcareous with gypsum beds at places; the sandstone is green to brown, fine to coarse-grained, gritty and calcareous. The limestone is grey to brown and fossiliferous. The volcanic rocks include agglomerate and conglomerate containing fragments and boulders of lava flows. The thickness of the Formation ranges from about 60 m to 1,500 m. It contains foraminifera, gastropods, bivalves, corals and echinoderms. The fossils include *Assilina dandotica, A. exponens, Opertorbitolites douvillei, Fasciolites oblonga, Echinolampas nummulitica* and *Velates perversus*. This fauna suggests an Eocene age for the Saindak Formation.

Amalaf Formation (HSC 1960): This Formation transitionally over-lies the Saindak Formation. It consists mainly of volcanic ash and agglomerate with subordinate andesitic lava flows. The volcanic rocks are interlayered with shale and sandstone. This Formation is exposed in the western part of the Chagai District. No fossils have been yet reported from it, but the Formation has been tentatively assigned an Oligocene age (Cheema et al. 1977).

Kakar Khorasan–Makran Flysch Belt

Ispikan Conglomerate (HSC 1960, Cheema 1977): It crops out only in the vicinity of Ispikan, 20 km northeast of Mand. It rests unconformably on Cretaceous marl. It contains unbedded and unsorted boulders and fragments of limestone (probably from Jurassic strata) and igneous rocks (including fragments from Sinjrani Volcanics). Blocks of the Eocene Nisai Formation rest on Ispikan Conglomerate. This conglomerate has been tentatively assigned a Paleocene age.

Nisai Formation (HSC 1960, Cheema 1977): This Formation is exposed along the eastern margin of the flysch belt and unconformably overlies various Mesozoic and Permian rocks. However, farther eastward it rests conformably on the Ghazij Formation. It is mainly comprised of grey to greyish brown massive, brecciated, reefoid and shelly limestone. At places the limestone is grey to black and well bedded and oolitic. The Formation contains subordinate amounts of marl, shale and conglomerate. The amount of these subordinate constituents varies from place to place. The shale is grey, calcareous, in places lateritic or carbonaceous; the sandstone is grey, brown, green or white, coarse, poorly sorted, cross-bedded, thick-bedded, and protoquartzitic to orthoquartizitic. The conglomerate occurs at different horizons in different areas and comprises angular to rounded pebbles and boulders of limestone, sandstone, jasper and various igneous rocks. The thickness of the Formation ranges from 30 m to 1,200 m. Fossils found in the Formation include *Assilina granulosa, A. subpapillata, Fasciolites globosa, F. oblonga, Discocyclina sowerbyi, Lockhartia conditi, L. newboldi, Nummulites atacicus, N. mammilla, Orbitolites complanatus, Pellatispira madaraszi, Fasciolites elliptica, Lepidocyclina dilatata, Nummulites intermedius* and *N. fichteli*. The age of the Formation is Eocene to Early Oligocene.

Khojak Formation (Vredenburg 1909a, HSC 1960): Conformably overlying the Nisai

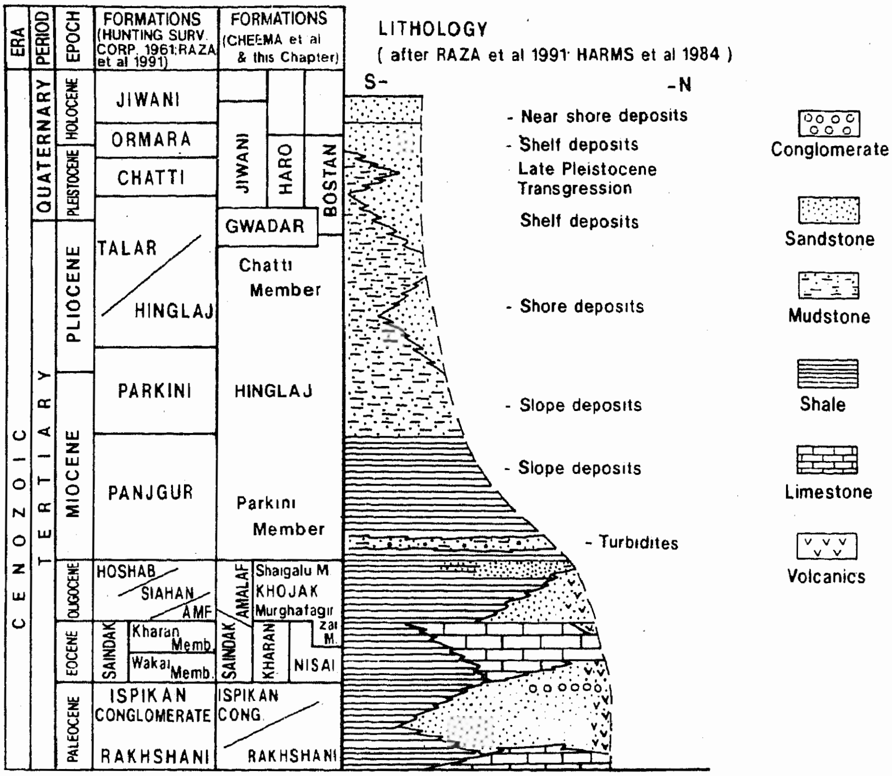


Fig 5.8. Stratigraphic column showing the Cenozoic sequence in the Makran region (From Harms et al. 1984)

Formation (Fig. 5.8), this **Formation** comprises a thick sequence of shale, sandstone and siltstone with subordinate limestone and conglomerate. Its lower part, the **Murgha Faqirzai Member**, consists mainly of grey, green to brown calcareous and arenaceous shales. It contains turbidites, prodelta-type laminated mudstones, and upward thickening sequences of hammocky sandstones. The upper part of the Formation, the **Shaigalu Member** (Cheema 1977), is largely grey, green to brown, cross-bedded, calcareous, micaceous sandstone with subordinate conglomerate. It is characterised by upward-fining sequences, multicoloured mudstones, paleosols and rare coal beds. Qayyum et al. (1996) interpret it as an upper deltaic-plain deposit. According to them the Khojak Formation is more than 6,300 m thick and was deposited as part of a major delta-deep sea fan system which formed in the Katawaz–Makran Sea, a remnant of the Neo-Tethys. It was fed by a ancestral drainage from the rising Himalayas in the east.

Paleocene to Oligocene foraminifera *(Assilina granulosa, Nummulites atacicus, N. globulus, N. cf irregularis, Discocyclina ranikoti; Lepidocyclina dilatata, Nummulites fischteli, N. intermedius* and *N. Vascus)* and Oligocene to Early Miocene molluscs *(Telescopium charpentieri, Tympanotonous margaritaceus, Ostrea gryphoides)* have been reported from the Khojak Formation (Eames 1939, HSC 1960). Its age ranges from Eocene to Early Miocene.

Balochistan Ophiolite-and-Thrust Belt and Kirthar Fold Belt

Ranikot Group: In the Kirthar area of the Indus Basin, the Paleocene succession consists of the Ranikot Group (Blandford 1876, 1879, Vredenburg 1909b). This Group has a basal marine sequence of sandstone and shale with interbeds of limestone and basaltic lava flows (Khadro Formation), a middle fluviatile to paralic sequence of sandstone and shale with coal and carbonaceous beds (Bara Formation), and an upper sequence consisting of marine limestone and some estuarine sandstone and shale (Lakhra Formation).

Khadro Formation (Williams 1959): It is widely distributed in the Kirthar and adjacent region, lies unconformably on the Late Cretaceous Pab Sandstone and the Moro Formation. The type locality is Bara Nai in the Lakhi Range. At the type locality, the basal part of the Formation consists of limestone containing oysters and reptile bones. This is followed by a series of olive, brown, green to grey, soft, ferruginous, medium-grained fossiliferous sandstone and olive, grey to brown gypsiferous shale with interbeds of fossiliferous limestone. A number of basaltic lave flows are also present. The thickness of the Formation is 67 m in the type section; in the Rakhi Nala, Sulaiman area, it is 170 m thick whereas in the test holes at Lakhra and Dabbo Creek, thicknesses of 140 m and 180 m were penetrated. Fossils described from the Khadro Formation include *Corbula (Vericorbula) harpa, Leionucula rakhiensis, Venericardia vredenburgi, Tibia (Tibiochilus) rakhiensis* (Eames 1952); *Cardita beaumonti* (Blanford 1878); *Globigerina pseudobulloides and G. triloculinoides* (Nagappa 1959); *Howecythereis multispinosa, H. micromma* and *Paracypris rectoventra* (Sohn 1959). The Formation has thus been considered to be Early Paleocene (Danian) in age.

Bara Formation (Cheema 1977): This Formation conformably overlies the Khadro Formation and is widely distributed in the Kirthar Range and adjacent areas. The type section is in Bara Nai in Lakhi Range. The Formation consists of interbedded sandstone and shale. The sandstone is fine- to coarse-grained, calcareous, ferruginous and at places glauconitic, ripple-marked and cross-bedded. Beds range in thickness from a few centimetres to over 3 m. The shale is soft, earthy, gypsiferous and commonly carbonaceous.

Oyster shells, some reptile remains and abundant leaf impressions have been found. *Ostraea talpur* has been reported from the base (Vredenburg 1928). The Formation has been given a Middle Paleocene age (Thanetian). (See Chapter 10 also).

Lakhra Formation (Cheema et al. 1977): It overlies the Bara Formation and crops out in the Kirthar and adjacent areas. The type section is in the southern part of the Lakhra anticline. It consists mostly of grey, thin- to thick-bedded, nodular, sandy and in places argillaceous, fossiliferous limestone, with interbeds of sandstone and shale in the upper part. Its thickness ranges from 50 m to about 242 m. The Formation contains a rich assemblage of foraminifera (Davies 1927, Nuttall 1931), corals (Duncan 1880), molluscs (Vredenburg 1909b, 1928) and echinoids (Duncan and Sladen 1882). The more significant larger foraminifera include *Miscellanea miscella, M. stampi, Lockhartia haimei, Lepidocyclina punjabensis, Discocyclina ranikotensis, Nummulites nutalli, N. thalicus, N. globulus, N. sindensis, Assilina ranikoti.* The Lakhra Formation is Late Paleocene in age.

Laki Formation (Noetling 1903, HSC 1960): It overlies the Ranikot Group unconformably and is exposed mainly in the southern Kirthar Range and in the southern Sulaiman

Range. The type locality is near Meting and near Mari Nai in the northern Lakhi Range. The Formation comprises cream coloured to grey limestone, with subordinate marl, calcareous shale, sandstone and lateritic clay. It contains a rich fossil assemblage of foraminifera, gastropods, bivalves, echinoids and algae. The fossils include: *Assilina granulosa, A. pustulosa, Lockhartia hunti, Flosculina globosa, Fasciolites oblonga, Gisortia murchisoni, Velates perversus;* and *Amblypygus subrotundus* and *Echinolampus nummulitica* (HSC 1960, Iqbal 1973, Nuttall 1925). The lower part of the Formation has been divided into the Sonari Member and Meting Limestone and Shale Member by HSC (1960).

The **Sonari Member** contains lateritic clay, shale, pockets of limonite and ochre, coal beds, sandstone and locally developed yellow, arenaceous limestone. The Sonari Member is overlain by the **Meting Limestone and Shale Member** (Vredenburg 1906) which consists of thin-bedded creamy white, fossiliferous, nodular limestone followed by a sequence of interbedded limestone and shale with subordinate sandstone. The age of the Laki Formation is Early Eocene (Ypresian).

Sonari Formation: The Sonari Member of HSC has been now excluded from the Laki and renamed Sonari Formation because it is distinct from the overlying and underlying Laki and Lakhra Limestone and it comprises a distinct nonmarine to brackish-water deposit (Outerbridge et al. 1989-91).

Kirthar Formation (Blanford 1876, Noetling 1903, Cheema et al. 1977): It overlies the Laki Formation conformably in the Kirthar area. It is distributed widely and covers also the Sulaiman area and parts of Waziristan. In these areas it overlies the Ghazij Group conformably. The type area is in the Kirthar Range (Gaj River section). The Formation is mainly fossiliferous limestone interbedded with subordinate shale and marl. The limestone is thick-bedded to massive, nodular in some areas, grey to white in colour and locally contains algal and corraline structures. In some localities the upper half of the Formation is massive cliff-forming limestone. The shale is olive, orange, yellow or grey, soft, earthy and calcareous. The thickness of the Formation ranges from 15 m to 30 m in western Kirthar Range to 1,270 m in the Gaj River type section (Cheema et al. 1977).

The Formation contains abundant foraminifera, gastropods, bivalves, echinoids and vertebrate remains (Oldham 1890, Vredenburg 1906, 1909a, Pilgrim 1940, Eames 1952, HSC 1960). The age of the Kirthar Formation is Middle Eocene to Early Oligocene.

Nari Formation (Blanford 1876, Williams 1959): It is exposed extensively in the Kirthar and Sulaiman region, whereas scattered outcrops are found in tectonised thrust blocks in the Balochistan ophiolite-and-thrust belt. In the Kirthar Province it conformably overlies the Kirthar Formation except in the Hyderabad anticlinorium where it oversteps and unconformably overlies the Kirthar and Laki Formations. The type section is in the Gaj River gorge in the Kirthar Range.

The upper part of Nari Formation is mostly brown, fine- to coarse-grained sandstone (locally conglomeratic) with interbeds of shale. The lower part consists of interbedded grey to brown, fossiliferous sandy limestone, calcareous sandstone and shale. At many localities the lower part of the Formation is a grey to brown shelly, nodular, thick-bedded to massive limestone which has been named the Nal Member (HSC 1960). The thickness of the Nari Formation ranges from 1,045 m to 1,820 m in the Kirthar area. It contains a rich fauna including echinoids, molluscs, corals, foraminifera and algae (Duncan et al. 1963, Khan 1968, Iqbal 1969b). Some of the significant larger foraminifera include *Nummulites* *intermedius, N. vascus, N. fichteli, N. clipens* and *Lepidocyclina dilatata*. The age of the Nari Formation is Oligocene to Early Miocene (Rupelian to Early Aquitanian; Latif 1964, Khan 1968).

Sulaiman Fold Belt

Dungan Formation: In the Sulaiman area the Dungan Formation (Oldham 1890, Williams 1959, Cheema et al. 1977, Kazmi 1988) comprises the Paleocene sequence and oversteps most of the Cretaceous rock units (Parh Limestone, Moghal Kot Formation, Pab Sandstone). It crops out extensively in the northern part of the Kirthar Range, and the whole of the Sulaiman Range. The type section is in the Mehrab Tangi 8 km northeast of Harnai. The Formation consists largely of thick- to medium-bedded to massive nodular limestone with subordinate marl, shale and sandstone. Its thickness varies from about 100 m to over 600 m. It contains a rich fossil assemblage of foraminifera, gastropods, bivalves and algae (Davies 1941, HSC 1960, Latif 1964, Iqbal 1969a, Kazmi 1988). Some of the more common and significant larger foraminifera found in the Formation are *Miscellanea miscella, M. stampi, Lockhartia tipperi, Kathina selveri, Operculina, Discocyclina, Nummulites nuttalli, N. thalicus, N. sindensis, Assilina dandotica* and *Fasciolites*. The age of the Dungan Formation is Paleocene to Early Eocene.

Ghazij Group (Oldham 1890, Williams 1959, Shah 1990): It conformably overlies the Dungan Formation and crops out extensively in the Sulaiman Range and the northern part of the Kirthar Range. The type section is at Spintangi. The Group dominantly consists of shale, interbedded with layers and lenses of claystone, mudstone, sandstone, limestone, conglomerate and alabaster (Kazmi 1962). It contains deposits of coal which are being mined. Its thickness ranges from about 160 m at Roara Nai (Kirthar Range), 590 m at the type section (Spintangi) to about 1,300 m at Moghal Kot. The Ghazij Group contains foraminifera, gastropods, bivalves, echinoids, algae and plant remains (Eames 1952, HSC 1960, Latif 1964, Iqbal 1969a). The foraminifera include *Flosculina globosa, Assilina sublaminosa, Lockhartia hunti, Coskinolina balsilliai* and *Dictyoconus indicus*. The Ghazij Group is Early Eocene in age, and has been further subdivided into following lithostratigraphic units.

Marap Conglomerate (HSC 1960): It crops out in the Kalat Plateau area and forms the basal part of the Ghazij Group. It consists of well rounded and poorly sorted pebbles and boulders of limestone, shale, and sandstone derived from the underlying formations including the Jurassic rocks. The conglomerate is interbedded with sub-ordinate shale, sandstone and, less-commonly limestone. The Marap Valley is the type locality where it is about 910 m thick.

Shaheed Ghat Formation (Shah 1987, 1990): This Formation is exposed in the Moghal Kot area. It overlies the Dungan Formation and forms the lower part of the Ghazij Group. The type section is at Shaheed Ghat 5 km southwest of Zinda Pir. The Formation consists of grey to olive green, laminated shale with marl. At places the shale contains layers with nummulites, gastropods and lamellibranchs. The thickness of the Formation ranges from about 340 m at Moghal Kot to over 680 m at Shaheed Ghat. The Shaheed Ghat Formation is Early Eocene in age.

Drug Formation (Shah 1987, 1990): It overlies the Shaheed Ghat Formation conformably and crops out in the Sulaiman Range and adjacent areas. The type section is at Drug

Tangi 3 km northeast of Drug Village. The Formation consists largely of limestone interbedded with subordinate shale. The limestone is orange, pale-olive, grey-green to creamish-white in colour. It is thin-bedded, hard, pebbly and nodular, crystalline, argillaceous and commonly has marly partings. At places it is bioclastic and of calcirudite type. The lower part of the Formation is mostly shale which is greenish-grey to dark grey, calcareous, at places pyritic or with ferruginous concretions and interbedded with minor limestone. The thickness of the Formation ranges from 40 m at Drug Tangi to 340 m in Zinda Pir area. The Drug Formation is Early Eocene in age.

Toi Formation (Shah 1987, 1990): It overlies the Drug Formation conformably, and, where the Drug Formation is missing, it overlies the Shaheed Ghat Formation. It crops out extensively in the Sulaiman Range; the type section is in Toi Nala. It consists mainly of interbedded sandstone and mudstone, siltstone, shale and conglomerate with locally-developed coal seams. The sandstone is coarse-grained, pebbly, poorly-sorted and cross-bedded, at places containing freshwater bivalves, gastropods and plant remains. The mudstone and siltstone are brown to reddish-brown, soft, blackish and calcareous. Locally the Formation contains thin beds of fossiliferous limestone interbedded with mudstone and shale. The maximum thickness of the Formation is about 1,196 m in the Moghal Kot section. The Toi Formation is Early Eocene in age.

Baska Formation (Hemphil et al. 1973): It overlies conformably the Toi Formation and forms the upper part of the Ghazij Group in the Sulaiman Range. The type locality is 2 km east-northeast of Baska Village. The Formation consists of green to grey shale and claystone interbedded with alabaster, gypsiferous limestone and marl. The alabaster beds range in thickness from a few centimetres to 10 metre. At places the shale is interbedded with thin limestone beds. The thickness of the Formation ranges from about 160 m in the north to about 820 m in the southern part of the Sulaiman Range. It contains foraminifera, bivalves and gastropods. The larger foraminifera include *Lockhartia hunti* and *Dictyoconoides vredenburgi*. The Baska Formation is Early Eocene in age.

Kirthar Formation: This Formation conformably overlies the Ghazij Group in most of the Sulaiman Range; it has been described earlier. However, in the Sulaiman Range and Kohat area, the Ghazij Group and its equivalent formations are conformably overlain in ascending order by the following formations (Table 5.7) which were previously considered as subunits of the Kirthar Formation (Cheema 1977, Shah 1987).

Habib Rahi Formation (Tainsh 1959, Meissner et al. 1968): It overlies the Baska Formation conformably and transitionally and the type section is proposed north of Zampost, along the Zhob–Dera Ismail Khan road. The Formation is largely greyish-brown, thin- to thick-bedded or massive, hard argillaceous, fossiliferous limestone, with nodules and cherty beds. Its thickness ranges from 15 m to 150 m. The Habib Rahi Formation is Middle Eocene in age.

Domanda Formation (Hemphil et al. 1973): It overlies the Habib Rahi Formation conformably and transitionally, and its type locality is west of Domanda. It consists mainly of dark brown to greenish grey claystone with intercalation of limestone at some places, and with subordinate grey to brown, fine to medium-grained thick-bedded to massive calcareous sandstone in the upper part. Its thickness ranges from about 130 m to 330 m. It contains foraminifera, gastropods, bivalves, echinoids and rare vertebrate fossils. The foraminifera include *Nummulites beaumonti, N. subbeaumonti, Alveolina elliptica* and *Dictyoconoides cooki*. The age of the Domanda Formation is Middle Eocene.

Pirkoh Formation (Hemphil et al. 1973): This unit conformably overlies the Domanda Formation and the Pirkoh anticline is its type locality. It consists of brown, grey to white, thin-bedded limestone with subordinate beds of soft argillaceous limestone and dark-grey calcareous claystone. Its thickness ranges from 10 m to 175 m. It is highly fossiliferous and contains foraminifera, gastropods, bivalves and echinoids of Middle Eocene age.

Drazinda Formation (Hemphil et al. 1973): It overlies the Pirkoh Formation conformably and transitionally; the type locality is northeast of Drazinda. It consists of brown to grey clay with subordinate fossiliferous marls and brown fossiliferous limestone interbeds. At places it is interbedded with grey, calcareous sandstone in the middle part. In the northern Sulaiman Range it contains celestite nodules. Its thickness ranges from 15 m to 500 m. It contains a rich fauna of foraminifera, bivalves, bryozoans and echinoids. Its age is Middle Eocene.

The Kirthar Formation and the Drazinda Formation of Eocene age are unconformably overlain by the **Nari Formation** of Oligocene age in the Sulaiman Range, which has been described earlier. In this region the Nari Formation is as much as 600 m thick.

Chitarwata Formation: In the Sulaiman Range the Oligocene sequence which was previously referred to as Nari Formation has now been renamed as the Chitarwata Formation (Hemphil and Kidwai 1973, Eames 1952, Shah 1987). The type section is at Chitarwata Pass. The Formation consists of red, grey and green claystone, siltstone. The lower part is largely sandstone which is ripple-marked and cross-bedded at places. At some localities hard ferruginous sandstone and conglomerate are present near the top of the Formation. The thickness of the Formation ranges from 150 m to 300 m. It contains freshwater gastropods and bivalves (Iqbal 1969b) and the plant specie *Croftiella escheri* on the basis of which the age of the Formation has been determined as Oligocene.

Himalayan Fold-and-Thrust Belt

Makarwal Group: In the Kohat–Potwar area the lower part of the Paleogene sequence comprising the Paleocene Makarwal Group of Gee (1989), has been deposited over the eroded surface of rocks ranging in age from Cretaceous to the Cambrian (Figs. 4.24 and 4.26). It extends rather uniformly over this region and includes a basal sandstone unit of terrestrial origin (Hangu Formation), a middle marine limestone unit (Lockhart Limestone), and an upper shale unit with subordinate sandstone (Patala Formation) which was deposited in the off shore region and in a back-barrier basin (Table 5.7).

Hangu Formation (Davies 1930, Fatmi 1973): It unconformably overlies various formations of Paleozoic to Mesozoic age. The type locality is south of Fort Lockhart in the Samana Range. It consists largely of grey to brown, fine- to coarse-grained, silty and ferruginous sandstone which grades upward into fossiliferous shale and calcareous sandstone. At places, the Formation is intercalated with grey argillaceous limestone and carbonaceous shale. In the Makarwal and Hangu areas it contains coal beds in the lower part. Its thickness ranges from about 15 m in Hazara to 150 m at Kohat Pass. The Formation contains molluscs, corals and foraminifera (Gregory 1930, Cox 1930b, Davies et al. 1937, Iqbal 1972, Smout and Haque 1956). Some of the important foraminifera include

Miscellania miscella, Lockhartia haimei, L. conditi, Lepidocyclina punjabensis, Operculina cf. canalifera and *Daviesina longhami*. The Hangu Formation is Early Paleocene in age.

Lockhart Limestone (Davies 1930, Fatmi 1973): This unit conformably overlies the Hangu Formation. Its type section is exposed near Fort Lockhart. It consists of grey, medium- to thick-bedded and massive limestone, which is rubbly and brecciated at places. In Hazara and Kalachitta it contains subordinate intercalations of grey marl and shale. Its thickness ranges from about 30 m to 240 m. It contains foraminifera, corals, molluscs, echinoids and algae (Cox 1931, Davies and Pinfold 1937, Davies 1943, Eames 1952, Haque 1956, Iqbal 1969a, Latif 1970c). Some of the important larger foraminifera comprise *Operculina subsalsa, O. juvani, Miscellanea miscella, M. stampi, Lockhartia haimei, L. newboldi, Lepidocyclina punjabensis, Discocyclina ranikotensis, Alveolina globosa, A. vredenburgi, Assilina dandotica, Nummulites nuttali, N. thalicus, N. sindensis* and *N. globulus*. The echinoderms include *Conoclypeus, Eurhodia morrisi, Hemiaster elongatus, Pleisolampas ovalis* and *P. placenta*. The age of the Lockhart Limestone is Paleocene.

Patala Formation (Davies and Pinfold 1937): It overlies the Lockhart Limestone conformably and its type section is in the Patala Nala in the western Salt Range. It consists largely of shale with subordinate marl, limestone and sandstone. Marcasite nodules are found in the shale. The sandstone is in the upper part. The Formation also contains coal (Warwick et al. 1988, 1990) and its thickness ranges from 27 m to over 200 m. It contains abundant foraminifera, molluscs and ostracods (Davies and Pinfold 1937, Eames 1952, Smout and Haque 1956, Latif 1970c). The larger foraminifera include *Lockhartia conditi, Nummulites globosa, Operculina canalifera, O. patalensis, Discocyclina ranikotensis, Assilina dandotica, Daviesina intermedia* and *Kathina nammalensis*. The age of the Formation is Late Paleocene.

The Patala Formation is overlain conformably by different sequences of Eocene age in different parts of its distribution area. In the Trans-Indus Ranges and the Salt Range the Eocene succession consists of the Nammal Formation, the Sakesar Limestone and the Chor Gali Formation.

Nammal Formation (Gee 1935, Fatmi 1973): The type section of this Formation in Nammal Gorge consists of shale, marl and argillaceous limestone. The lower part tends to be more shaly and marly, the upper part is dominantly limestone. Its thickness ranges from 35 m to over 130 m. The Formation contains molluscs and foraminifera (Cox 1931, Haque 1956, Khan in Fatmi 1973). Some of the important foraminifera are: *Nummulites atacicus, N. subatacicus, N. mamilla, N. lahirii, N. irregularis, Assilina granolosa, A. leymeriei, A. laminosa, Rotalia trochidiformis, Discocyclina ranikotensis* and *Fasciolites oblonga*. The age of the Nammal Formation is Late Paleocene to Eocene.

Sakesar Limestone: With increase in limestone beds, the Nammal Formation transitionally passes into the overlying Sakesar Limestone (Gee 1935, Fatmi 1973), the type locality of which is the Sakesar Peak. It consists of grey, nodular to massive limestone which is cherty in the upper part. Near Daudkhel, the Sakesar Limestone laterally grades into massive gypsum. Its thickness ranges from 70 m to about 300 m. It contains, amongst others *Nummulites atacicus, N. mamilla, Assilina granulosa, A. leymeriei, A. subspinosa, Fasciolites oblonga, F. globosa, Lockhartia conditi, L. hunti, Operculina nummulitoides, Orbitiolites complanatus* and *Rotalia trochidiformis*. Its age is Early Eocene.

Chor Gali Formation (Pascoe 1920, Fatmi 1973): This Formation rests conformably on the Sakesar Limestone (type locality Chor Gali Pass). It consists largely, in the lower part, of thin-bedded grey, partly dolomitised and argillaceous limestone with bituminous odour, and, in the upper part, of greenish, soft calcareous shale with interbeds. Its thickness ranges from 30 m to 140 m. It contains molluscs, ostracods and foraminifera (Davies et al. 1937, Eames 1952, Gill 1953, Latif 1970c). The more significant larger foraminifera include *Assilina spinosa, A. granulosa, A. daviesi, Flosculina globosa, globorotalia ressi, Lockhartia hunti, L. tipperi, L.conditi, Nummulites atacicus, N. mamilla, Orbitolites complanatus, Rotalia crookshanki, R. trochidiformis.* The age of the Chor Gali Formation is Early Eocene. It is overlain unconformably by the Neogene sequence.

Panoba Shale: In the Kohat area, the Eocene sequence conformably rests on the Paleocene Patala Formation and includes the Panoba Shale (with Bahadur Khel Salt), the Shekhan Formation, the Kuldana Formation and the Kohat Formation. The Panoba Shale (Eames 1952, Fatmi 1973, type locality south of Panoba village) consists of grey to olive, silty and calcareous shale with burrow markings. At places it is ferruginous, gypsiferous, and contains subordinate alum shale or thin flaggy limestone. Southwards it grades into **Bahadur Khel Salt Formation** (Gee 1945, Meissner et al. 1968). The Panoba Shale is 40 m thick and contains *Globorotalia aequa, G. quadrata, Assilina postulosa, Orbitolites complanatus* and *Nummulites sp.* The age of the Panoba Shale is Early Eocene.

Shekhan Formation (Davies 1924, Fatmi 1973): It rests conformably on the Panoba Shale and consists of yellow to grey, thin-bedded to massive, nodular limestone in the lower part, and gypsiferous shale in the upper part. In the vicinity of Mami Khel the upper part becomes gypsiferous and farther south in the Jatta area the Formation grades into the Jatta Gypsum Formation (Meissner et al. 1974). Its thickness ranges from about 50 m to over 70 m. It contains foraminifera, molluscs, echinoids and corals (Eames 1952, Nagappa 1959, Pascoe 1963). The larger foraminifera include *Assilina daviesi, A. laminosa, Nummulites atacicus, N. pinfoldi, Orbitolites complanatus, Discocyclina sp., Operculina sp.* and *Fasciolites oblonga* etc. The echinoids include *Conoclypeus pilgrimi* and *Hemiaster digonus*. The age of the Formation is Early Eocene.

Kuldana Formation (Wynne 1874, Fatmi 1973): It conformably overlies the Shekhan Formation and the Jatta Gypsum Formation. It consists dominantly of vari-coloured gypsiferous and arenaceous shale and brown to grey, gypsiferous marl, interbedded with red sandstone and calcareous conglomerate. It is about 17 m to over 160 m thick. It contains foraminifera, oysters and fresh water gastropods. Vertebrate remains are found in the middle part (Pinfold 1918, Meissner et al. 1968). *Assilina granulosa, A. subspinosa, A. exponens* have been reported from Hazara (Latif 1970c), and *Gaudryina, Lockhartia (L. hunti)* has been reported from Kohat (Meissner et al. 1968). The Formation has been assigned an early Middle Eocene to Late Eocene age.

Kohat Formation (Davies 1926a, Meissner et al. 1968, Fatmi 1973): This Formation rests on the Kuldana Formation conformably. It consists mainly of grey to cream-coloured, massive, nodular limestone and olive shale. It includes a lower **Kaladhand Limestone Member,** a middle **Sadkal Shale Member** and an Upper **Habib Rahi Limestone Member.** The Kohat Formation ranges in thickness from 50 m to 170 m and contains *Assilina postulosa, A. granulosa, Fasciolites oblonga, Dictyoconoides vredenburgi, Flosculina* *globosa, Lockhartia hunti, Nummulites mammilla* and *Orbitolites complanatus.* It is Eocene in age. The Kohat Formation is overlain unconformably by the Neogene sequence which has a thin conglomerate at the base.

Margalla Hill Limestone: In the Potwar, Kalachitta and Hazara areas, the Eocene sequence overlies the Paleocene Patala Formation conformably and includes the Margalla Hill Limestone, the Chor Gali Formation, and the Kuldana Formation. The Kohat Formation overlies the Kuldana in upper parts of Potwar area and is missing in the Kalachitta and Hazara areas. The Margalla Hill Limestone (Latif 1970a, Fatmi 1973, type locality Burjianwala) consists largely of grey, medium-bedded to massive, nodular limestone with subordinate marl and shale. The Formation ranges in thickness from about 80 m to 120 m and contains *Assilina granulosa, A. spinosa, A. dandodica, Discocyclina ranikotensis, Lockhartia conditi, L. hunti, L. tipperi, Fasciolites delicatissimus, F. ellipticus, Lepidocyclina punjabensis, Nummulites atacicus, N. globulus, N. mammilla, Operculina jiwani, O. patalensis, Rotalia trochidiformis* etc. The Margalla Hill Limestone is Early Eocene in age.

Balakot Formation: In the Hazara"Kashmir Syntaxis, the basal part of the Murree Formation comprises cross-bedded sandstone and laminated and bioturbated sand and siltstone of shallow marine to tidal environment. These beds contain Late Paleocene to Early Eocene nummulites and assilines. Bassart and Ottigar (1989) have named this sequence Balakot Formation.

The overlying Chor Gali, Kuldana and Kohat Formations have been described earlier. The Paleogene sequence in the Potwar, Kalachitta and Hazara area is unconformably overlain by the Neogene formations.

Neogene

Kakar Khorasan–Makran Flysch Belt

In the Kakar Khorasan–Makran region, the Neogene sequence consists mostly of flysch sediments comprising basin turbidites, slope, shelf and near shore deposits. These are capped by Pleistocene molasse-type deposits. HSC (1960) divided this sequence into a number of formations the names of which were adapted by other workers (Harmes et al. 1984, Raza et al. 1990). However, the Geological Survey of Pakistan (Cheema et al. 1977) have classified these units into fewer sub-divisions, shown in Figure 5.8. Based on this classification, the Neogene sequence of the region is described below.

Hinglaj Formation (Vredenburg 1909a, Cheema et al. 1977): This Formation forms the lower part as well as the bulk of the Neogene succession in Balochistan. The Hinglaj Mountains are considered as the type locality. The Formation rests transitionally on the Oligocene Khojak Formation, though in the southeastern part of the region, the Hinglaj Formation overlies unconformably the Nisai Formation, the Parh Limestone and other older rocks, with a conglomerate at the base. The Formation consists of sandstone with mudstone, shale, shelly limestone and minor conglomerate. The sandstone is grey to brown, thin- to thick-bedded, cross-stratified, ripple-marked, calcareous and protoquartzitic. The shale is grey to brown and grades into mudstone. The limestone is commonly present in the lower part and it is grey, shelly to coquinoid and argillaceous to sandy. In the Makran region, the Formation includes a lower **Parkini Mudstone Member** which is nodular and dark grey, and an upper **Chatti Mudstone Member** which is more calcareous and marly. The thickness of the Formation ranges from 3,303 m to over 4,545 m. The Formation contains foraminifera, gastropods and bivalves (Hunting Survey Corp. 1960). The age of the Hinglaj Formation ranges from Miocene to Pleistocene.

Gwadar Formation (Vredenburg 1921, HSC 1960, Cheema et al. 1977): In Gwadar Peninsula this Formation overlies the Hinglaj Formation unconformably. It consists of soft, poorly consolidated, poorly-bedded or massive, buff coloured sandy clays, interbedded with subordinate brown medium- and coarse-grained poorly-consolidated, thick-bedded, sandstone and a few beds of conglomerate. The thickness of the Formation ranges from 60 m to about 900 m. It contains foraminifera, bivalves and gastropods and has been assigned a Late Pliocene to Pleistocene age.

Sulaiman–Kirthar Fold Belt

Gaj Formation: In the Kirthar and Sulaiman Ranges, the lower part of the Neogene consists of marine near-shore to estuarine sediments of the Gaj Formation (Blanfold 1876, Williams 1959, Pascoe 1963, Cheema et al. 1977). The type locality is at the Gaj River. The Formation rests conformably and transitionally over the Nari Formation. In the Sulaiman Ranges, however, it oversteps and rests unconformably on older rocks at some places.

The Gaj Formation is mostly shale which is variegated, grey and gypsiferous. It is interbedded with grey to brown, calcareous, ferruginous and cross-bedded sandstone and fossiliferous brownish, argillaceous limestone. However, in the southern part of the Kirthar Range in the Karachi area, the Formation predominantly consists of yellowish-brown sandstone and cream-coloured or pinkish-white argillaceous limestone. Its thickness ranges from about 90 m in the Quetta area to 600 m in the Kirthar area. The Formation contains foraminifera, molluscs, echinoids, bryozoans, corals and other fossils (Duncan and Sladen 1885, Nuttall 1926, Vredenburg 1928, Hunting Survey Corp. 1960, Khan 1968, Cheema 1977). Some of the more significant fossils include *Miogypsina globulina, M. cushmani, Lepidocyclina marginata, L. blanfordi, Ostrea vestita, Glycimeris (Pectunculus) sindensis, Breynia carinata, Clypeaster depressus* and *Caleloplearus frobesi*. The Gaj Formation is Early Miocene (Aquitanian to Burdigalian) in age but in places may extend into Middle Miocene (Pascoe 1963, Khan 1968).

Himalayan Fold-and-Thrust Belt

Rawalpindi Group: The Neogene sequence in the Kohat–Potwar region consists entirely of fluvial sediments of the Siwalik Group and the Rawalpindi Group (Pinfold in Cheema et al. 1977). The rocks of the Siwalik Group have been described earlier. The Rawalpindi Group rests disconformably on various Eocene formations and it correlates with Gaj Formation of the Kirthar and Sulaiman region. It includes the Murree and Kamlial Formation. In Hazara–Kashmir Syntaxis, Paleocene to Eocene silt and sandstone with forams occur at the base of the Murrees (Bossart and Ottigar !989).

Murree Formation (Wynne 1874, Pilgrim 1910, Cheema et al. 1977; type section north of Dhok Maiki) is a thick monotonous fluvial sequence of red and purple clay, and interbedded greyish sandstone with subordinate intraformational conglomerate. The thickness of the Formation increases from 180 m to 600 m in the Salt Range to 3,030 m in northern

Potwar. It is poorly fossiliferous though plant remains and some vertebrate bones have been found including those of *Anthracotherium bugtiense, Brachyodus giganteus, Palaeochoerus pascoei, Hemimeryx sp.* and *Teleoceras fatehjangensis*. This fauna indicates an Early Miocene age.

Kamlial Formation (Pinfold 1918, Pascoe 1963, Lewis 1937, Fatmi 1973, Pascoe 1963, Cheema et al. 1977; type section southwest of Kamlial), overlies the Murree Formation conformably and transitionally; though at some localities it lies unconformably on the Eocene Sakesar Limestone. The Formation consists mainly of grey to brick-red medium- to coarse-grained sandstone interbedded with purple shale and intraformational conglomerate. A number of mammalian fossils have been found including *Listriodon pentapotramiae, Hemimeryx blanfordi, Anthracotherium sp., Trilophodon cf. angustidens, Dinotherium indicum* and *hyaenae lurus lahirii* (Pascoe 1963). The age of the Kamlial Formation is Middle to Late Miocene.

Siwalik Group: In the Kirthar and Sulaiman Ranges, the Gaj is overlain by the Siwalik Group (Meddlicot 1864, Pilgrim 1913, Lewis 1937, Cheema et al. 1977) which is composed of molasse-type sediments. It overlies the Gaj Formation conformably, though at several places, particularly in the Sulaiman Range, it overlaps the older Formations. An angular unconformity may be seen at some places. In the Kohat–Potwar region this Group rests conformably on the Rawalpindi Group. The term Siwalik Group now includes the Manchar Series of Blanford (1876) and Vredenburg (1906), the Sibi and Urak Groups of HSC(1960) and the Uzhda Pasha Formation, the Shinmati Formation and the Urak Conglomerate of Kazmi (1961). The Siwalik Group includes (in ascending order) the Chinji, Nagri, Dhok Pathan and Soan Formations and the Lei Conglomerate.

The **Chinji Formation** (Pilgrim 1913, Lewis 1937, Eames 1952, Cheema 1977; type section south of Chinji village) is confined to the eastern Sulaiman Range and the Kohat–Potwar region and is not developed in the Kirthar area. In the Sulaiman Range it overlies the Nari Formation disconformably and consists mainly of red clays with subordinate grey to brown, fine- to medium-grained, gritty, soft, cross-bedded sandstone. The Formation has yielded a rich vertebrate fauna of crocodiles, turtles, monitor lizards, aquatic birds, dinotheres, primitive trilophodonts, suidae, water deer, hominoides, pythons and chelonias (Pilgrim 1913, Mathew 1929, Colbert 1933, Lewis 1937, Pascoe 1963). The age of the Chinji Formation is Late Miocene (Late Tortonian to Sarmatian).

Nagri Formation (Pilgrim 1913, 1926, Lewis 1937, Cheema 1977; type locality Nagri village, principal reference section at Gaj River): This Formation rests conformably on the Chinji Formation in most places in the Sulaiman Range, though elsewhere in the Sulaiman and Kirthar Ranges it overlies the Gaj and earlier formations unconformably. At some places an angular unconformity may be seen. The Formation is largely thick-bedded to massive, greenish-grey, medium- to coarse-grained, salt and pepper textured, calcareous sandstone, interbedded with subordinate brown to reddish sandy clay and conglomerate. The thickness of the Formation ranges from 200 m to 3,000 m. The Formation has yielded a rich vertebrate fauna including proboscideans, rhinoceratides, crocodiles, chelonians and artiodactyles. From the Potwar area perissodactyles, carnivores and primates also have been reported (Pilgrim 1913, 1926, Anderson 1928, Colbert 1933, Lewis 1937, Gill 1952, Pascoe 1963). The age of the Nagri Formation is Early Pliocene (Pontian).

Dhok Pathan Formation (Blanford 1876, Griesbach 1893, Pilgrim 1913, Cotter 1933, Cheema et al. 1977; type section at Dhok Pathan village): It has a transitional contact with the underlying Nagri Formation and consists of a thick pile of interbedded sandstone and clay. The sandstone is thick-bedded, grey to brown, calcareous and cross-bedded; the clay is brown, orange to red. Lenses of conglomerate are present in the upper part. Its thickness ranges from about 1,330 m to 2,000 m. A rich vertebrate fauna has been obtained from the Formation. It is less fossiliferous in the Sulaiman and Kirthar Ranges where *Hipperion punjabiense, Rhinoceros sivalensis* and *Pachyportax latidens* have been found. The Potwar region has, however, yielded a much richer vertebrate fauna including carnivora, proboscidea, perissodactyla and artiodactyla (Pascoe 1963). The age of the Formation is Early to Middle Pliocene.

Soan Formation (Meddlicot 1864, Blanford 1976, Pilgrim 1913, Kravtchenko 1964, Cheema et al. 1977; type section near Mujahad village, principal reference section near Urak village): This Formation rests disconformably on the Dhok Pathan Formation with a marked coarsening of the clastic sediments and appearance of massive conglomerate beds. The Formation largely consists of massive conglomerate with subordinate sandstone, siltstone and clay. The pebbles and boulders in the conglomerate range from 5 cm to 30 cm. The Soan Formation is 300 m to 3,000 m thick. It is poorly fossiliferous and vertebrates such as *Mastodon sivalensis, stegodon clifti, Elephas planifrons, Sivatherium giganteum, Proamphibos lachrymans, Dichoryphochoerus durandi* and *Sivafelis potens* have been reported from the Potwar region. The age of the Soan Formation is Late Pliocene to Early Pleistocene (Astian to Villafranchian).

Quaternary

Quaternary stratigraphic sequence in Pakistan represents a wide range of depositional environments including the marine coastal deposits, shore and offshore deposits, the volcanic deposits of the Koh-i-Sultan, the aeolian deposits of the Thar and other deserts, evaporites of the salt lakes in Sindh, playa and lacustrine deposits in the intermountain basins, deeply-weathered residual soil in the Himalayas, glacial and fluvioglacial deposits in the Himalayas, Karakoram and Hindukush Ranges, and a complex of fluvial deposits in the vast piedmont zone, flood plain and delta of the Indus River.

Makran Coast

On the Makran Coast the Late Pliocene to Pleistocene Gwadar Formation is overlain by the **Jiwani Formation** (Hunting Survey Corp. 1960; type section east of Jiwani) with an angular unconformity on top of the Gwadar Formation. At places it overlies the Hinglaj Formation unconformably. The Jiwani Formation consists of littoral deposits of shelly limestone, sandstone and conglomerate. The limestone is comprised of shelly fragments set in a sandy calcareous matrix. It is medium- to thick-bedded, hard and porous. The sandstone is thin- to thick-bedded, cross-laminated, pebbly, medium- to coarse-grained and well sorted. The Formation is about 30 m thick and contains bivalves such as *Fossularca pectunculiformis, Timoclea arakanensis, Martesia striata, Ostrea pseudocrassissima* and other fossils. The Jiwani Formation has been assigned Pleistocene to Sub-Recent age by the HSC (1960).

The Makran coast is characterised by a number of Pleistocene to Holocene marine terraces and these have been discussed earlier in Chapter 2.

Volcanic Deposits

The Chagai District contains a number of extinct volcanoes such as Koh-i-Sultan, Damodin, Koh-i-Dalil, Chota Dalil, Mit Koh, Koh-i-Malik and many smaller, satellite cones or plugs (Fig. 6.36, Photo. 15). These volcanic centres are surrounded or covered with volcanic rocks- the **Koh-i-Sultan Volcanic Group** (HSC 1960), which are largely comprised of agglomerate and tuff and subordinate lava (see Chapter 6). Because these deposits are not folded and the volcanic cones are well preserved, it is inferred that they are not older than Pleistocene.

Thar Desert

The Thar Desert, along the eastern margin of the Indus Plain, is largely covered with longitudinal and complex seif-type stabilised sand dunes (Fig. 2.18). Some of the longitudinal sand dune ridges have a relief of as much as 100 m and are more than 40 km long (Kazmi 1977, 1985). The interdunal depressions are filled with playa-type Recent sediments or contain salt lakes. The aeoline deposits of the Thar and Cholistan Deserts range in age from Pleistocene to Recent (see also Chapter 2).

Intermontane Basins

In the intermontane basins a thick sequence of conglomerate with subordinate sandstone and some mudstone commonly rests unconformably over the Neogene or earlier deposits. These conglomerates have been deformed and folded and form basinal deposits. In the Balochistan Basin they have been named **Haro Conglomerate** (HSC 1960; type locality in the Kech Valley in the Kharan District). In the Indus Basin similar conglomerates have been named **Lei Conglomerate** (Gill 1952, Cheema et al. 1977; type locality Lei River section southeast of Rawalpindi, principal reference section near Spintangi Railway Station). They were previously given different names by various authors (Pilgrim 1910, Gee 1945, HSC 1960, Kazmi 1961, Pascoe 1963, Kazmi et al. 1970). The Conglomerate is comprised largely of rounded to subrounded pebbles and boulders. Its thickness ranges from about 150 m to over 900 m. The Lei Conglomerate has yielded pre-Soan-artifacts and vertebrate fossils including *Elephas hysudricus, Sivatherium giganteum, Dicerorhinus platyrhinus, Equua sp., Bos sp.,* and *Camel sp.* (Pilgrim 1913, Wadia 1928, Pascoe 1963). The Lei Conglomerate has been related to the last major episode of the Himalayan Orogeny and has been assigned a Pleistocene age (Pascoe 1963, HSC 1960).

In the Quetta area a thick sequence of folded lacustrine to playa-type deposits form the valley fill and unconformably overlies the Neogene and earlier rocks. These deposits have been named the **Bostan Formation** (HSC 1960; type sections at Pishin Lora and Kamerod village in Siahan Range). This Formation overlies the older formations (including Jurassic), generally with a marked angular unconformity. However in Dalbandin Valley it is believed to be transitional with the Khojak Formation. It consists of gently folded basinal deposits that were laid down in tectonic depressions. It consists mainly of white, grey to brick-red, poorly consolidated gypsiferous clays interbedded with subordinate reddish-brown to grey, salt and pepper textured thin-bedded sandstone and conglomerate. No fossils have yet been reported from the formation. The Bostan Formation is believed to be Pleistocene in age.

In some regions of Pakistan such as Hazara, Swat, the Pir Panjal Range and the western Himalayas, remnants of relict landscapes in the form of ancient valley or basin floors that have escaped erosion are still preserved (Grinlinton 1928, DeTerra 1935, Porter 1970, Lawrence et al. 1985). Some of the relicts like those of the Chattar Plain of Hazara are covered with thick residual soils. In the Chattar Plain, the upper part of the soil comprises a thick clay-rich saprolite with core stones, at depth resting over disintegrated or jointed bed rock. These soils have formed over a prolonged period of tropical weathering in an area of low relief. Lawrence et al. (1985) believe that these soils developed as a result of prolonged weathering between 20 to 5 m.y. ago and were intermittently eroded due to the Himalayan uplift of the past 5 m.y. We are, however, of the view that these relict residual soils are essentially of Pleistocene age. It is most unlikely that these soils could have escaped erosion due to severe tectonics and uplift, particularly during the past 4–5 m.y. (Zeitler et al. 1982).

The intermontane basins and the mountain valleys outside the glaciated areas are largely covered with Quaternary alluvial valley fill which, at many localities, consists of interbedded fluvial, aeoline (loessic), lacustrine and playa deposits (Kazmi et al. 1970, Burbank et al. 1985).

The Peshawar, Kohat, Potwar, Bannu, Zhob, Pishin and Kharan are amongst the largest intermontane basins in Pakistan. The Quaternary geological record of these basins is apparently dominated by tectonic rather than climatic events (see Chapter 4). There is no evidence of glacial advance in any of these or other basins which are at altitudes of below 1,500 m. The lowest limit of Pleistocene glacial advance in the Himalayas and Karakoram has been established between 2,400–1,500 m by previous workers (Middlemiss 1896, Norin 1925, Porter 1970). Pleistocene glacial deposits are therefore confined to the smaller basins and valleys of the Himalayas and the Karakoram and we have discussed them in the following sections.

Almost all the intermontane basins of Pakistan are characterised by a typical loessic silt which is interbedded with alluvial deposits. According to Rendell (1992) it provides a terrestrial record of continental aridity and has been correlated with glacial or pluvial episodes (DeTerra et al. 1939). Thermoluminiscence (TL) dating of the loess deposits of Potwar and Peshawar Basins, commonly referred to as the **Potwar Loess**, has provided the following chronology (Rendell 1992).

18,000 yr BP.	–End of loess deposition in Pakistan. (It coincides with last glacial maximum in NW Europe).
75,000–18,000 yr BP	–Main phase of loess deposition (unweathered loess).
170,000 yr BP	–Early phase of loess deposition.

Apart from the Potwar Loess, the Potwar region bears other evidence of Himalayan glaciation in the form of huge erratics of granites, gneisses, basalts and ultramafic rocks (Photo. 30) which dot the landscape along the Indus, Haro and Sil Rivers in NW Potwar, and are believed to be ice-rafted glacial remnants brought down due to catastrophic floods (Wynne 1879, Desio 1983).

Glacial deposits in the Himalayas and the Karakoram

Pleistocene to Recent glacial moraines, fluvioglacial deposits, loess, glacial lake deposits, ancient rock terraces as well as fluvial terraces are commonly seen in the upper reaches of the Chitral, Swat, Hunza and Indus Valleys. The Quaternary deposits in the upper Indus

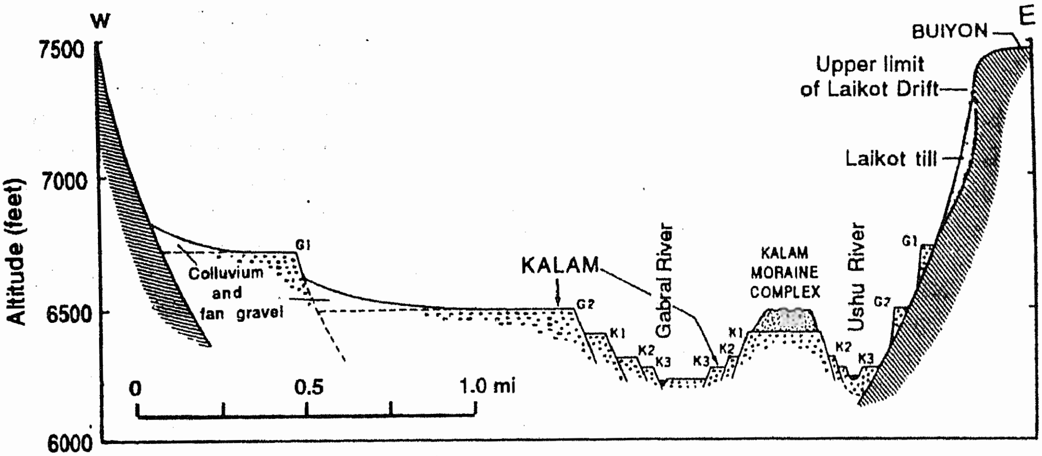


Fig. 5.9. Cross section through glacial terraces at Kalam, Swat (from Porter 1970).

Valley have been described by Olson (1982), and Cronin (1982, 1989), in the Swat area by Porter (1970), and in the Kashmir–Potwar area by De Terra and Petterson (1939). The latter established four major glaciations in the Himalayas during the Pleistocene on the evidence of distinct glacial moraines. Shroder et al. (1989) have, however, described three major episodes of Pleistocene glaciations in the Indus Valley between Darel and Gilgit.

In the Upper Swat region, above an altitude of 2,280 m the valley floor morphology is largely attributed to glacial erosion and remnants of ancient valley floor are preserved as high-level bedrock benches near Kolalai, Laikot, Kalam and Utrot. Between 2,280 m and 1,770 m the valleys are covered with glacial deposits which are up to 200 m thick. Based on moraine and terrace morphology, weathering characteristics and degree of soil development, the glacial deposits have been differentiated into three principal drift sheets. These have been named **Laikot** (which is oldest), **Gabral** and **Kalam** Drifts (Porter 1970). Each of these glacial sequence has been further subdivided into lesser stratigraphic units as shown and described in Figure 5.9 and Table 5.8. Porter and Rendell (quoted in Shroder 1992) consider the Swat glaciations to be relatively young, perhaps Middle to Late Pleistocene and Holocene.

From the Skardu Valley of Baltistan, Cronin (1982) has described an Early Pleistocene sequence, the **Bunthang Sequence**, which consists of a till at the base, overlain by a locally deformed fluvial to glacio-lacustrine conglomerate and sandstone. It grades upwards, into grey massive siltstone and mudstone (Fig. 5.10). The Middle Bunthang Sequence rests on the lower one and is comprised of three wedge-shaped units of boulder conglomerates. It is overlain by the Upper Bunthang Sequence of fine-grained sandstone and massive mudstone interpreted as braided stream deposits. The Bunthang Sequence is about 1,200 m thick and is overlain by alluvial terrace deposits. The upper age limit of the sequence, based on magnetic polarity studies (end of Matuyama Chron) is 0.73 m.y. (Cronin et al. 1989).

The sediments above the basal till show a reversal of magnetic polarity and have been tentatively correlated with the Jaramillo subchron (0.91–0.98 m.y.) by Cronin et al. (1992). These authors correlate the Bunthang Sequence with the Jalipur Till (Fig. 5.11).

The Quaternary deposits of the upper Indus Valley between Darel and Gilgit and in the Hunza Valley include glacial moraines, tills, and fluvioglacial deposits, lake deposits and fluvial deposits ranging from Pleistocene to the Recent (Derbyshire et al. 1984, Shroder et al. 1989).

Table 5.8. Pleistocene stratigraphy of Swat Kohistan (from Porter 1970).

	Rock-Stratigraphic Units	Geologic-Climate Units
Kalam Drift (K)	Drift comprising the Kalam Moraine complex, the Matiltan Moraine, and their correlatives, and associated outwash bodies extending beyond the end moraines. Named for the town of Kalam, the older part of which is built on outwash terraces of this drift sheet.	Kalam Glaciation
Kalam III drift (KIII)	Drift comprising the Matiltan Moraine, the lowest outwash terrace (K3) at Kalam, and correlative deposits.	Late stade
Kalam II drift (KII)	Drift comprising the K2 moraine of the Kalam moraine complex, the intermediate Kalam terrace (K2) at Kalam, and correlative deposits.	Intermediate stde
Kalam I drift (KI)	Drift comprising the outermost Kalam end moraines of the Kalam moraine complex, the highest Kalam terrace (K1), and correlative bodies of drift in other alleys.	Early stde
Gabral Drift (G)	Drift comprising the Gabral Moraine, the two highest outwash teraces at Kalam (G1 and G2), and correlative deposits. Named for the Gabral River, which flows through drift of this age below Utrot.	Gabral Glaciation
Gabral II drift (GII)	Drift comprising the prominent terrace (G2) on which the newer part of the town of Kālam has been built.	Late stde
Gabral I drift (GI)	Drift comprising the outermost element of the Gabral Moraine, the highest terrace (G1) at Kalam, and correlative deposits in other valleys.	Early stde
Laikot Drift (L)	Drift in the form of isolated patches of deeply oxidized till and scattered erratic stones on mountain slopes above Kalam, and beyond the limit of Gabral Drift. Named for the village of Laikot, which stands at the approximate downvalley limit of the confluent Gabral-Ushu glacier system during this advance.	Laikot Glaciation

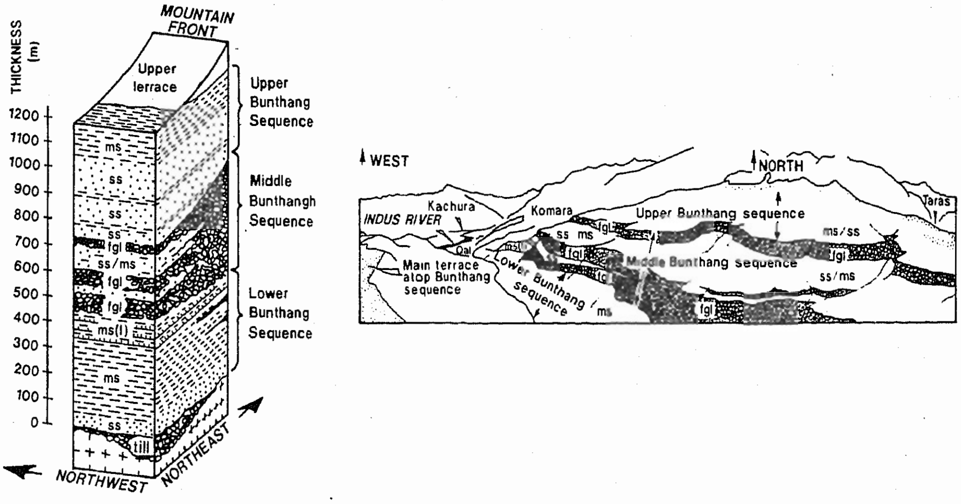


Fig. 5.10. Stratigraphic section and view of the Bunthang sequence in Ghothamal Canyon near Skardu (from Cronin et al. 1989).

Three periods of major glaciation have been identified, with the middle glaciation having two and the last having four phases of ice advance. The geochronology has been compiled using data from thermoluminiscence, magnetostratigraphy and fission track dates. These deposits are shown in Figure 5.11. Zhang and Shi (1980) first drew attention to at least three Pleistocene glacial stages in Batura region (upper Hunza) and named them Shanoz (oldest), Yunz and Hunza. Derbyshire et al. (1984) found a similar sequence in the main Hunza Valley, though they changed the name of the last one to Borit Jheel. They also recognised a fourth glacial stage-Gulkin I. Shroder et al. (1989) traced the continuation of the Hunza glacial sequence to the Indus Valley between Shatial and Gilgit, though, to avoid confusion, they used the terms early, middle and last instead of the geographic names used previously.

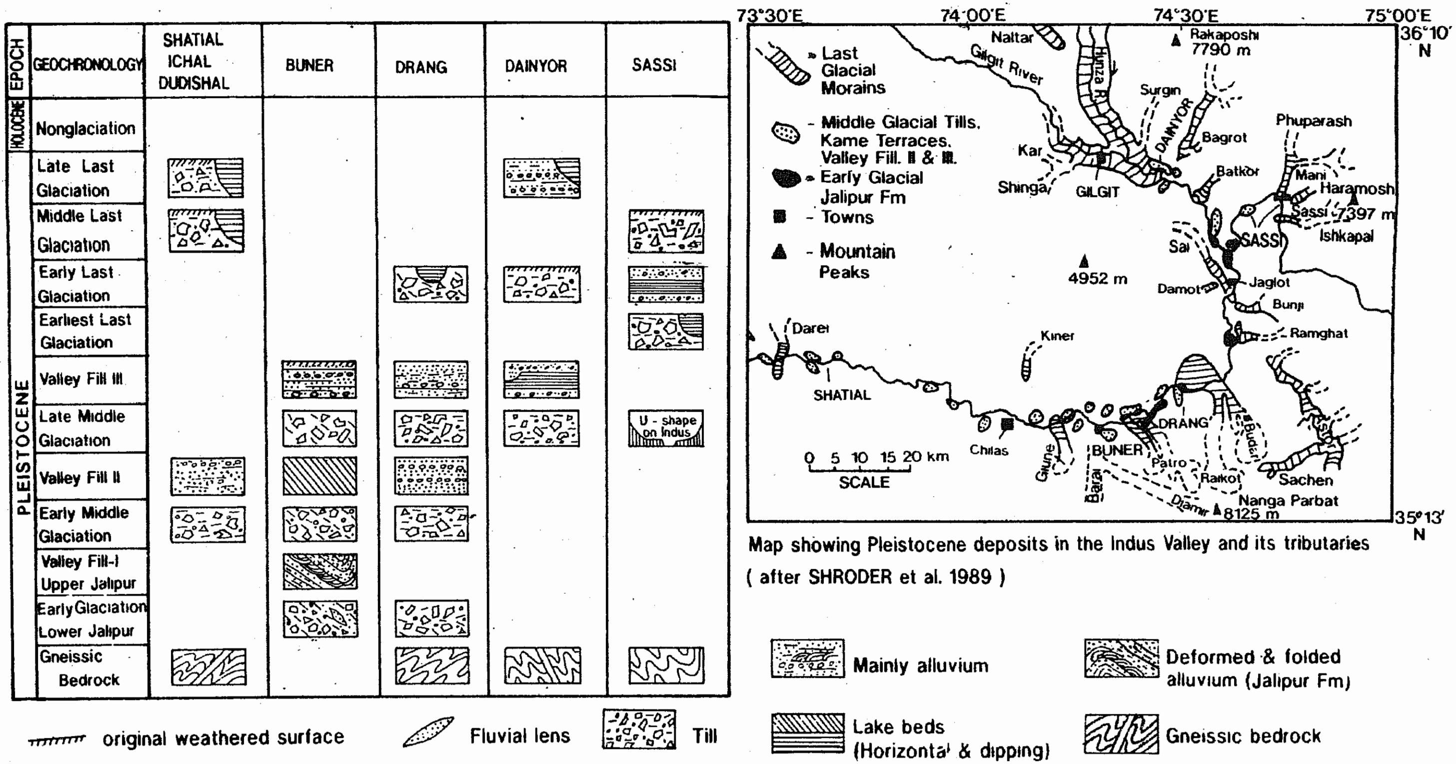
The earliest glacial deposits (**Shanoz**) are found on some of the Pre-Pleistocene surface (valley floors) at altitudes between 2,600–>5,000 m (e.g, the Patundas surface- 4,100–4,200 in Pasu area; Mirschiken benches- 5,445 m opposite Aliabad; glacial benches ~ 2,600 m, near Gilgit). These surfaces with till remnants have been reported in Haramosh, Nanga Parbat, Deosai and Skardu regions also. A sequence of indurated and folded alluvial sediments with till at the base, named **Jalipur Sequence** occurs in the Indus Valley between Bunji and Chilas. It has been referred to the Early Glaciation (Shanoz) by Shroder et al . (1989) on the ground that it contains tillites at the base which donot contain Nanga Parbat Gneiss. The Jalipur sediments were deposited prior to rapid uplift of NPHM and unlike other and younger deposits, this sequence has been indurated and folded. According to Owen (1989a), however, this sequence is younger and is actually a lodgement till and that induration may be due to glacial pressure; if this sequence was earliest, its traces should occur higher up on valley sides also, which is not the case.

Glacial deposits of the Middle Glaciation (**Yunz Stage**) occur on rock terraces and diffluence cols (3,000–3,650 m) in the Hunza Valley. Derbyshire et al. (1984) have shown that these deposits are older than 139,000 TL yr BP and that this was a period of extensive glaciation with glaciers advancing into the Gilgit Valley. Near Gilgit, a lacustrine silt overlying consolidated lodgement till has been dated >100,000 TL yr BP (Shroder et al. 1989). Paleomagnetic data on this same silt suggests that they are younger than 780,000 yr BP (Owen 1989b). Tills from Yunz Glaciation occur in the Gilgit Valley and in the Indus Valley as far down-stream as Shatiyal (Shroder 1992).

Tills from the Last Glaciation (**Borit Jheel**) occur extensively in Upper Hunza Valley below 3,000 m altitude and are predominantly of lodgement and subglacial melt out origin (Derbyshire et al. 1984, Shroder 1992). Glaciolacustrine silts overlie lodgement tills of Batura Glacier, Minapin Glacier and at other places. The glaciolacustrine silts in Batura area give dates of ~ 50,000 ± 3,300 TL yr BP (Derbyshire et al. 1984).

Vestiges of the last glaciation, **Ghulkin I** stade or the fourth glaciation (Derbyshire et al. 1984) were recognised in Upper Hunza Valley. This phase was characterised by "expanded foot" type glaciers and its traces extend into the Gilgit and middle Indus Valley in the shape of large regular lobate landforms comprised of poorly sorted bouldery polymictic diamicrons. At places (e.g., Sachen and Main Glaciers) modern examples of these lobate landforms are directly contiguous to existing ice-cored moraines (Shroder 1992). Lacustrine silts intercalated with tills of this glacial stade from Minapin and Pisan Glaciers have been dated 47,000 ± 2,350 TL yr BP.

In the Upper Hunza Valley, Derbyshire et al. (1984) have identified still younger minor



Map showing Pleistocene deposits in the Indus Valley and its tributaries (after SHRODER et al. 1989)

Fig. 5.11. Quaternary sequence of the upper Indus Valley (from Shroder et al. 1989).

Table 5.9. Quaternary sequence of the Indus Plain and adjacent regions (from Kazmi 1977).

PERIOD		PIR PANJAL (PUNCH.) (DE TERRA et al 1939)	SULAIMAN-KIRTHAR PIEDMONT ZONE	UPPER INDUS PLAIN	LOWER INDUS PLAIN	THAR DESERT
RECENT		Stream bed deposits	Piedmont and Subpiedmont deposits	Flood plain deposits ancient channel deposits of Indus & its tributaries	Flood plain deposits	Barchan dunes of desert fringe zone
EARLY RECENT	ICE ADVANCE	T5 - Terrace deposits	Extension of alluvial fan deposits T5 - Gravel fan terrace	Erosion and degradation Extension of alluvial fan deposits from foot hills onto the Indus flood plain	Degradation	Transverse dunes of desert fringe zone
	INTERGLACIAL	Erosion	Erosion	Aggradation DHAREMA FORMATION (flood plain deposits of Indus and tributaries)	Upper fine sand member of TANDOJAM FM. Lower gravelly member of Tandojam Formation	Lower Barr terrace deposits
PLEISTOCENE	LATE GLACIATION	Terminal moraines Aggradation T4 - Terrace deposits	T4 - Terrace deposits Erosion T3 - Terrace deposits Erosion T2 - Terrace deposits Erosion T1 - Terrace deposits Erosion KACHHI CLAY	Erosion and degradation Upper part of PASRUR, GUJRAT & KHUSHAB CLAYS	Erosion & degradation LARKANA FORMATION (Silt deposits of extended alluvial fans) NABISAR FORMATION (Deltaic deposits)	NABISAR FORMATION Transverse sand dune deposits Longitudinal sand dune deposits
	INTERGLACIAL	Warping - Erosion, followed by T3 - Terrace deposits		Aggradation SHEKHUPURA FORMATION (Flood plain deposits) Upper Barr terrace		
	LATE MIDDLE GLACIATION	Terminal moraines Aggradation. Deposition of gravel & POTWAR Boulder LOESS. T2 - Terrace		CHUNG FORMATION (loessic silt deposits) PASRUR CLAY, GUJRAT CLAY & KHUSHAB CLAY in subpiedmont zones		
	INTERGLACIAL	Erosion T1 - Terrace deposits LEI CONGLOMERATE ? Tilting & folding. Lastphase of Himalayan orogeny	Tilting & folding. last phase of Himalayan orogeny	Aggradation RECHNA FORMATION (flood plain deposits)		
	EARLY MIDDLE GLACIATION	Moraines deposition of Boulder Conglomerate	LEI CONGLOMERATE	Erosion and degradation LAHORE FORMATION (silt & clay deposits - widespread extension of alluvial fan deposits)	LEI CONGLOMERATE	
	INTERGLACIAL	Folding, uplift & erosion PINJOR FORMATION	Folding, uplift & erosion SOAN FORMATION	Aggradation BARI FORMATION (flood plain deposits)	SOAN FORMATION	
	EARLY GLACIATION	Kashmir valley terminal moraines TATROT FM	SOAN FORMATION	TATROT FORMATION ? (sandstone, clay & conglomerate)		

episodes of glacial advances, which have been named Ghulkin II, Batura, Pasu I and Pasu II stades and are largely Holocene in age. Ghulkin II, however, is considered as a late retreat stage of Ghulkin I.

Quaternary deposits of the Indus Plain

The vast-alluvial plain of the Indus River and its tributaries covers an area of about 100,000 km^2 and contains fluvial deposits which are at places over 590 m thick and range in age from Pleistocene to Recent (see Chapter 2). Data from several hundred water wells drilled in the Indus Plain (Kazmi 1964) provides a lucid picture of the nature, structure and stratigraphy of these deposits. Kazmi (1966, 1977, 1984) has worked out the Quaternary stratigraphy of the Indus alluvium and has correlated it with the Himalayan Quaternary sequence (Table 5.9). This chronology is based on the following hypothesis.

Present day ephemeral or intermittent streams draining the outer hill ranges around the Indus Plain carry heavy silt load, which are deposited as wide and thick silt and clay deposit in the piedmont zone (Fig. 2.19). The Indus and its major perennial tributary streams,

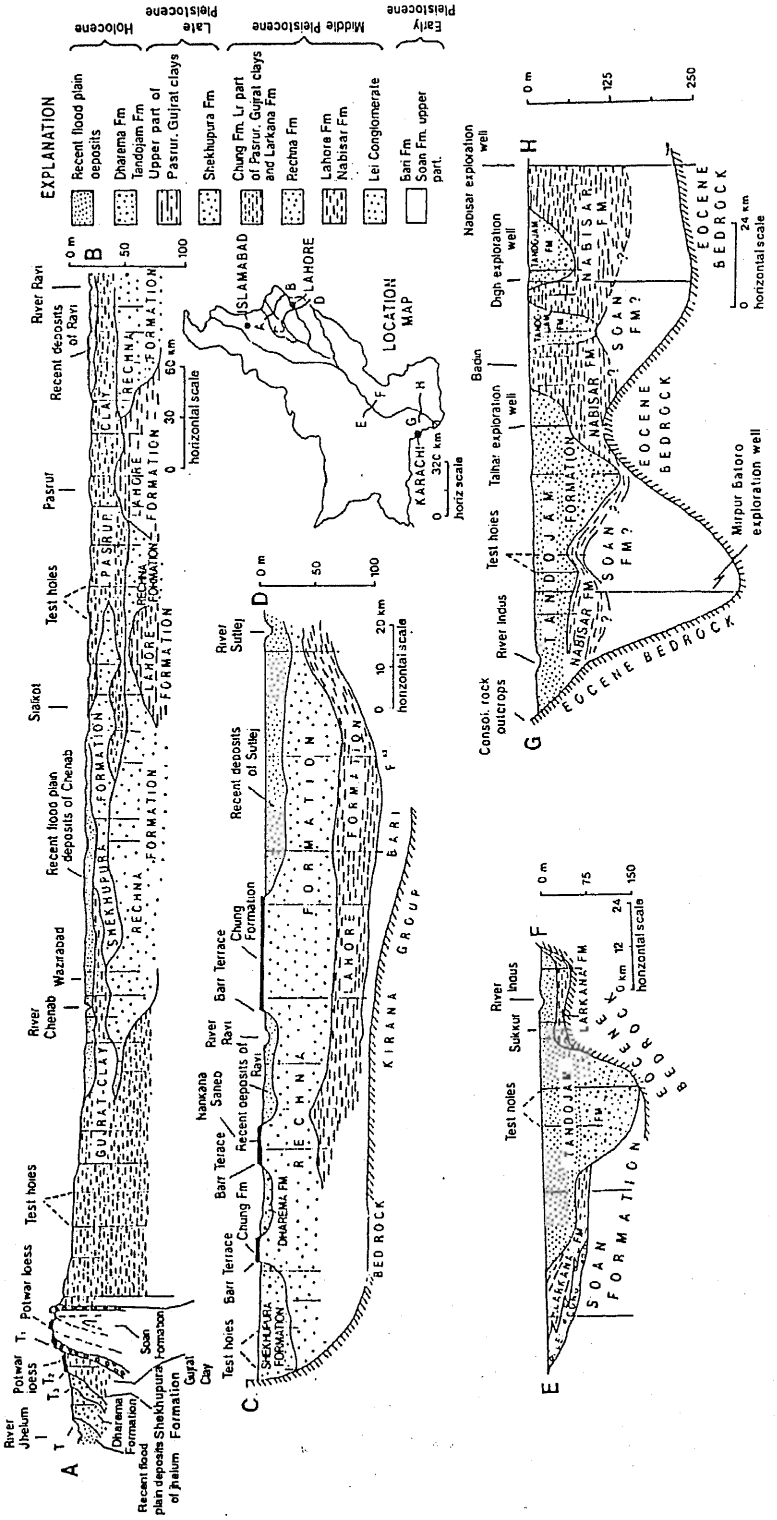


Fig. 5.12. Subsurface geological sections of the Quaternary sequence of the Indus Plain (from Kazmi 1977).

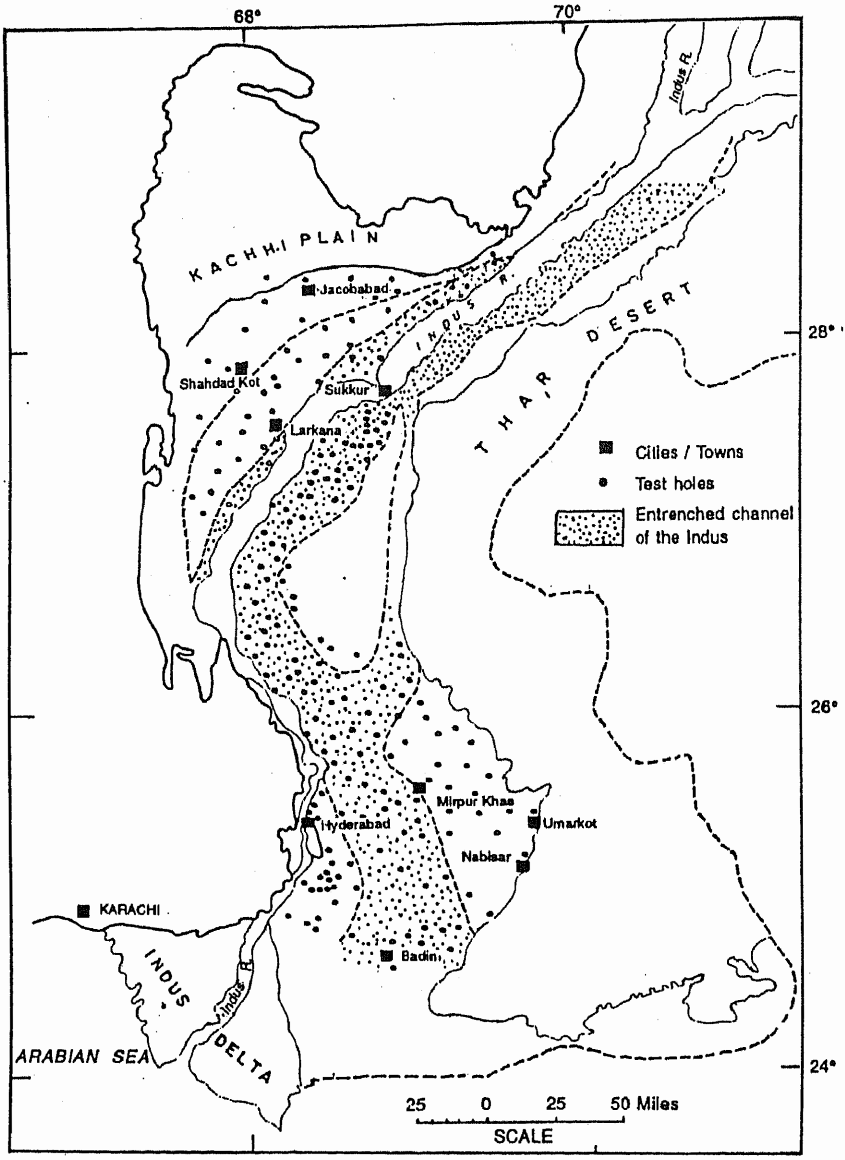


Fig. 5.13. Map of the lower Indus Plain showing location of test-holes and the Late Pleistocene buried channel of the Indus (from Kazmi 1984).

however, carry a relatively low silt and high sand load (Fig. 2.16). During glacial periods most of the precipitation in the catchment of the perennial Himalayan rivers would have been in the form of snow, thus reducing the seasonal flood discharge and the deposition on the flood plains. On the other hand during the glacial periods (pluvial in Indus Plain), the precipitation on the outer hill ranges fringing the Indus Plain, and the deposition of silt in the Piedmont zone would have increased. The silt and clay deposits (**Pasrur**, **Gujrat** and **Khushab Clays**, **Larkana** and **Lahore Formations**; Fig. 5.12) were thus formed and overlapped the flood plain deposits. Lowering of the sea level during each glaciation would have degraded and entrenched the Indus and its tributaries into relatively narrow channels, forming terraces and causing extensive desertification amongsts other physiographic changes. The **Potwar Loess**, **Chung Formation** and the aeoline sand deposits of the Thar and Cholistan Deserts were thus formed. Each interglacial period must have caused widespread aggradation, extensive melting of the ice, dispersal of ice rafted boulders and erratics (Photos. 30 and 31), channel filling with coarser poorly sorted sand (**Tando Jam Formation**, **Dharema Formation**, **Shekhupura Formation** and **Rechna Formation**), changes in river courses, shifts in the position of the Indus Delta and the formation of wide flood plains and deltaic deposits (**Nabisar Formation**).

The Jacobabad–Khairpur upwarp divides the Indus Plain into two parts which are tectonostratigraphically different. Whereas the lower part is comprised of a sloping monocline disrupted by extensional faulting, the northern part is traversed by the Sargodha–Shahpur buried ridge (Fig. 4.6), and the subsurface structure is more complicated. The upper Indus Plain is the product of erosional-cum-depositional activities of the five major rivers (see Chapter 2) and there is a much higher concentration of sand in the subsurface Quaternary deposits. Lithologic logs of hundreds of test-holes reflect sharp differences in the nature and structure of the unconsolidated subsurface sediments of the two parts of the Indus Plain and these are therefore described separately in the following sections.

Lower Indus Plain

Lei Conglomerate (Gill 1952): In the Kirthar foredeep on the western side of the lower Indus Plain, the Lei Conglomerate (Dada Conglomerate of HSC 1960) overlies the Siwalik Group and earlier formations with a sharp angular unconformity. It is comprised of thick beds of boulder conglomerate and subordinate coarse, crossbedded sandstone. It is essentially a piedmont outwash deposit. In the foot-hill region and in the Piedmont zone of the Lower Indus Plain, it is overlain by younger unconsolidated piedmont deposits. Towards the Indus Plain it interfingers with the silt and clays of the Larkana Formation (Fig. 5.12). It is unfossiliferous and its age has been inferred mainly from its relationship to the Himalayan Orogeny. In the lower Indus Plain, Kazmi (1984) refers it to the early Middle Glaciation.

Nabisar Formation (Kazmi 1966, 1984): South of Nawabshah, a subsurface sequence of more than 125 m thick deltaic hard clay, interbedded with lenses of very fine sand, overlies the Eocene bed-rock and the Siwaliks (Manchhar Formation of Blandford 1876). The clay is laminated, ranges in colour from dark grey, brown, greenish brown to greyish white. It is calcareous and at places contains abundant mollusc shells, resembling the present day deltaic deposits. The exploratory oil well at Nabisar, and other groundwater test holes between Mirpurkhas and Nabisar and farther northwards, encountered a thick sequence of this Formation, which has been named after the town of Nabisar. According to Kazmi (1984) this deep inland position of the deltaic deposits is due to the rise in sea level and consequent inland encroachment of the delta during the last and earlier interglacial stages.

Larkana Formation (Kazmi 1966, 1977, 1984): This Formation comprises silt and clay with thin interbeds of very fine sand ranging in thickness from 20 to more than 120 m. The Formation has been encountered in bore holes upstream of Sehwan. Near Khairpur it overlies the Eocene limestone, whereas west of the Indus it interfingers with the Lei Conglomerate and is overlain by the younger piedmont deposits (Fig. 5.12).

Kazmi (1966) interprets it as an ancient subpiedmont deposit and correlates it with the Lei Conglomerate. It was probably deposited during the early Middle, late Middle and Late Glacial periods.

Tando Jam Formation (Kazmi 1966, 1984): Test holes drilled on the lower Indus Plain clearly define a 130 to more than 200 m deep and about 50 km wide buried channel of the Indus (Fig. 5.13). This channel is filled with very fine to coarse sand with pebbles and cobbles. This sand unit has been named Tando Jam Formation (Kazmi 1966), after the town of Tando Jam where the Formation was first recognised and is well developed. This Formation is comprised of an upper fine-sand member and a lower gravelly coarse-sand member. The upper ~70 m of the Formation is uniformly comprised of fine to very fine sand similar to the sand presently deposited by the Indus. Some borehole logs show a transition zone at depth of 70 to 80 m in which sands of all grades with some gravel are found.

The lower gravelly coarse-sand member occurs below a depth of 80 m. This unit is devoid of fine-sand and is largely comprised of medium- to coarse-sand with pebbles and large cobbles (Photo. 31). In the gravel beds facetted boulders up to 25 cm in diameter have been encountered in some of the boreholes. The rock-types represented in these clasts include Eocene limestone, Siwalik sandstone, granite, gabbro, gneiss, phyllite, hornfels, quartzite, and a variety of schist. Most of these rock types are typical of the Himalayan igneous and metamorphic sequences. The facetted pebbles and cobbles indicate a cold glacial regime with considerable advance of the glaciers and the ice-front in the Himalayas. Their presence, hundreds of kilometres to the south and near the delta, is attributed to ice-rafting during catastrophic floods. We have earlier mentioned presence of huge erratics of igneous and metamorphic rocks dotting the landscape in NW Potwar (Photo. 30).

According to Kazmi (1966, 1984), the 130–200 m deep Indus trench formed due to the degradation in the Last Glacial period. It was filled in by the Lower Member of the Tando Jam Formation during the early aggradational stage of the Indus during Early Holocene, when the glaciers retreated and the sea level began to rise. The Upper Member of the Formation is Late Holocene, and ranges from the Pre-Moenjodaro period (± 6,000 BP) up to the present as it includes the Recent flood plain deposits of the Indus. Near Tando Jam, Kazmi (personal communication) noted a large Moenjodaro-type brick and pottery fragments encountered in a wide diameter test hole at a depth of about 30 m, in the Upper Member of the Tando Jam Formation.

In the eastern part of the lower Indus Plain, the Tando Jam Formation overlies the Nabisar and Larkana Formations. In the western part it overlies the Larkana Formation and the Piedmont and Subpiedmont deposits.

Upper Indus Plain

Bari Formation (Kazmi 1966, 1977): Along the eastern extension of the Sargodha–Shahpur buried ridge, south of Lahore in Bari Doab, a number of bore holes have encountered bed rock at a depth of 300 to 350 m (Fig. 5.12). It is comprised of hard reddish brown claystone and siltstone with gravel containing pebbles of granite, chert, quartzite and limestone. Named after Bari Doab, Kazmi (1977) has tentatively correlated it with the Early Pleistocene Soan Formation. It is overlain by the thick clays of the Lahore Formation.

Lei Conglomerate: This Conglomerate is exposed in the foot-hill region, along the margins of the upper Indus Plain and has been described earlier. In the subsurface it is overlain by the thick sequence of Gujrat and Pasrur clays.

Lahore Formation (Kazmi 1966, 1977): This Formation comprises thick, massive clay and silt, interbedded with sand. It occurs at a depth of 165–200 m and overlies the Bari Formation near Lahore. It has been encountered in test holes drilled in the Rechna and Chaj Doabs as well. Unlike the present day sediments, some of the sand in this Formation is brownish grey, much less micaceous and contains greater percentage of epidote and zoisite. The thick silty clays of this Formation extend up to the subpiedmont zone and probably represent an extension of that environment. Kazmi (1966), has thus concluded that this Formation was deposited during a pluvial regime and has correlated it with the Lei Conglomerate.

Rechna Formation (Kazmi 1966, 1977): This Formation consists of 100 m to 160 m thick sand and silt. Over 70 percent of these sediments comprise fine to medium grained grey coloured sand. The sand contains occasional thin layers of gravel which is comprised of kankar, fragments of sandstone, siltstone, slate and schist. There are only occasional or rare clasts of quartzite, granite or other igneous rocks. It is noteworthy that in most of the test holes, the upper most beds of this Formation contain kankar nodules, the occurrence of which suggests an old soil profile and a time gap before the deposition of the later deposits. The sediments of this Formation are similar to those of the present flood plain deposits and in the subsurface, are spread beneath the greater part of the upper Indus Plain. According Kazmi (1966, 1977) they were deposited during the Second Interglacial period (Table 5.9). The Rechna Formation overlies the thick persistent clays of the Lahore Formation. In the subpiedmont zone it is covered by the thick Pasrur Clays or interfingers with the Gujrat Clays, whereas in the central part of the Plain it is variously overlain by Shekhupura Formation, Chung and Dharema Formations, or the Recent flood plain deposits (Fig. 5.12).

Chung Formation (Kazmi 1964, 1966): In the Chaj, Rechna and Bari Doabs, the edges of the southern parts of the Barr Plain comprise the Barr Terraces which are composed of an up to 20 m thick sequence of loessic silt studded with kankar. The lower part of the Formation consists of hard red clay. It contains lenses of sand in the form of old channel fillings and fossil dunes. It overlies the Rechna Formation and has been correlated with the Potwar Loess.

Pasrur, Gujrat and Khushab Clays (Kazmi 1966, 1977): These units comprise thick deposits of clay and silt in the subpiedmont zones of the Pasrur, Gujrat and Khushab region. They contain lenses of channel fill sands. These clays overlie the Lei Conglomerate and southward they interfinger with the Rechna and Shekhupura Formations. They comprise subpiedmont deposits which during the glacial periods extended far down the Indus Plain but during the interglacials their deposition was reduced and they formed a narrow fringe along the mountain front. These deposits range from the Middle Pleistocene up to the Recent.

Shekhupura Formation (Kazmi 1966, 1977): This Formation is largely comprised of fine to medium grained, grey coloured micaceous sand with layers of silt and clay. These sediments represent the ancient flood plain deposits of the Chenab, the remnants of which are exposed between Hafizabad and Shekhupura (Figs. 2.17 and 2.19). They are identical with the present day sediments of the Chenab. Similar deposits of the ancient flood plain of the Jhelum occur in the central part of the Chaj Doab and have been grouped with this Formation. These deposits are up to 80 m thick and overlie the Rechna Formation in the central part of the Plain. In the subpiedmont zone they interfinger with the Pasrur and Gujrat Clays (Fig. 5.12). This Formation has been referred to the Third Interglacial (Table 5.9).

Dharema Formation (Kazmi 1966, 1977): This Formation is comprised of alluvial sand and silt representing relatively younger channel filling of the ancient courses of the Punjab rivers which occur on the Lower Barr Terrace (Figs. 2.17, 2.19 and 5.12). They are probably Early Holocene.

6

Magmatism

Pakistan contains a good record of magmatic rocks ranging from Early Proterozoic to Quaternary; and almost all types of common igneous rocks have now been recognised. In broad terms, most of the igneous rocks of Pakistan can be divided into three major paleogeographic regimes: (1) Pre-Jurassic, related to various orogenic, crustal thinning and rifting processes in the Gondwanaland, (2) Late Mesozoic, associated with spreading and hot spot processes, and (3) Cretaceous to Quaternary, related to convergent processes in the Tethys and at the edge of the "Asiatic" continent.

The Precambrian magmatic rocks of Pakistan can be divided into two groups (1) Early Proterozoic granitic rocks with minor amphibolites, in the Himalayan region, which are probably an extension of those of the Indian and Nepal Himalayas, and (2) Late Proterozoic bimodal magmatic rocks in northern Punjab and southeastern Sindh, the latter being an extension of those of Rajasthan; and ultrapotassic volcanics in the Salt Range. During Early Paleozoic, several granitic plutons were emplaced in the Himalayan foothills. These have their equivalents in Afghanistan and the rest of the Himalayas. Late Paleozoic rifting in the northwestern edge of the Indian plate resulted in generally bimodal basic and acid (with some alkaline) magmatism in Kashmir, Hazara and Peshawar region. There are Jurassic-Cretaceous alkaline plutonic and volcanic rocks in Balochistan which have been related to intraplate magmatism.

The Cretaceous-Tertiary period is marked by a dramatic increase in magmatic intrusions and outpouring of volcanic material in northern and western Pakistan. This magmatism owes its origin mostly to the closure of the Neo-Tethys and ultimate collision of India with "Asia". This resulted in (1) subduction-related magmatism in island arcs, continental margins and, possibly, marginal basins, and (2) obduction of several ophiolitic complexes onto the Indian plate. The formation of the island arcs in the north and west, and their obduction, along with the slices of oceanic crust/upper mantle, resulted in a formidable scenario of accretion and thickening of the continental crust. Despite the closure of the Neo-Tethys, magmatism continued in the Himalayas (Nanga Parbat region) till Plio-Pleistocene, and there are Holocene volcanoes in Balochistan.

In this chapter, we summarise the salient petrological aspects of the Pakistani igneous rocks. Despite considerable progress in our understanding of the magmatism due to enhanced research activity in recent years, there still remain unresolved issues related to petrology and paleotectonic environments of many of the rock units discussed here. Clearly, more detailed field work and isotopic studies are required.

LATE ARCHEAN TO EARLY PROTEROZOIC

Igneous rocks of this age have recently been reported from the Nanga Parbat–Haramosh Massif (NPHM) and Besham area in northern Pakistan. The Kaghan area presumably contains

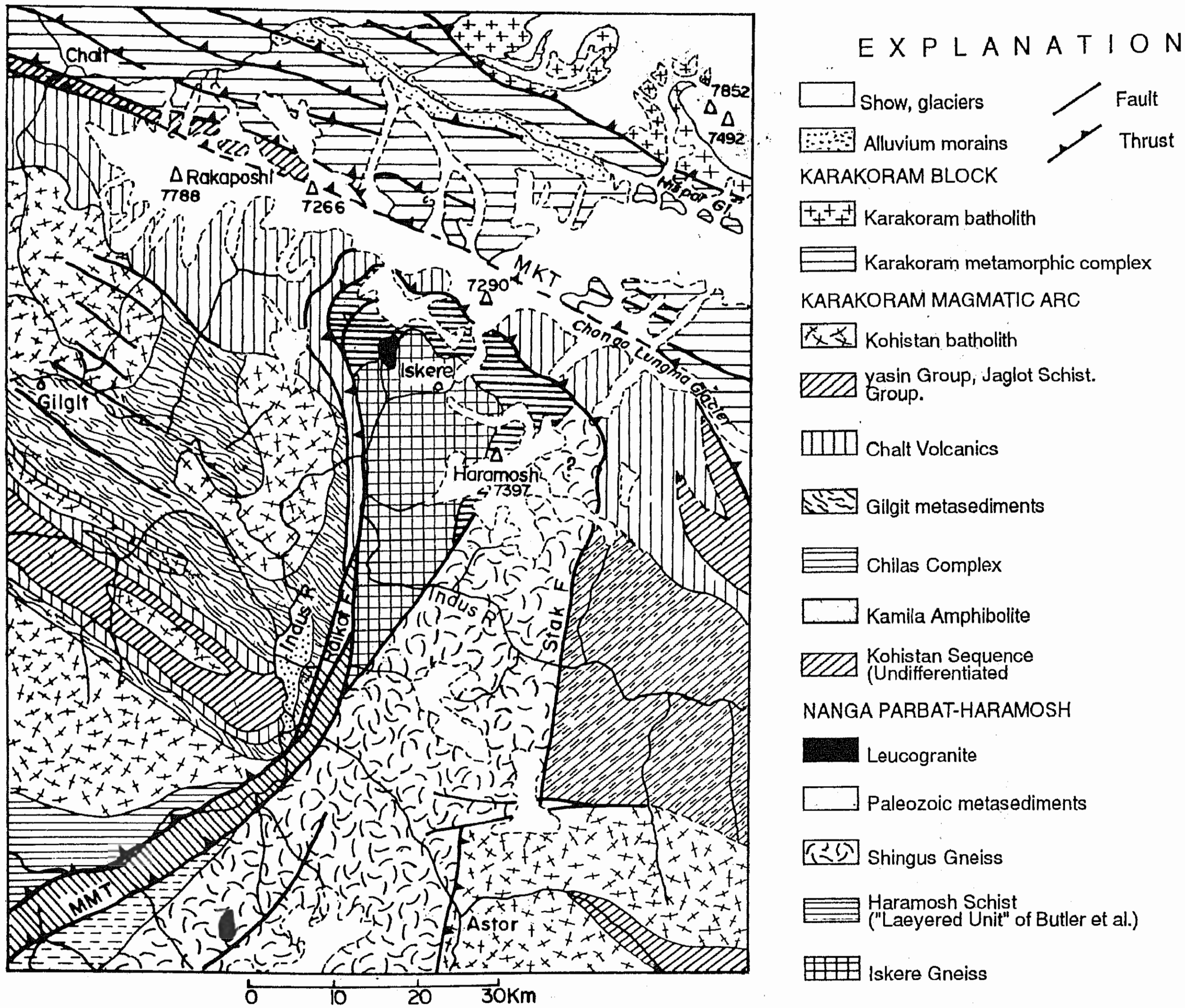


Fig. 6.1. Geological map of the Nanga Parpat-Haramosh Massif and adjacent areas. (Modified from Madin 1986, Verplank 1986, Butler et al. 1992, and Searle and Khan, not dated).

Proterozoic rocks which may be coeval with these. There also are suggestions that the Koh-i-Sufaid crystallines in Kurram and Tirah may be of this age. Further geochronological work may reveal additional occurrences.

Nanga Parbat-Haramosh Massif

The N-S extending NPHM was considered as a promontory (Wadia 1933) or edge of a west-facing embayment (Madin et al. 1989) of the Indian plate. Butler et al. (1992) show that this northern margin of the Indian plate at NPHM is controlled by an anticlinal fold structure (cf., Coward 1985, Coward et al. 1988). The rapid unroofing/rise of the massif has been accommodated initially along ductile shears (i.e., MMT) and subsequently by cataclastic faulting (Raikot or Liachar on the western and Stak on the eastern edge), along with Shahbatot strike-slip zone. The northern part of the massif forms a N-S trending antiform. The Indus Suture, along which rocks of the Cretaceous Kohistan magmatic arc were thrust over the Indian plate during Paleocene, appears to be folded around the NPHM in the form of a loop (Tahirkheli and Jan 1979). On its western margin, however, the Indus Suture was rotated along the still active Raikot Fault where Indian plate rocks are carried back over Kohistan (Fig. 6.1).

The NPHM is a half window of high-grade Precambrian basement gneisses that have been overprinted by Himalayan metamorphism. Correlation of the deformation histories with the Hazara area have been discussed by Treloar et al. (1991). Misch (1949) presented the rocks as a type example of granitisation of batholithic dimension. Shams and Ahmad (1979) reported that the massif is made up of migmatitic gneisses with interwoven micaceous folia (Photo. 32). The gneisses consist of two feldspars, quartz, biotite, muscovite, chlorite, opaque grains and remnants of garnet, kyanite and staurolite of metamorphic origin. According to Rehman and Majid (1989), they are mostly granite and adamellite. Madin et al. (1989) divided the rocks of the northwestern part of the massif into three units.

The Shengus Gneiss, forming the lowest structural unit, consists of fine-grained, finely laminated pelitic and psammitic gneisses with subordinate amphibolites and calc-silicate gneisses (Photo. 33), at least 5 km thick. The Iskere Gneiss is coarse-grained biotite gneiss with subordinate biotite schist, amphibolite, and calc-silicate gneiss, with a minimum thickness of 8 km. The gneisses are medium- to coarse-grained and contain quartz and feldspar megacrysts. The structurally high Haramosh schist unit (the layered unit of Butler et al. 1992), which is more than 2.5 km thick, contains medium- to coarse-grained biotite schist and gneiss, with marble, calc-silicate gneiss, and subordinate amphibolite. The range in lithologies, which form up to a metre thick layers, is the same as that of the Iskere Gneiss, except for a lack of coarse biotite orthogneiss. The amphibolites might be the components of Kohistan interlayered tectonically (Butler et al. 1992), but they can also be Permian. Rare-earth geochemistry (Smith et al. 1992) supports the widely held view that the Nanga Parbat granitic gneisses are derived from a pelitic protolith.

U-Pb zircon ages show that the Iskere Gneiss is about 1,850 Ma and the structurally lower Shengus Gneiss 400–500 Ma in age (Zeitler et al. 1989). Madin et al. (1989) correlate the older unit with the 1,600–1,900 Ma Lesser Himalayan granites (Valdiya 1983), whereas Zeitler et al. (1989) regard that the Shengus Gneiss may be a metamorphosed equivalent of the Mansehra Granite and other Lower Paleozoic S-type granites found throughout the Himalayan foothills.

There are young, generally undeformed, granitic rocks in the NPHM. These are dominantly coarse-grained biotite-muscovite granite pegmatites locally mined for green tourmaline, aquamarine, topaz, and garnet (Kazmi et al. 1985). Tourmaline granite and aplite occur as dykes, lensoid bodies, and selvages on the pegmatites (Shams 1983, Madin et al. 1989). Very young granitic rocks of Pliocene-Pleistocene age have been reported from the massif. These are probably related to very rapid uplift, and denudation which has accelerated over the past 10 m.y. to a maximum of 7 mm/year (Zeitler 1985, Zeitler and Chamberlain 1991, Zeitler et al. 1993).

Besham antiform

A window of Early Proterozoic basement, bounded by steep N-trending faults both on it eastern and western sides, is exposed in the Besham area to the south of the Indus Suture (Fig. 6.2). Following a preliminary account by Jan and Tahirkheli (1969), the Besham area was described by Ashraf et al. (1980), Butt (1983) and several other workers. La Fortune et al. (1992) have presented detailed petrography and geochemistry of the Besham area. From oldest to youngest they divide the rocks into five groups:

1) The Besham Group, which forms the basement sequence, consists of metasediments, quartzo-feldspathic gneisses and sodic quartzo-feldspathic gneisses formed in situ from a sedimentary protolith of variable composition. The sodic gneisses are the equivalent of previously named Lahor Granite (Ashraf et al. 1980).

2) Mafic dykes in (1), now forming lenses and concordant layers commonly less than 2 m in width but reaching up to 50 m. Metamorphosed to epidote-bearing amphibolites, these are tholeiitic and have island arc geochemical affinities. However, Baig and Snee (1991) reported that some of them are basaltic komatiite, locally preserving pillow structures.

3) The third group of rocks consists of cogenetic, small granitic intrusions and associated pegmatites. The Shang and Duber granite/granodiorite intrusions are coarse-grained, foliated, biotite ± hornblende-bearing, and strongly deformed on their margins. The Shorgara pegmatite containing blue and/or white microcline may be related to these granites. This group of rocks is younger than the Besham Group and older than the Karora Group. The basal conglomerate of the Karora Group also contain rare boulders of muscovite-tourmaline granite, probably derived from nearby intrusions yet to be located.

4) The Karora Group conglomerate, calcareous and carbonaceous metasediments, which form the cover sequence and provide evidence for more than one metamorphic event in the area.

5) Undeformed leucogranites intrude both the Karora Group and the Besham Group in two places in the form of small sills, up to 25 m thick. They are medium-grained, equigranular, and made up of oligoclase, quartz, microcline, biotite and accessory pyrite and sphene.

Radiometric ages have been determined on some rocks of the Besham Group. Treloar et al. (1989c) reported an Ar-Ar age of 1,920 ± 24 Ma on hornblende from an amphibolite sheet or pod within a granite body near Duber. Pb isotope ratios suggest that the stratiform exhalative type Pb-Zb deposits of the area formed 2,120 to 2,200 Ma ago and re-equilibrated during 1,950 Ma metamorphism (Shah et al. 1992). Additional Ar-Ar ages determined by Baig (1991) and Baig et al. (1989) have been interpreted as suggestive of one or more metamorphic episodes in the range of 2,031 ± 6 to 1,865 ± 3 Ma. They report that the (Lahor) sodic gneisses were emplaced about 1,500 Ma ago. A 550 ± 20 Ma Ar-Ar age on biotite from the granitic rock north of Duber (Treloar et al. 1989c) suggests that the group

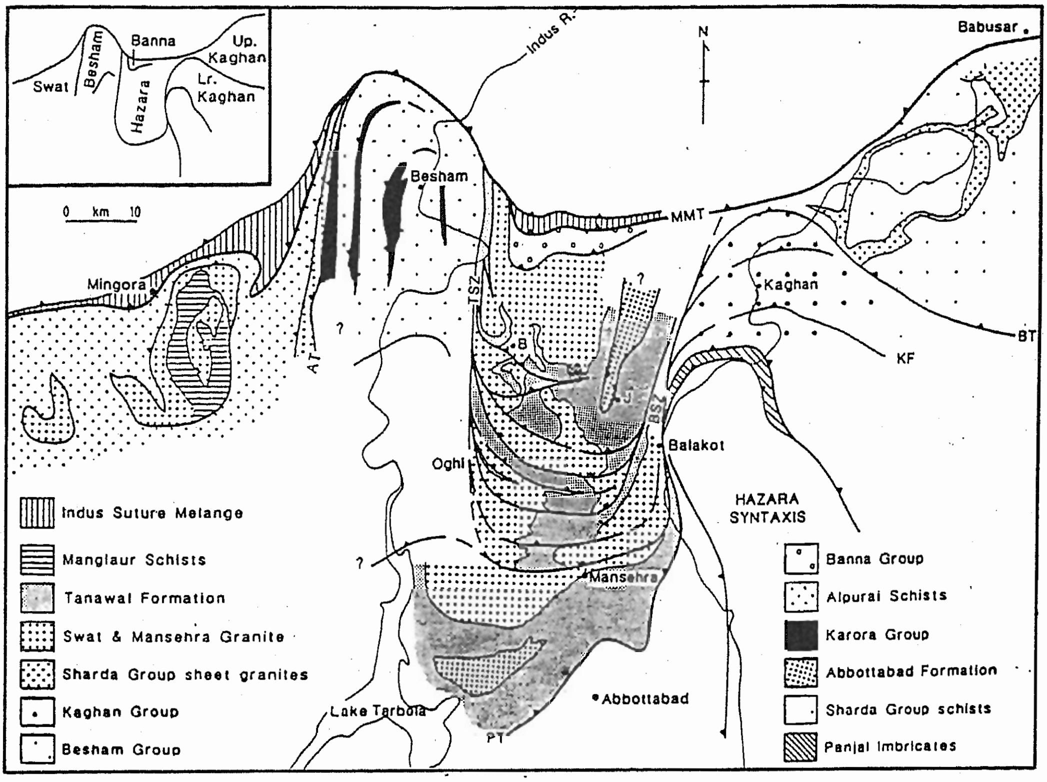


Fig. 6.2. Map of the internal zones of the Indo-Pakistan plate underlying the MMT between Swat and the Kaghan Valley, with an inset showing the boundaries of the tectonostratigraphic blocks. B–Batgram, J–Jabori, AT–Alpuri Thrust, PT–Panjal Thrust, BT–Batal Thrust, MBT–Main Boundary Thrust, KF–Khanian Fault, BSZ–Balakot shear zone, TSZ–Thakot shear zone (from Treloar 1989, Williams 1989).

3 intrusive rocks in the basement gneisses are of the same age as the Mansehra pluton. To sum it up then, the basement gneisses in the Besham and Nanga Parbat areas appear to be coeval. They have an Archean or very Early Proterozoic protolith containing Early Proterozoic granitic and amphibolitic rocks metamorphosed by 1,865 Ma.

Kaghan area

The Precambrian basement of the Kaghan Valley and adjacent area of Azad Kashmir has been studied by several workers (for references, see Chaudhry and Ghazanfar 1987, Greco and Spencer 1993, Chaudhry et al. 1994). The basement appears to pass into that of the Nanga Parbat area in the northeast, but no correlation has so far been proposed between the two. Chaudhry and Ghazanfar (1987) divided the Kaghan basement into (1) Archean to Proterozoic Sharda Group which is thrust along the Main Central Thrust over (2) Proterozoic Kaghan Group, itself thrust over Paleozoic and Mesozoic rocks along the Panjal Thrust. The Kaghan Group, referred to as the Lesser Himalayan Salkhala Formation by Greco (1991), occurs in a band around the Hazara–Kashmir Syntaxis. It consists of unfossiliferous mica schists, calcareous rocks (± graphite ± chlorite) with marbles and calc-silicates, siliceous rocks and amphibolites, metamorphosed up to garnet grade. There are two-micas augen gneisses and granite bodies, possibly Cambrian in age.

The Sharda Group consists predominantly of granetiferous calc-pelites and marbles with subordinate psammitic, pelitic and concordant amphibolitic bands metamorphosed up to sillimanite grade. Abundant granites characterise the group. Chaudhry and Ghazanfar (1987) reported that the area to the east of Kaghan is entirely occupied by the Sharda Group. According to Greco and Spencer (1993), this group of rocks represents a Higher Himalayan crystalline nappe comprising a basement of Cambrian and older granites overlain by Lower Paleozoic rocks and a cover of Paleozoic and Mesozoic metasediments. The amphibolites, according to them, belong mostly to the Panjal magmatism, metamorphosed locally to eclogites.

Chaudhry and Ghazanfar (1987) have described the granitic rocks in considerable detail. They occur in two stratigraphic horizons and have been complexly folded along with their enclosing metasediments and associated amphibolites, but some are discordant. They display considerable textural and some mineralogical variations, with banded microgranite gneiss being the most abundant. There are migmatites, porphyritic to non-porphyritic and deformed granites, aplites and pegmatites, some of which are related to the Himalayan Orogeny. On field and petrographic grounds, Chaudhry and Ghazanfar (1987) concluded that the granitic rocks are S-type and derived from the associated metasediments by partial melting. No radiometric dates are available to compare the granites with those of the neighbouring areas, but they may belong to Proterozoic, Cambrian, and Himalayan (Tertiary) episodes.

Koh-i-Sufaid

A basement of presumed Early Proterozoic rocks occurs in the gently flexuring Parachinar re-entrant. The basement is truncated on its north by the Indus Suture and its west by the Chaman Transform Fault. Much of the basement occurs across the border in Afghanistan, but its southern margin, which is thrust over Mesozoic sedimentary rocks, occurs as a thin strip in Kurram and Tirah (Meissner at al. 1975, Butt 1988). Although radiometric dates are lacking, the basement has been considered to correlate favourably with the Himalayan crystalline schuppen zone (Shroder 1984). Some details on structure were presented by Butt (1988) who equated the rocks with Nanga Parbat and Besham areas.

According to Butt the basement is an imbricate zone containing NW-trending and NE-dipping high-angle reverse faults separating various rock units. Rocks in the adjoining fault blocks show extreme variations from augen gneisses and migmatites displaying ductile deformation to amphibolites, schists, phyllites and dolomites some of which show brittle deformation.

Scanty petrographic information is available on the rocks. The basement, according to Badshah (1983a) and Ahmed (1985a), consists of pre-Paleozoic migmatites, gneisses, schists, calcareous rocks and amphibolites. Biotite gneisses of granitic to granodioritic composition and showing various deformation textures (augen- to "micro-granitic" gneisses and mylonites) are the most common rocks. Some of these and associated pegmatites contain abundant tourmaline. Minor bodies of amphibolites are closely associated with the gneisses, but the amphibolites may be of two generations: 1) Precambrian and 2) Permian rift-related. The basement also contains a younger group of fine- to medium-grained, porphyritic to non-porphyritic granites and granodiorites which are commonly sheared and occur in small bodies. A similar range of rocks has been reported from the

Photo. 32.

Photo. 33.

Photo. 34.

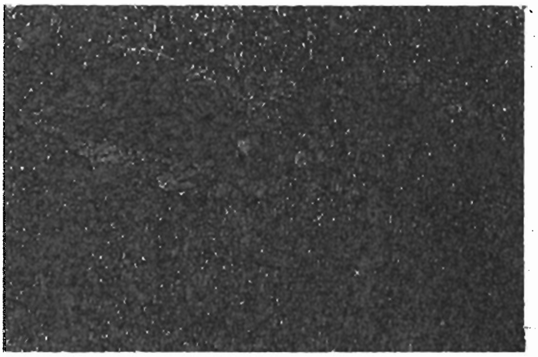


Photo. 35.

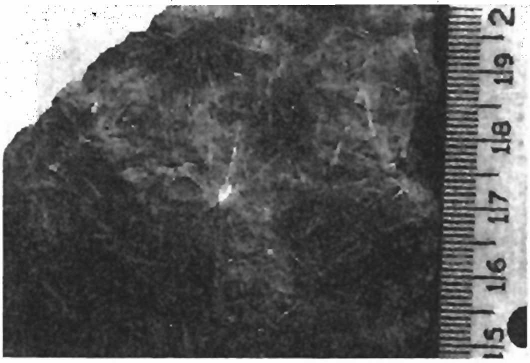


Photo 36.

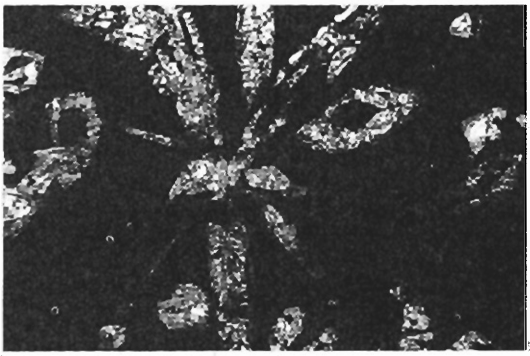


Photo. 37.

Photo. 32. Migmatitic Iskere orthogneisses, Nanga Parbat Syntaxis. (Photo. *P. J. Treloar*).

Photo. 34. Grey (right) and pink (left) granites of the Nagar Igneous Complex. Town of Nagar Parkar is in the background. (Photo. *M. Q. Jan*).

Photo. 36. Hand specimen of Khewra trap showing acicular, radiating phenocrysts. (Photo. *M. Q. Jan*).

Photo. 33. Migmatitic and mylonitic Shengus paragneisses, Nanga Parbat–Haramosh Massif. (Photo. *P. J. Treloar*).

Photo. 35. Composite mafic-felsic dyke in mafic basement, Judheg Jo Wandio, (Nagar Igneous Complex. (Photo. *M. Q. Jan*).

Photo. 37. Photomicrograph of Khewra trap showing radiating phenocrysts in an almost glassy matrix. (Photo. *M. Q. Jan*).

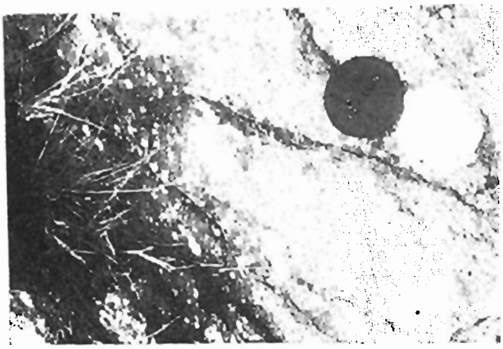


Photo. 38.

Photo. 39.

Photo. 40.

Photo. 41.

Photo. 42.

Photo. 43.

Photo.38. Undeformed Mansehra granite, Mansehra. (Photo. *P. J. Treloar*).

Photo. 40. Olistoliths of sandstone within sediments of Paleocene Rakhshani Formation in the Ras Koh Range. (Photo. *R. H. Siddiqui*).

Photo. 42. Layered gabbro in the Zhob valley ophiolites, near Khanozai. (Photo. *R. H. Siddiqui*).

Photo. 39. Intensely deformed Mansehra granite near Batgram. (Photo. *P. J. Treloar*).

Photo. 41. Folated harzburgite of the Zhob valley ophiolites, near Khanozai. (Photo. *R. H. Siddiqui*).

Photo. 43. Sheet joints in Chagai granodioritic batholith. (Photo. *R. H. Siddiqui*).

Tirah area to the east of Kurram Agency (Jan 1975). This area also contains slightly deformed leucogranites some of which have abundant sodic plagioclase and local garnet.

LATE PROTEROZOIC

Kirana Hills

Some three dozen inliers of Precambrian basement, ranging from a few hundred square metres to about 20 km^2 in area, rise above the alluvial plain of the Punjab to the SE of Sargodha. The largest of these are the Kirana Hills, 16 km from Sargodha, which rise abruptly to 330 m above the plain, itself about 180 m above sea level. The hills consist mainly of quartzites and slates, interbedded with equally abundant material of igneous origin. The rock succession attains a maximum thickness of about 825 m at Sheikh Hill; whether it is true or apparent thickness is not clear. Davies and Crawford (1971) noted that the rocks underwent tight folding that was obscured by intense shearing. Drilling at Karampur, 235 km south of Sargodha, suggests that these rocks are fairly extensive and probably form the basement to the Salt Range (Gee 1989).

Igneous component is more abundant than the sedimentary rocks in some of the hills. The Buland Hill, covering ~ 5 km^2 area, and the small hill in Chiniot town consist predominantly of rhyolitic flows, tuffs (some welded), breccia, and concordant to semi-concordant sheets (? sills) of dolerite. Basic flows, tuff, and thin dykes are rare as compared to rhyolite intrusions one of which, in Buland Hill, forms a 90 m thick lens (Davies and Crawford 1971, Alam et al. 1992).

The silicic rocks are commonly porphyritic and consist of two feldspars, quartz, and small amounts of chlorite, altered oxides and, in some, glass. Average composition of six rhyolite samples shows 0.65 wt% CaO, 0.72 wt% Na_2O and 4.07 wt% K_2O (Davies and Crawford 1971). The mafic rocks range from ophitic dolerite (bearing plagioclase of andesine-labradorite range, augite, olivine, opaque oxide) to highly uralitised or metamorphosed diabase containing abundant amphibole, epidote and chlorite (Khan and Chaudhry 1991). The tuffs and pyroclastic rocks display compositional similarity with their corresponding flows and intrusions. Some of the rhyolitic flows and tuffs are difficult to distinguish from the quartzites and slates. The abundance of the mafic over silicic rocks, as shown by Alam et al. (1992), may thus be a matter of conjecture.

The Kirana Hills represent a good example of bimodal continental magmatism. Although Heron (1913) and Alam et al. (1992) have suggested the possible presence of intermediate rocks, these are negligibly small in comparison to the rhyolitic and mafic rocks. Davies and Crawford (1971) opined that the dolerite sheets were emplaced soon after the deposition of sediments which was frequently interrupted by outpouring of silicic flows and glowing avalanches now represented by welded tuffs. However, the presence of mafic flows and tuffs suggests that mafic and silicic magmatism may have been contemporaneous. It is worth noting that many of the mafic rocks were shown as flows by Heron (1913).

The Kirana Hills form the western extension of the Precambrian shield of India. The igneous suite has a Rb-Sr isochron age of 870 ± 40 Ma (Davies and Crawford 1971), which groups it with the 850–750 Ma post-Delhi tectonic/anorogenic magmatic event (Roy 1988). Large outpourings of volcanic rocks (the Malani suite) occurred in Rajasthan towards the end of this period (~735 ± 15 Ma).

Nagar Parkar area

A basement of Late Proterozoic rocks, surrounded by sandy desert at the southeastern tip of Sindh, is a remarkable feature of the geology of Pakistan. Centred around Nagar Parkar town (24°22'N;70°43'E), the basement rocks are exposed at the edge of Thar Desert in the northeastern margin of the Rann of Cutch (Photo. 34). Wynne (1867) reported the granitic rocks of Nagar Parkar in 1867. Kazmi and Khan (1973) presented a detailed map of the area and reported a variety of Quaternary deposits, subordinate and scattered Jurassic to Tertiary sandstone and clay, overlying the basement that they termed the Nagar Igneous Complex (Fig. 6.3). From gravity anomaly data, Butt et al. (1989) concluded that much of the Nagar Parkar area is underlain by mafic rocks, whereas the granitic intrusions form thin sheet-like masses.

The igneous rocks of Nagar can be classified into: 1) metabasites (oldest), 2) acid dykes in the metabasites, 3) grey granite, 4) pink granite, and 5) mafic dykes (youngest). The metabasites are medium to- coarse-grained volcanic and plutonic rocks commonly metamorphosed to epidote amphibolites (Photo. 35). They display two distinct cleavages in the gneissosity which itself is folded. The metabasites contain deformed acidic dykes of rhyolite to quartz trachyte composition. These consist of phenocrysts of perthite, plagioclase ± quartz in an allotriomorphic matrix of these minerals and accessory Fe-Ti oxide, blue amphibole, biotite, zircon, apatite and secondary epidote. The dykes are mostly small but one body is 35 m thick and extends for >2 km. They might represent feeders to (Malani-type) volcanics now eroded away or covered by the Quaternary deposits.

The grey granite forms the maximum exposures, especially in the southern half of the area. It intrudes the metabasites and is itself intruded by the pink granite. The grey granite is mostly undeformed and medium- to coarse-grained, however, its veins in the metabasites may be pegmatitic, with abundant epidote. It is equigranular to subporphyritic and essentially composed of perthite, albite and quartz. Riebeckite and Fe-Ti oxides occur in almost all, and aegirine and minor biotite in many samples. Zircon, apatite ± allanite ± titanite are minor accessories along with local epidote. The mineral composition of the rocks is similar to the Warsak Granite in northwestern Pakistan (Kempe 1973) and indicative of their sodic alkaline nature.

The pink granite is mostly medium-grained, equigranular to subequigranular and commonly homogeneous. The largest body of this granite covers about 9 km^2. It is generally leucocratic and essentially made up of perthite (± local microcline and plagioclase) and quartz, but some contains sufficient plagioclase to be termed adamellite. Accessories include biotite, opaques, zircon, titanite, apatite, fluorite, and ?dumortierite. Locally, the rocks contain riebeckite and aegirine, thus grading modally into the grey granite. The pink colour of the rocks is due to the altered nature of the feldspar which contains iron oxide stains and specks. In several places, the grey granite has turned pink upon alteration. The distinction between the grey and pink is, thus not easy everywhere.

All the major rock units of the area are intruded by fine-grained, undeformed mafic dykes up to 3 m thick. These range from dolerite to lamprophyre and display alkaline affinity (e.g., lilac titaniferous augite). Also present are two small rhyolite domes containing feldspar and quartz phenocrysts in glassy groundmass. The relationship of these with the rest of the rocks is not clear. Preliminary geochemistry of the granites and their basic dykes suggests that they have formed in an anorogenic, within continental-plate environment (Butt et al. 1995).

For a very long distance the Aravalli Mountain Range of western India runs consistently

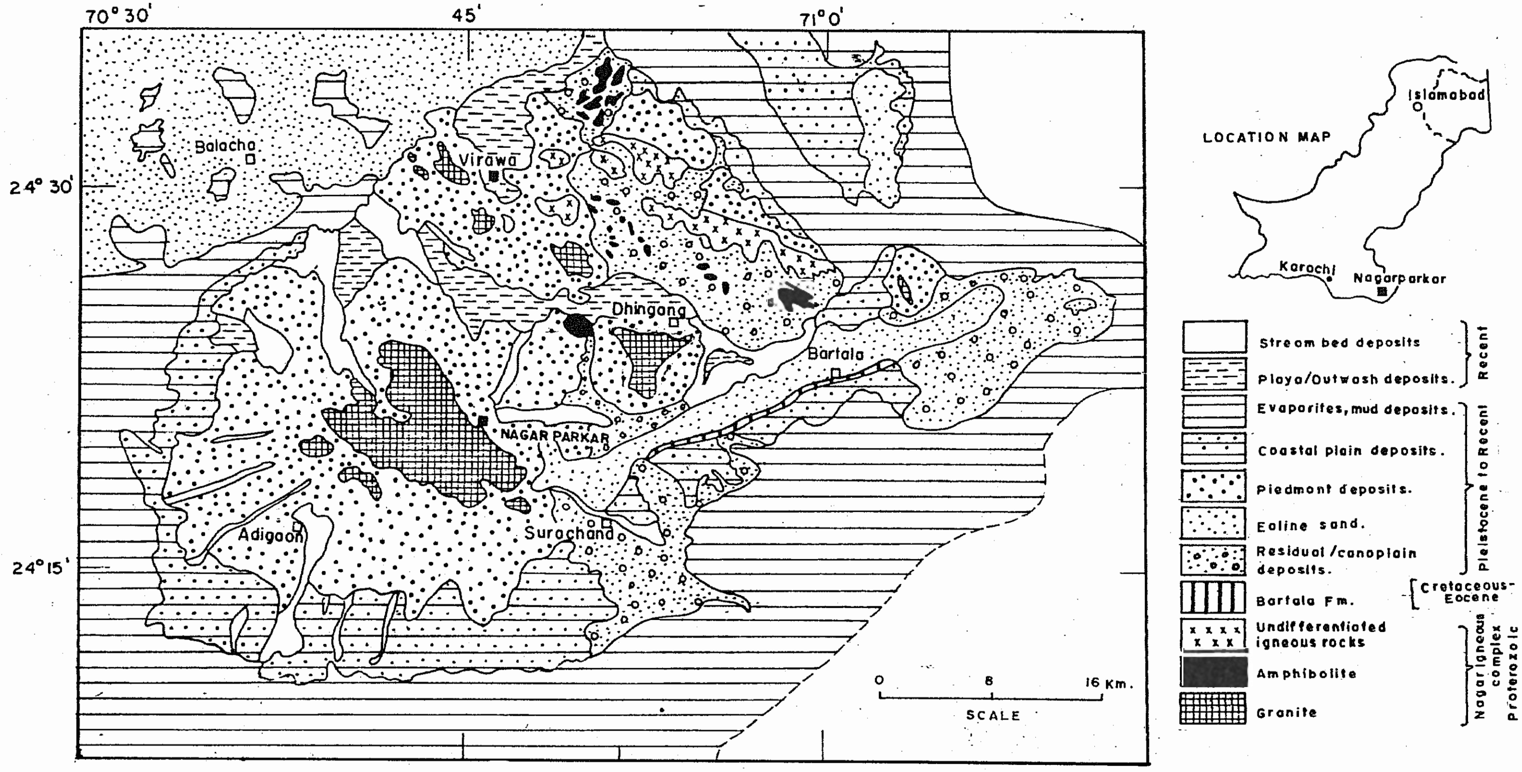


Fig. 6.3. Geological map of the Nagar Parkar area (from Kazmi and Khan 1973).

NE-SW, but towards the southern end it forms a great arc bending eastwards, possibly as a result of fault drag on a major scale along the Narmada-Son lineament (West 1962, Crawford 1978). Four major tectonic, magmatic, and metamorphic events have been recognised in the range (Roy 1988), with approximate closing ages of 3,000 Ma (Archean), 2,000 Ma (Aravalli Orogeny), 1,450 Ma (Delhi Orogeny), and 850–750 Ma (post-Delhi tectonic/anorogenic magmatic event). The range contains many granites of Late Proterozoic age. Of these, the Mount Abu and Erinpura Granites, and the Malani Igneous Suite occur on the convex side of the dragged area of the range. Since the Rb-Sr age of the Malani and Mount Abu suites is the same (i.e., 740 ± 20 Ma), Crawford (1975) regarded that the former may be the high level equivalent of the latter. He also hinted that this igneous activity could have taken place in a tectonic environment created by the sinistral shear along the Narmada-Son lineament.

The Nagar Parkar exposures occur some 200 km SW of Mount Abu and 150 km S of Malani, the intervening area being covered by sand. These granites have been encountered at shallow depth in several test holes to the southeast of Chachro, 80 km N of Nagar Parkar (Fasset et al. 1994). They seem to underlie the eastern Thar Desert. The granites display a general similarity with those of Mount Abu, Idar (Jolar) and Erinpura (Heron 1932). They also share common features with the Malani Igneous Suite (Srivastava et al. 1989), e.g., both are locally alkaline, and both contain rhyolitic and younger mafic dykes. We, thus, conclude that the Nagar Parkar granites may be a southwesterly extension of the post-Delhi magmatism towards the end of the Proterozoic. Their host amphibolites may represent a basement of an older age.

Salt Range

The classical stratigraphic sequence of the Salt Range contains thin flows of an ultrapotassic rock in its base. Commonly known as Khewra Trap, these occur in the top of the very late Proterozoic or Eocambrian rocks consisting of marly anhydrite, gypsum, and oil shales overlying evaporites. Since first reporting by Fleming (1953) the trap has attracted the attention of several geologists because of its highly unusual composition, texture and paragenesis. The trap occurs throughout the Salt Range. Drill core at Dulmial, 20 km NE of Khewra, shows two zones of mostly altered trap (total core length of 96.6 m) separated by 23.8 m of rock salt, dolomite, anhydrite and bitumen (Faruqi 1986).

The petrography of the trap has been presented by Wynne (1878), Martin (1956), Mosebach (1956), Shuaib et al. (1993) and Jan and Faruqi (1995). The trap is a purple, reddish brown, orange to buff, rarely dark green rock, frequently mottled (Photo. 36). It is very fine-grained, porphyritic and vesicular to amygdaloidal. The phenocrysts are euhedral to skeletal spinifex, and commonly grown radially (Photo. 37). They range from 1 mm to 3 cm in length and up to 1.5 mm in breadth. Irrespective of size and shape, they are completely pseudomorphed by fine-grained talc with subordinate Mg-rich clays ± quartz. Their identification, therefore, is a matter of opinion, but Mosebach (1956) and Jan and Faruqi (1995) regarded them as altered enstatite. A little olivine may have existed in at least some rocks. There also are a few microphenocrysts of titanomagnetite. The groundmass is micro- to cryptocrystalline, locally glassy, and K-feldspar (sanidine, orthoclase) in composition. It commonly contains abundant granules of hematite. Vesicles are partly to completely filled by talc, Mg-rich clays, quartz, dolomite ± chalcedony.

Preliminary geochemistry shows that the rocks are made up of 60 wt% SiO_2, 0.7% TiO_2, 11% Al_2O_3, 4–6% Fe_2O_3 (total), 0.02% MnO, 10% MgO, 0.4% CaO, 0.5% Na_2O, 9% K_2O, and 0.04% P_2O_5 (Jan and Faruqi 1995). These values classify them as ultrapotassic, despite that some silica may have been added and iron oxide lost during alteration. Rocks of similar texture, mineralogy and composition have not been reported from elsewhere in the region; indeed the trap is so unique that Mosebach (1954) assigned it a new name–Khewraite.

Several evaporite deposits of the world can be related to rifting and continental fragmentation (Windley 1984). The Salt Range evaporites and ultrapotassic rocks are closely associated in space and time, and it is tempting to relate them also to rifting. However, the major element chemistry of the rocks groups them with ultrapotassic rocks of continental areas, whereas their TiO_2 content is similar to K-rich rocks of orogenic areas. Untill the age of the Khewra Trap is precisely known and Precambrian tectonics of the region well-understood, a sound hypothesis for the plate tectonic configuration of the Khewra Trap is hard to be put forward.

CAMBRIAN–ORDOVICIAN

The Lesser Himalayan granitic plutons of Lower Ordovician and Cambrian age occur in a belt stretching for 1,600 km from Kathmandu to the Indus (Le Fort et al. 1980, 1983). Similar plutons also occur in the southern margin of the Tibetan slab and in the central mountains of Afghanistan. The Mansehra pluton with a well-defined whole rock Rb-Sr isochron age of 516 ± 16 Ma is an extension of this belt in Pakistan (Le Fort et al. 1980). Similar granitic rocks occur in other parts of the Himalayas of Pakistan, e.g., Nanga Parbat, Besham, Swat, and Azad Kashmir, all of which are confined to the northern edge of the Indian plate to the south of the Indus Suture (Fig. 6.4).

These granitic rocks are characterised by a strong gneissose fabric and generally porphyritic/porphyroblastic aspect, with feldspar megacrysts up to 15 cm long. However, the Mansehra Granite is non-foliated in the southern part, the deformation front being marked by a line passing just north of Mansehra (Shams 1969, Coward et al. 1982). The Swat Granitic Gneisses, according to Martin et al. (1962), are non-foliated at the base. Here we summarise the petrographic aspects of these plutons.

Mansehra Granites

Covering more than 2,000 km² area, these granites (Photos. 38 and 39) have been studied in detail by Shams and associates (1961a, 1966, 1969, 1980, 1983). According to Calkins et al. (1975) and Le Fort et al. (1980), they constitute a sheet tightly folded along with country rocks. They range from granite to granodiorite, are calc-alkaline in chemistry, and composed of quartz, albite-oligoclase, K-feldspar (orthoclase to microcline, mostly perthitic), biotite and small quantities of a number of other minerals, including garnet. Modal analyses of 21 samples from Batal–Battgram area of the complex (Saleemi 1978) range from potassic to normal granites with a variation towards quartz monzonite.

Shams (1969) classified the rocks into Susalgali Gneiss, Mansehra porphyritic granite, andalusite granite (commonly also containing sillimanite grown at the expense of micas), and associated minor bodies of pegmatites, aplites, albitites, and granite porphyrites. These minor bodies have been studied in detail by Ashraf (1974, 1983, 1992). A massive to weakly porphyritic tourmaline granite with albite, microcline, garnet, etc., intruding schists on its western margin, is considered by Shams (1969) to be a younger derivative.

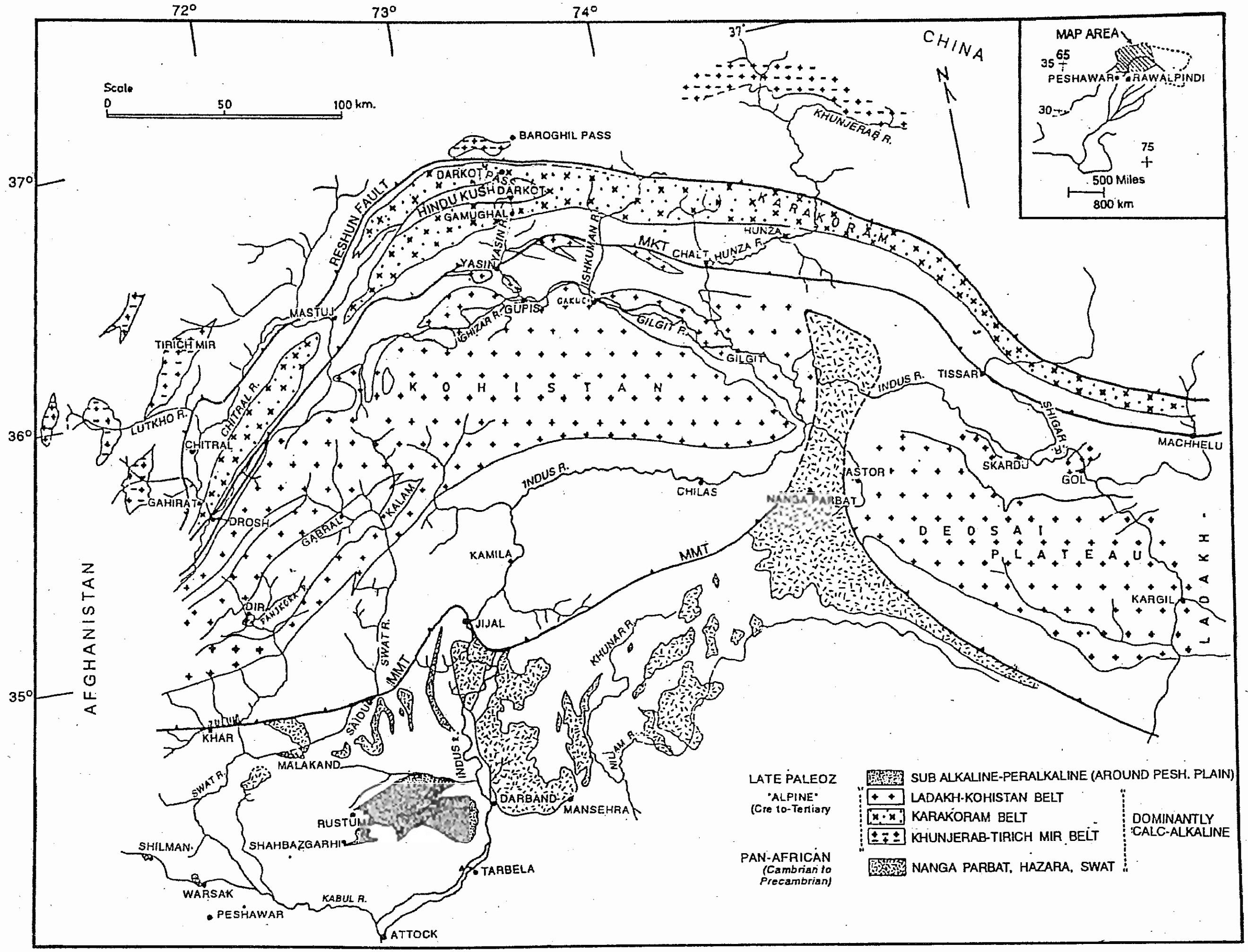


Fig. 6.4. Sketch map showing distribution of the major granitic rocks in northern Pakistan (modified from Jan et al. 1981b).

The largest outcrop of this type covers 20 km^2 area. The occurrence of cordierite (Le Fort et al. 1980), sillimanite and andalusite in the granites on one hand and of garnet on the other, coupled with staurolite and kyanite in the nearby rocks is interesting. The area has been affected by Barrovian-type, Himalayan metamorphism which increases northwards towards the MMT, as does deformation (Le Fort et al. 1980, Coward et al. 1982). It is not clear whether the granites have passed through an earlier (Late Pan-African) episode of low-pressure metamorphism or the minerals cordierite, andalusite and sillimanite are the result of contamination/contact metamorphism (Jan et al. 1981b).

Like the Nanga Parbat and Nauseri Gneisses, a metasomatic "origin of pre-existing sediments through the agency of hot permeating fluids of ultimate magmatic" derivation was also suggested for the Mansehra Granites by Shams (1969). However, in view of the current thoughts on granites, the gradually diminishing support for the hypothesis of large-scale granitisation and the information provided by Le Fort et al. (1980) for the Mansehra Pluton, we would favour a partial melting origin of the Precambrian sediments for the magma of these complexes.

Swat Granitic Gneisses

Briefly described by Martin et al. (1962) and Jan and Tahirkheli (1969), these rocks are very similar to those of Mansehra and their 515 Ma Ar-Ar biotite age (Maluski: personal communication) confirms that the two are of the same age. Like those of Mansehra, they also appear to have been emplaced in a sheet and tightly folded. The geological map by DiPietro et al. (1996) indicates that the eastern bodies of these granites pass into those of Mansehra across the Indus. They are composed of quartz, two feldspars, two micas, opaques, epidote, apatite (± garnet) and, according to Shams (1983), plot in the quartz-rich granitoid field with "two pronged extension towards granite and quartz monzonite field." A younger tourmaline-bearing granite/gneiss is a minor marginal variant of the Swat Granitic Gneisses (see also Di Pietro 1990). Wollastonite-bearing calc-silicate rocks occur in contact marbles near Manglaur (Shams 1961b) and Pir Baba.

The biotite (± tourmaline) gneisses at Chakdarra, north of Malakand, were considered to be the extension of the Swat Granitic Gneisses (Martin et al. 1962). Chaudhry et al. (1974) thought these to be syntectonic and older than the Malakand Granite. However, in a later publication (Chaudhry et al. 1976), they considered the gneisses (potash granite) to be magmatically related to the "soda granite" of Malakand proper. Jan et al. (1981b) thought that the close association of the two granites was merely accidental and not an undisputed proof of a similar age and common parentage. The gneisses show a tectonic fabric and are probably an extension of those of Swat whilst the Malakand Granite is undeformed and possibly belongs to the alkaline province discussed in a following section. Modally also, the Chakdarra Gneisses resemble those of Swat. Ahmad and Lawrence (1992) have equated these with the tourmaline-bearing facies of the gneisses found further east.

Humayun (1985) divided the granitic rocks in Lower Swat into (1) porphyritic calc-alkaline granodiorite gneisses with white microline megacrysts, (2) equigranular to porphyritic biotite granites of probable alkaline affinity and displaying local rapakivi texture, and (3) equigranular, subsolvus, tourmaline-muscovite granites. The rapakivi granites were considered anomalous and emplaced tectonically from the basement.

Granitic gneisses, similar to those of Swat, have been reported to the west of Malakand from Sillai Patti (Hussain et al. 1984) and Shamozai–Utmankhel tribal territory (Badshah

1979, M. Rafiq pers. comm.). The Chingalai Granodiorite Gneiss (K-feldspar phenocrysts + oligoclase + quartz + "abundant" biotite ± hornblende) was considered by Siddiqui et al. (1968) to be genetically related to the alkaline rocks of Koga area in Buner. Recent investigation of this area casts doubt on the relation since a major fault separates the two types of rocks. We tentatively place this granodiorite in the Cambro-Ordovician.

Azad Kashmir

Several, relatively small granitic bodies have been reported from Azad Kashmir. They are porphyritic, medium- to coarse-grained, and can be grouped into deformed (gneissose) and undeformed types commonly containing biotite and muscovite. The Nauseri, Jura and Kel Granitic Gneisses appear to be potassic, with K-feldspar phenocrysts up to 13 cm in length (Shakoor 1976, Ghazanfar et al. 1983). The Reshian Augen Gneiss contains plagioclase porphyroblasts, along with microperthite, biotite, garnet, rutile, epidote, sphene, tourmaline and pyrite, and is cut by tourmaline-quartz veins (Greco 1986, Rehman and Chaudhry 1981). The Neelum Porphyritic Granite intrudes the Jura granite and is itself intruded by the 1.5 km broad Dainyar adamellite(?) containing up to 17% tourmaline (Ghazanfar et al. 1983).

The granitic rocks of the Nauseri area were studied in some detail by Khan et al. (1994). These constitute an over 40 km long and up to 5 km broad belt that extends towards Kaghan. Most of the rocks are granite in composition and contain muscovite, biotite, apatite ± sphene ± zircon; hornblende occurs in the Thora body. The rocks display shearing, mylonitisation and foliation, probably related to Himalayan deformation. The main Nauseri body consists of coarse-grained augen gneisses with finer grained margins. Petrography and major element analyses suggest that the rocks are peraluminous and a product of partial melting of the associated sediments.

Radiometric dates are not available on the rocks and their equivalence with the Mansehra Granites is a matter of opinion. The stratigraphic studies of Greco (1986) show that the Reshian Granitic Gneiss is older than the Late Paleozoic Panjal Volcanics. A similar conclusion can be drawn about the Neelum Granite which is traversed by abundant dolerites probably related to the Panjal magmatic episode. On purely stratigraphic grounds, Shakoor (1976) placed the Nauseri Granitic Gneisses in the Late Cambrian. Greco et al. (1989) have reported the occurrence of such rocks in the Kaghan area also, where they contain metamorphic kyanite and garnet. Keeping in view the likely occurrence of Precambrian granites in the adjacent Kaghan area, it would not be surprising if some of the Azad Kashmir granites also turn out to be pre-Paleozoic.

Nanga Parbat and Besham

Reference to granitic rocks of this age in the two areas was made earlier. The two-mica Shingus Gneiss in the Nanga Parbat Massif is a fine-grained, finely-laminated unit comprising a range of lithologies, but petrographic details are not available to us. It has a U-Pb zircon age of 400–500 Ma, and is considered as a metamorphic equivalent of the Mansehra Granite (Zeitler et al. 1989). In Besham area, there are small granitic intrusions of Cambrian age in the Early Proterozoic basement. Of these the Shang and Duber bodies with associated pegmatites are worth mentioning. These are coarse-grained, with up to a centimetre long feldspar phenocrysts, deformed (especially on margins), and locally banded. They range in composition from granodiorite to adamellite and consist of K-feldspar, quartz, medium plagioclase, hornblende and/or biotite, opaque oxide, and traces of sphene, epidote, tourmaline, zircon and local garnet. The granite in the north of Duber has an Ar-Ar biotite age of 550 ± 20 Ma (Treloar et al. 1989c).

Western Karakoram

In upper Yarkhun Valley of northeastern Chitral, Le Fort et al. (1994) and Tongiorgi et al. (1994) reported pre-Ordovician granitoids occurring to the north of the Karakoram Axial Batholith. The principal pluton at Ishkarwaz is an altered and deformed biotite-hornblende adamellite, forming an alumino-cafemic association with calc-alkaline affinity. The granite body is 4 to 5 km wide. It intrudes low-grade quartzite and migmatite, and is covered transgressively by fossiliferous litharenite and slate of Early Ordovician age. Other smaller bodies, probably forming apophyses 1 to 4 km wide and intruding the same formation of metaquartzite and mica schist, may be coeval. Early Paleozoic granitoids have also been reported from the neighbouring region.

The Ishkarwaz Granite has low Mg/(Mg + Fe) ratio, high Ba (~1,000 ppm), and resembles closely the darkest member of the Bumburet Pluton of Kafiristan in western Chitral (Le Fort et al. 1994). The latter is an augen-gneiss of granodioritic composition containing mica ± amphibole (Khan 1986). This area also contains a variety of pegmatites but their relation to the Early Paleozoic granitoids is not clear. The occurrence of Early Paleozoic granites in the Karakoram Range suggests that the Karakoram plate is Gondwanic in origin and was probably contiguous to India before Permian.

Origin of Cambro-Ordovician Granites

According to Le Fort et al. (1980, 1983), it is difficult to relate the generation of these granites either to a definite orogeny (Late Pan-African, for example) or to another phenomenon–like an extended thinning of the crust or a strike slip movement giving rise to a sort of thermal corridor. Some deformation, according to them, is still recognisable but to a limited extent and no indication of a previous metamorphism has yet been observed. One aspect worth a careful study is the occurrence of andalusite, sillimanite, and cordierite in some of the granites. Are these low pressure minerals due to contamination and contact metamorphism, or have they been produced by a regional metamorphic episode during Late Pan-African Orogeny? Shams (1969) observation regarding the growth of sillimanite at the expense of micas hints at the second possibility. On the other hand, there is ample evidence to suggest that the granites and the surrounding rocks have been affected by alpine metamorphism (Barrovian-type) and deformation.

The composition, high initial Sr^{87}/Sr^{86} ratio, and evolutionary trend for the Mansehra pluton suggest that the rocks resemble closely the S-type granite series derived from an old crustal basement (Le Fort et al. 1980). The petrography, stratigraphic details and, in a number of cases, "gradational" contacts between the granites and country rocks, as presented by some workers, suggest to us that the Cambro-Ordovician granites were derived by the partial melting of Precambrian sediments. It is not totally surprising, thus, that some workers proposed a metasomatic origin for them. The Swat, Besham, Mansehra, and Kashmir gneisses were intruded as sheets and folded tightly along with country rocks. The whole association, according to Le Fort et al. (1980), was overthrust along the Main Central Thrust.

LATE PALEOZOIC – EARLY MESOZOIC

Based on the occurrence of alkaline rocks in Warsak, Shewa–Shahbazgarhi (Coulson

1936), Koga (Siddiqui et al. 1968) and Tarbela, Kempe and Jan (1970) suggested that an alkaline igneous province stretched across north Pakistan. Subsequent geochemical and petrographic data on granitic rocks of Ambela (Ahmad and Ahmed 1974, Rafiq 1987) and Malakand (Chaudhry et al. 1974), and the discovery of carbonatite complexes in Shilman (Jan et al. 1981) and Sillai Patti (Ashraf and Chaudhry 1977) provided further support to the idea that the Peshawar Plain alkaline igneous province (PAIP) extends for a distance of at least 150 km between the Indus River and Pak-Afghan border (Fig. 6.5). Field studies summarised in Kempe and Jan (1980) indicate that the alkaline complexes are (1) generally emplaced along fault zones, and (2) restricted in occurrence to Paleozoic and Precambrian rocks. To date none has been reported in Mesozoic and Tertiary rocks.

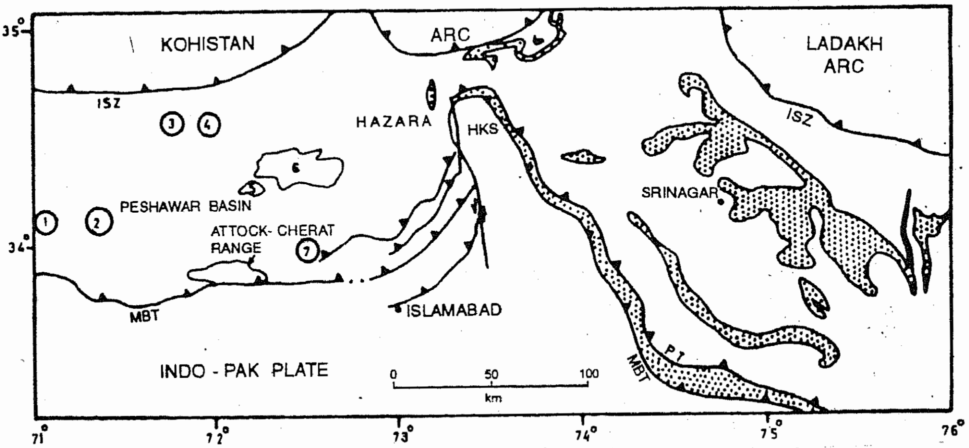


Fig. 6.5. Sketch map showing the location of the Peshawar Plain alkaline igneous province (PAIP). Numbers refer to locations of the outcrops of the PAIP discussed in text. Panjal Volcanics shown by dots. ISZ–Indus Suture zone, PT–Panjal Thrust, HKS–Hazara Kashmir Syntaxis (from Jan and Karim 1990).

Kempe (1973) and Kempe and Jan (1980) suggested that the rocks of the alkaline igneous province are associated with Tertiary rifting. Their conclusions, unfortunately, were based on only two K-Ar dates from Koga (50 Ma) and Warsak (41 Ma). Although additional Tertiary ages have also been reported for the Malakand Granite (20–23 Ma fission track and Ar-Ar), Warsak Granite (40–43.5 Ma Ar-Ar), Loe Shilman and Silli Patti carbonatites (31 ± 2 Ma K-Ar), and Ambela Syenite (47.5 ± 1.5 Ma Ar-Ar) (Zeitler et al. 1982, Maluski and Matte 1984, Le Bas et al. 1987), it is likely that these dates represent tectonometamorphic events rather than magmatism (see Table 6.1). The few reliable ages (Le Bas et al. 1987: Ambela, Zeitler 1988: Malakand, Khan et al. 1990: Shewa-Shabazgarhi) and the total absence of these rocks in the post-Paleozoic sedimentary sequence favour a Late Paleozoic age for the PAIP (Table 6.1).

The later part of the Paleozoic and beginning of the Mesozoic appears to have been a fertile time for magma generation in this part of the subcontinent. In addition to the PAIP, numerous dolerite dykes were emplaced in the metasediments presently occurring between the Indus Suture and the Khairabad Thrust and there was volcanic activity of the Panjal cycle in Kashmir and adjoining areas. This rather extensive continental magmatism, spanning Carboniferous to Triassic period and ranging from granite to carbonatite on the one hand and mafic to rhyolitic on the other, was regarded by Jan and Karim (1990) to be connected to a major episode of continental swelling and rifting.

Here we present an intergrated summary of the three groups of rocks to elucidate

Table 6.1. Radiomeric ages from the Alkaline Igneous Province. (Modified from Jan and Karim 1990).

LOCALITY	ROCK	SYSTEM	AGE (Ma)	REFERENCE
1) Loe Shilman	Carbonatite	K-Ar (Bio)	31 ± 2	Le Bas et al. (1987)
2) Warsak	a) Granite	K-Ar (Amp)	41	Kempe (1973)
	b)	Ar-Ar (Amp)	43.5 ± 5	
	c)	Ar-Ar (Amp) (Bio)	40 ± 5 42 ± 4	Maluski & Matte (1984)
	d) Hbl schist	K-Ar	184 ± 17	Kempe (1986)
3) Malakand	a) Granite	F.T (Ap-Zir)	20	Zeitler (1982)
	b)	Ar-Ar. (Mus)	22.8 ± 2.2	Maluski & Matte (1984)
	c)	U-Pb (Zir)	271 ± 11	Smith et al. (1994)
4. Silai Patti	a) Carbonatite	K-Ar (Bio)	31 ± 2	Le Bas et al. (1987)
	b)	F.T (Zir) (Ap)	32.3 ± 1.4 21.8 ± 0.4	Qureshi et al. (1990)
5) Ambela	a) Syenite	K-Ar (W. Rock)	50	Kempe (1973)
	b)	Ar-Ar (Bio)	47.5 ± 1.5	Maluski & Matte (1984)
	c) Syenite & ijolite	Rb-Sr (W. Rock)	315 ± 15 297 ± 4	Le Bas et al. (1987)
	d) Syenite	U-Pb (Zir)	280 ± 15	Smith et al. (1994)
6) Tarbela	Albitite	K-Ar (Hbl)	350 ± 15	Kempe (1986)

their petrography, emplacement ages and tectonic setting based mostly on published data.

The Peshawar Plain Alkaline Igneous Province (PAIP)

Of the three groups of rocks discussed in this paper, the PAIP is the most diversified in petrography. Kempe and Jan (1980) and Kempe (1983) considered ten occurrences that comprise the province. In the following we describe seven of these, from W to E (Fig. 6.5), ignoring those of eastern Afghanistan and Mohmand for want of authentic data and the Mansehra albitites for their association with the Early Paleozoic granitoids. We have included the Shilman and Sillai Patti carbonatites with the PAIP although they may well be a separate entity of mid-Tertiary age.

1. Loe Shilman, Khyber Agency: Sill-form bodies of carbonatite are emplaced along an E-W-trending and N-dipping fault zone in pre-Mesozoic rocks of Khyber Agency. The main intrusion, reaching 170 m in width, extends for 2.5 km, passing westward into Afghanistan. Isolated sheets of carbonatite occur further east. The carbonatite comprises amphibole sovite which is intruded by biotite sovite and amphibole ankeritic carbonatite. There is a zone of fenitisation up to 100 m broad (Mian and Le Bas 1987). Syenites, lamprophyric rocks and Fe-rich hydrothermal veins are associated with the carbonatites (Jan et al. 1981d), and Butt (1990) has described K-rich rocks from the complex.

Gabbro and dolerite sills, generally not more than a few metres thick, occur to the east and north of the carbonatite complex. These are amphibolitised and resemble those of the Warsak area, 35 km to the east. The relationship between the gabbros and carbonatite is not known, but Jan et al. (1981d) speculated that the two may be genetically related. A K-Ar biotite date of 31 ± 2 Ma was considered by Le Bas et al. (1987) to be the age of formation of the carbonatite.

*2. **Warsak:*** A series of sill-form alkaline granites, microgranites, gabbros and dolerites occurs in a 5 x 8 km N-plunging syncline at Warsak (Coulson 1936, Ahmad et al. 1969, Kempe 1973, 1983). The granitic rocks, some of which are foliated and sheared, contain aegirine, riebeckite, astrophyllite and/or biotite ± garnet. The entire area has been metamorphosed in upper greenschist facies, and mafic rocks contain abundant amphibole and relics of pyroxene. There are mafic tuffs, agglomerates and pillow lavas, raising the possibility that some of the acidic rocks may also be volcanic (Kempe 1978). Small bodies of similar gabbroic and granitic rocks occur not only in Shilman to the west, but also 10 km to the south (Khan et al. 1970) and perhaps to the north.

The age relationship of the granitic and gabbroic rocks is not clear but the volcanic members seem to have preceded the intrusive phase. The Warsak area is, apparently, a good example of contemporaneous bimodal magmatism. On the basis of limited data, Kempe (1978) suggested that the whole may constitute a differentiated series. Recent geochemical investigation (Khan 1991, Tahirkheli et al. 1990) shows that the granites are A-type and characteristic of continental rift magmatism. A K-Ar amphibole date of 41 Ma was considered by Kempe (1973, 1983) as age of emplacement. This age, however, is at odds with the 187 ± 17 Ma K-Ar age (hornblende) on a metamorphosed mafic tuff (Kempe 1986). Maluski and Matte (1984) thought that their 40 to 43 Ma Ar-Ar ages on amphiboles and biotite from Warsak were indicative of tectono-metamorphic activity. The Warsak rocks are strikingly similar in petrography and geochemistry to those of Shewa–Shahbazgarhi (Coulson 1936, Kempe 1973, 1983), which have a Carboniferous stratigraphic age (Khan et al. 1990).

*3. **Sillai Patti:*** Ashraf and Chaudhry (1977) reported a carbonatite occurrence about 20 km west of Malakand. Le Bas et al. (1987) suggested that the carbonatite was emplaced along a thrust in the form of a 12 km long and 20 m thick sheet. Butt (1989) presented geochemical data for the carbonatite and noted that it is made up of dyke-like bodies emplaced either within the metasediments or at their contact with granitic rocks.

The carbonatite, consisting of biotite-apatite sovite, amphibole-apatite sovite and alkaline pyroxenites, has fenitised the country rocks, especially the granitic gneisses. The latter appear to be an extension of the Swat Granitic Gneisses (Jan et al. 1981b). K-Ar dating on an undeformed biotite in the carbonatite has yielded 31 ± 2 Ma (Le Bas et al. 1987). An identical fission-track zircon age (32.3 ± 1.4 Ma) has been determined by Qureshi et al. (1990). These authors have argued that the carbonatite, like that of Shilman, formed during Oligocene. Butt (1989), however, thinks that this may be the age of metamorphic overprinting and that the carbonatite may be Late Paleozoic. This possibility was preferred by Jan and Karim (1990).

Small carbonatite bodies (with biotite, pyroxene, amphibole) and associated fenites also occur in Jambil area of Swat. It is likely that further search would reveal additional bodies of carbonatite in the region.

4. ***Malakand:*** In the Malakand Pass area, granitic rocks intrude an ESE-plunging anticline made up of gneisses and schists (Shams 1983). Khan (1965), Chaudhry et al. (1974, 1976) and Hamidullah et al. (1986) have presented petrographic details and geochemistry of the rocks. The granite contains albite, microcline, quartz, muscovite, epidote, biotite, calcite, garnet, etc. Allanite and sphene are common in samples from Benton Hydroelectric Tunnel (Kempe and Jan 1980). Pegmatites (some carrying tourmaline and fluorite) and aplites are commonly associated with the granites. The gneisses have more or less similar modes and have been chemically divided into siliceous, silica-rich and normal granitic types by Hamidullah et al. (1986). These authors and Chaudhry et al. (1976) suggested that the granites were related to the granitic gneisses.

The Malakand Granite has a 20 Ma fission-track apatite and zircon age (Zeitler 1982) and an Ar-Ar muscovite age of 22.8 ± 2.2 Ma (Maluski and Matte 1984). In the light of chemical data Chaudhry et al. (1976) thought it to be "basically a soda granite" while Kempe and Jan (1980) considered that the Malakand Granite may belong to the PAIP. Hamidullah et al. (1986) report that the granite is peraluminous and calc-alkaline, and not related genetically to the PAIP. Recent work on U-Pb zircon systematics shows that the Malakand Granite is Permo-Carboniferous in age (Zeitler 1988). We regard that (1) the Malakand Granitic Gneisses are an extension of the Swat Granitic Gneisses of Early Paleozoic age, and (2) the Malakand Granite is a product of the magmatic episodes of Late Paleozoic and probably related to the PAIP.

5. ***Shewa–Shahbazgarhi:*** Acidic porphyrites/micrograntes with basic intrusions cover a 35 km^2 triangular area near Shahbazgarhi. Isolated outcrops of such rocks occur 15 km to the south (Martin et al. 1962) and further north near Rustam (Rafiq 1987), suggesting that these sheared volcanic/subvolcanic rocks may once have covered a much larger area. On the basis of petrography and geochemistry, these rocks have been considered to be consanguineous with those of Warsak (Coulson 1936, Kempe 1973, Kempe and Jan 1970). The acidic rocks consist of an earlier garnetiferous group containing biotite and lacking alkaline ferromagnesian minerals, and later alkaline microgranites with aegirine, riebeckite and biotite (Kempe 1983). Different petrographic and geochemical divisions have been suggested by Chaudhry and Shams (1983) and Ahmad et al. (1990). The mafic intrusions, like those of Warsak, are epidote amphibolite facies (?) metamorphosed gabbros and dolerites containing hornblende (hastingsite) and epidote.

Chaudhry and Shams (1983) presented major element geochemistry of the acidic rocks and concluded that they are the product of anatectic melts of deep crustal origin, emplaced during alternate periods of tension and compression related to subduction of the Indian plate during Late Cretaceous-Early Tertiary. Ahmad et al. (1990) noted that the basic rocks share the characteristics of continental flood basalts and were (along with the alkaline and peralkaline acidic rocks) emplaced during continental rifting.

According to Khan et al. (1990) and Pogue et al. (1992), the Shewa porphyritic alkaline acidic rocks are intercalated with Carboniferous sediments. This discovery is important because, (1) the Shewa–Shahbazgarhi porphyrites may at least partly be volcanic (? tuffs and flows), (2) this is the first stratigraphic age on any rock in the PAIP. This age is in conformity with those determined radiometrically for Malakand, Tarbela, and parts of Ambela, and (3) the close similarity of these rocks with those of Warsak leads to speculate that the latter may also be Carboniferous.

6. ***Ambela Granitic Complex (AGC):*** Covering about 900 km^2, the AGC is by far the largest body of the PAIP. Following the initial brief description (Martin et al. 1962), Siddiqui et al. (1968) presented the petrology of the alkaline rocks from the western part of the AGC in Koga area. Since then much more work has been done on this complex, of which that by Chaudhry et al. (1981), Rafiq (1987) and Mian (1987) is particularly important. Rafiq and Jan (1988, 1989) presented details of petrography and geochemistry, and classified the complex into three major groups of rocks. Group I, the product of the first magmatic episode, consists of granites and alkali granites which occupy about 70 % of the batholith. Group II, following the granites sequentially, comprises quartz syenites, feldspathoidal syenites, ijolite and carbonatite. Considerable metasomatism accompanied the successive phases of intrusions in this group. Finally, the complex was invaded by dolerite and lamprophyre dykes which occupy 5% of the area and constitute the group III. As elsewhere, the dolerites display alteration/metamorphism and contain substantial amounts of hornblende and epidote. These have continental tholeiitic affinity (Rafiq and Jan 1990).

The granitic rocks range from dominantly peraluminous through metaluminous to mildly alkaline, and are derived from magmas related to crustal thinning and rifting. Different degrees of partial melting and fractional crystallisation led to variation in the composition of these rocks. With deepening of the zone of magma generation in the crust, the underlying mantle was activated, resulting in influx of volatiles and alkalies. This led to the generation of magma batches which were successively more SiO_2-undersaturated and alkaline, resulting in production of Group II rocks (Rafiq and Jan 1989).

Based on a 50 Ma K-Ar syenite date, Kempe (1973) proposed that the complex formed during Eocene. This idea was further substantiated by a 47.5 ± 1.5 Ma Ar-Ar biotite age from a syenite (Maluski and Matte 1984). Kempe and Jan (1980) equated the Ambela rocks with those of Warsak, Shewa and Tarbela, and placed them in the PAIP (see Ahmad and Ahmed 1974). Le Bas et al. (1987) reported that the alkaline rocks of Koga have Rb-Sr isochron ages of 297 ± 4 to 315 ± 15 Ma. Smith et al. (1994) determined a 280 ± 15 Ma U-Pb zircon age for the syenite. The AGC is mainly intrusive into the sedimentary rocks of Siluro-Devonian age. However, late pegmatites and veins occur in Early Triassic rocks (S.R. Khan personal communication). The batholith contains xenoliths of (? cover) volcanics similar to those of Shewa–Shahbazgarhi (Rafiq 1987). On these grounds it can be concluded that the AGC was emplaced mainly during Carboniferous between 300 and 350 m.y. ago.

7. ***Tarbela:*** The Tarbela "alkaline" complex comprises gabbroic rocks (oldest), dolerites, a variety of albitites, granites (some of which are sodic), albite-carbonate rock/breccia, and (?) carbonatites (youngest) (Jan et al. 1981a). The rocks stretch for at least 4 km, but many outcrops have been removed or covered during the construction of the Tarbela dam. The complex may have been intruded along a fault zone. Some of the gabbroic intrusions display in situ differentiation, with one intrusion grading from pyroxenitic outer margin to leucogabbroic/dioritic interior, with a core of "intrusive" albitites. Considerable metasomatic activity accompanied the rocks with the development of scapolite, albite, carbonate, quartz and pyrite. Some albitised country rocks resemble adinoles and it has been suggested that most albitites may be metasomatic (Le Bas, personal communication, 1985).

Amongst typical alkaline minerals, sodic amphiboles and pyroxenes are restricted to sodic granites, now removed. Trace elements in albite-carbonate rocks and the high quantity of albite and carbonate along with consistent presence of zircon, rutile and/or sphene in most albitites are suggestive of their alkaline character. The alkaline affinity of the gabbroic rocks is indicated by the abundance of amphibole (hornblende, kaersutite, hastingsite), low anorthite content of plagioclase, clinopyroxene composition, the general absence of primary quartz, and the possibly high Ti content reflected in amphibole, sphene and ilmenite. Preliminary geochemistry supports this view. Kempe (1986) has reported a K-Ar date on hornblende from an albitite as 350 ± 15 Ma, which is close to the formation age of the nearby Ambela Granitic Complex, provided the date does not reflect excess argon.

Before closing this section, it is worth mentioning that Spring et al. (1993) have recently described alkaline granites ($K_2O > Na_2O$) in upper Lahul and SE Zanskar. With a U-Pb zircon age of 284 ± 1 Ma, these are coeval with those of PAIP and have been related to rifting.

The Panjal Volcanics

Having a wide distribution and covering about 12,000 km^2 area, the Panjal Volcanics are a prominent feature of NW Himalayas. They occur in an up to 40 km broad belt of intermittent outcrops extending NW from Zanskar through Pir Panjal Range, Neelum and Kaghan Valleys, turning around the western Himalayan Syntaxis to Balakot. The volcanics are a succession (up to 2,500 m thick) of thin (mostly a few cm to 3 m) basaltic flows with interbedded tuffs and limestone, underlain by agglomeratic slates, grits, pyroclastics, limestone/marble, graphitic schists, pelitic schists and metaconglomerate. Wadia (1931) also reported dykes and laccoliths of mafic composition in the volcanics. Included here are the volcanic rocks of Suru and those of Zanskar which consist of Late Carboniferous-Early Permian basaltic flows (Srikantia et al. 1978). It is not clear but some volcanic members of the PAIP may be the equivalent of the Panjal Volcanics.

The Panjal Volcanics have been stratigraphically assigned to Late Carboniferous to Triassic, but much of the volcanic activity took place during the Permian (Pareek 1982, Gupta et al. 1982). Ghazanfar et al. (1987) and Chaudhry et al. (1987) divided the rocks into Panjal Formation (volcanics) of Permian age and Chushal Formation (agglomeratic slate, etc.) of Late Carboniferous age. The volcanics are dominantly basaltic in composition, however, andesite, trachyte, and subordinate nepheline basalt, ankeramite and limburgite also occur (Honegger et al. 1982). The basaltic members are mostly non-porphyritic, but some have pyroxene and/or plagioclase phenocrysts in a fine-grained or glassy matrix. In Azad Kashmir and Kaghan the top of the succession contains a high proportion of tuffs and ashes (mostly intermediate to acidic), and the flows locally appear to have pillow structures (Chaudhry et al. 1986).

The volcanics have undergone varied degrees of alteration, metamorphism and deformation. Tight folding is common throughout and the rocks may display shearing, schistosity, and flow banding. They are greenish and amygdaloidal, but locally "mottled" due to black and dark green amygdules in red and white mineral aggregates (Greco 1986, Sinha 1981). Epidote, calcite, chalcedony and jasper are widespread alteration products (Pascoe 1949). Metamorphic grade has been reported to increase from south with zeolite and greenschist facies (Papritz and Rey 1989) to greenschist (Chaudhry et al. 1986) or even amphibolite facies in the north (Greco et al. 1989). Spencer et al. (1990) report eclogites (first record in Himalayas) derived from the Panjal Volcanics in the upper part of the Kaghan Valley. The Swat amphibolites are also derived from volcanic flows of this age (Pogue et al. 1992).

The Peshawar-Hazara-Kashmir dolerites

Between the Indus Suture and the Main Boundary Thrust (MBT), there are numerous dolerite bodies in the Precambrian to Paleozoic rocks of NWFP and Kashmir. Notable occurrences are those of theAttock–Cherat Range (Wadia 1931, Tahirkheli 1970), Khyber Agency (Khan et al., 1970, Shah et al. 1980), Hazara (Shams and Ahmed 1968, Calkins et al. 1975), and Kashmir (Wadia 1961, Pascoe 1949). The association of such rocks with granites in the PAIP has already been described.

The dolerites commonly occur as sills and dykes, but locally as plugs or irregular bodies. They rarely exceed ten metres in thickness and a few hundred metres in length. At places the dykes occur in swarms. Available data on the orientation of the dykes commonly show E-W trends in Khyber Agency (Shah et al. 1980), Attock–Cherat (Karim and Sufyan 1986) and Hazara (Calkins et al. 1975). The dykes were probably subjected to late thrusting events during Himalayan Orogeny, which may have modified their regional trends, however, their parallelism is generally preserved.

Only limited mineralogical data have been presented for these rocks in Attock–Cherat Ranges (Karim and Sufyan 1986), Mansehra (Ahmed 1985b) and northeast of Tarbela (Majid et al. 1991). Common mineral constituents are mostly plagioclase (An_{45-71}), augite, with some Fe-Ti oxide, biotite, pigeonite, hornblende, epidote, and chlorite. A few rocks contain olivine (Fo_{38-52}). Many can be classified as dolerites, but some are lamprophyres, norites and gabbros. Some of the latter in Kashmir have acted as feeder dykes, sills and bosses to the Panjal Volcanics (Pascoe 1949). The rocks have experienced low-grade metamorphism and weathering: chlorite, epidote, secondary amphibolite and sodic plagioclase are common. Some garnet amphibolites in the basement rocks of Kaghan Valley have been considered the subsurface equivalent of the Panjal Volcanics (Papritz and Rey 1989).

The dolerites intrude Late Paleozoic rocks (upper Permian in Khyber Agency: Shah et al, 1980, and Carboniferous in Attock–Cherat Range); but they have not been reported from Mesozoic and Tertiary rocks, including the Triassic of Kalachitta (A. Hussain personal communication). Therefore, it can be concluded that many of the dolerites may be of Permian to Early Triassic age, but some may be older and related to the PAIP and early Panjal cycle (for further details, see Pogue et al. 1992). Mafic dykes intruding the Mansehra Granite yielded $^{39}Ar/^{40}Ar$ plateau dates of 284 ± 4 Ma and 262 ± 1 Ma (Baig 1990).

Geochemistry

Geochemical data for the three groups of rocks have important implications regarding their paleotectonic setting. The precise genetic relationship and the processes involved in their evolution cannot be ascertained on the basis of existing geochemical data. The carbonatitic complexes are chemically distinct, by virtue of their trace elements and high REE, from the rest of the rocks of the PAIP. These have been regarded as the product of alkaline magmatism rather than remobilised limestones (Le Bas et al. 1987).

The silicic rocks of the PAIP are predominantly granitic in composition, but some are adamellite, granodiorite and monzonite, and range from peraluminous to alkaline and peralkaline in character (Rafiq and Jan 1989). Figure 6.6 shows the Nb vs SiO_2, Rb vs SiO_2, Y vs SiO_2, Nb vs Y and Rb vs Y + Nb plots. These unequivocally classify the rocks to have originated in within-plate environments. Furthermore, the PAIP granitic rocks follow the trend of type extensional tectonic suites (Jan and Karim 1990). The gabbroic rocks of the PAIP dominantly plot in the field of tholeiites on the MFA diagram (Fig. 6.7). Additional geochemical data of petrogenetic significance are partial trace element analyses on these rocks. On multi-element spidergrams, their patterns closely resemble continental tholeiites (Jan and Karim 1990, Tahirkheli et al. 1993).

The Panjal Volcanics show a variable degree of alteration, rendering major elements chemistry unreliable for inferring magmatic affinities (Humayun et al. 1987). The volcanics are apparently tholeiitic to mildly alkaline in character (Gupta et al. 1982). Trace and REE data show that they are similar to plateau basalts (enriched P-MORB type magma) originated in within-plate environments (Honegger et al. 1982, Papritz and Rey 1989, Vannay and Spring 1993). Figure 6.8 shows the mantle normalised trace elements spidergrams for the Panjal basalts. Their patterns compare well with continental tholeiites. Khan et al. (1992) have, however, suggested a transitional within-plate to ocean floor setting for the Kaghan Panjal basalts on the basis of incompatible trace elements (see Greco 1991, p.48). The Swat amphibolites also display continental flood basalt affinity (Humayun 1986).

The Peshawar, Hazara and Kashmir dolerites are mostly-tholeiitic basalts with some alkaline compositions (Majid et al. 1991) and have continental affinities. Mantle normalised trace elements spidergrams for dolerites from Attock–Cherat Range show enrichment in large ion lithophile (LIL) elements relative to high field strength (HFS) elements and Nb trough, and their overall patterns closely resemble average continental tholeiites (Fig. 6.9). Clinopyroxenes from Attock–Cherat Range and Mansehra and Tarbela gabbros occupy the field of non-orogenic suites on Ti + Cr vs Ca discrimination diagram (Fig. 6.10). Late Paleozoic dolerites in Lahul–Spiti and Zanskar also have broadly similar characteristics, but have been considered unrelated to the Panjal Volcanics (Vannay and Spring 1993).

Salt Range

A vitric lava flow or tuff, only 10 to 30 cm thick but having a strike length of 4 km, has been recently reported by Butt et al. (1994). It occurs within the Early Permian Warcha sandstone in the eastern part of the Salt Range near Khewra. It is made up of brown glass (78%), iron oxide/hydroxide (10%), microcrystalline calcite (5%), and xenocrystic feldspar (3%) and quartz (2%). Four chemical analyses show average amounts of 15.6% SiO_2, 0.3% TiO_2, 6.5% Al_2O_3, 6.6% Fe_2O_3, 0.3% FeO, 0.4% MnO, 0.5% MgO, 42.9% CaO, 1.0% Na_2O, 0.7% K_2O, 0.4% P_2O_5, 4.3% H_2O, and 7.7% CO_2. These highly undersaturated rocks have been related to Late Paleozoic-Early Mesozoic rifting of the Gondwanaland.

Origin and geotectonic implications

Synthesis of field, petrographic and geochemical data suggests that the northwestern margin of the present day Indo-Pakistani landmass experienced widespread magmatic activity presumably in response to an extensional tectonic regime in a continental interior setting. The rocks can be grouped into Panjal Volcanics, Kashmir–Hazara–Peshawar dolerites, and a number of plutonic complexes constituting the Peshawar Plain alkaline province. The first two groups of rocks have been assigned to mainly Upper Paleozoic on stratigraphic grounds, but a controversy emerged from two sets of radiometic ages on the rocks of the PAIP. Based on post-Cretaceous ages, the rifting was assumed to be related to rebound relief tension, following compression release, initiated by collision between the Kohistan arc and the Indian plate (Kempe and Jan 1980, Kempe 1983). Jan and Karim

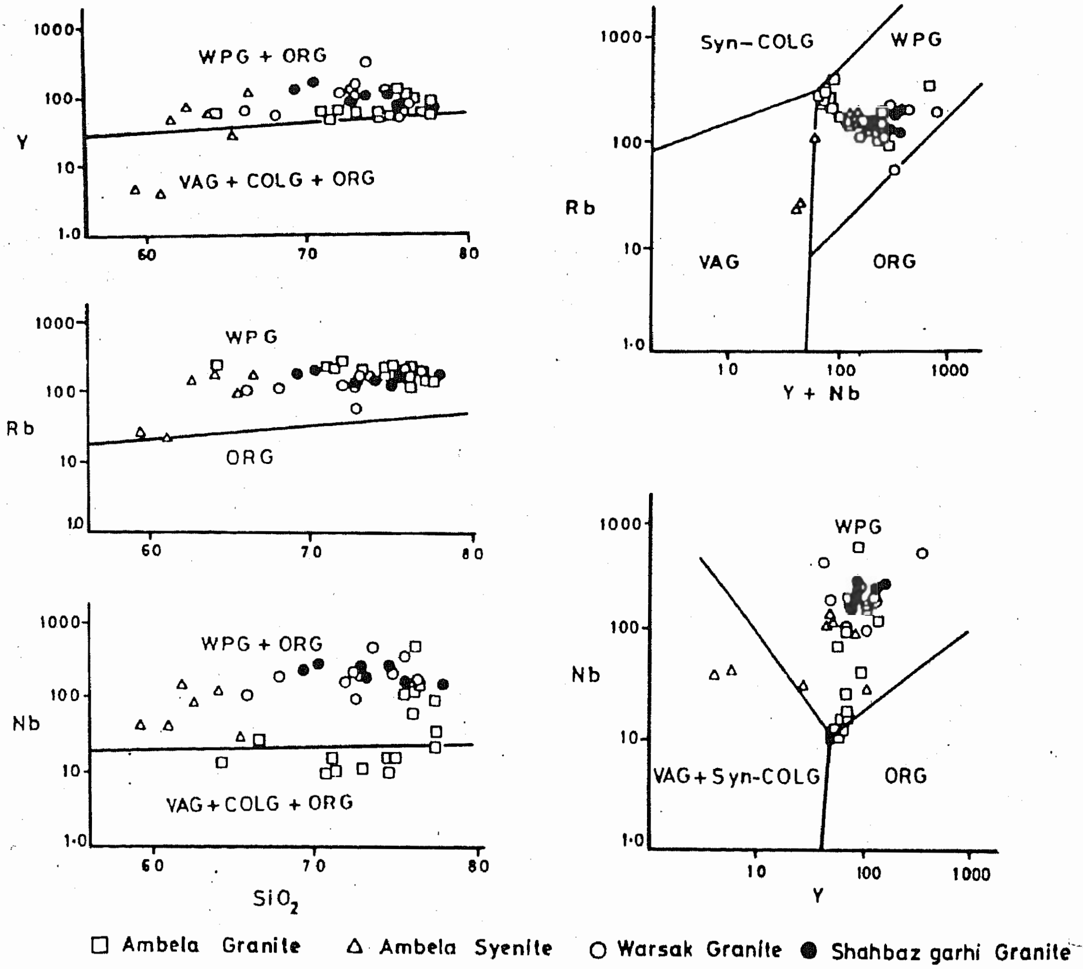


Fig. 6.6. Tectonic discrimination diagrams for silicic rocks. The PAIP rocks are characterised as within-plate on these plots. Field boundaries after Pearce et al. (1984). Source of data: Ambela–Rafiq and Jan (1990), Warsak–Khan (1991), Shahbazgarhi–Ahmad et al. (1990).

(1990) interpreted the post-Cretaceous set of dates to be related to thermotectonic events, and the Late Paleozoic-Early Mesozoic set, backed by stratigraphic ages, as the emplacement period. If so, this would mark an extensive rifting episode of Permo-Carboniferous age, covering the entire region between the Indus Suture and Khairabad Thrust from eastern Afghanistan to Zanskar and beyond. Pogue et al. (1992) have provided some field and stratigraphic data on the extensional tectonics in the region. Sedimentary environments during eruption of the Panjal Volcanics were coastal subaqueous to subaerial and terrigenous (Nakazawa et al. 1982, Honegger et al. 1982, Greco 1991), and the volcanics and sediments accumulated in a tectonically active basin (Acharyya 1973). The rocks indicate rifting on a previously stable continental shelf (Andrew-Speed and Brookfield 1982, Honegger et al. 1982, Humayun et al. 1987, Papritz and Rey 1989).

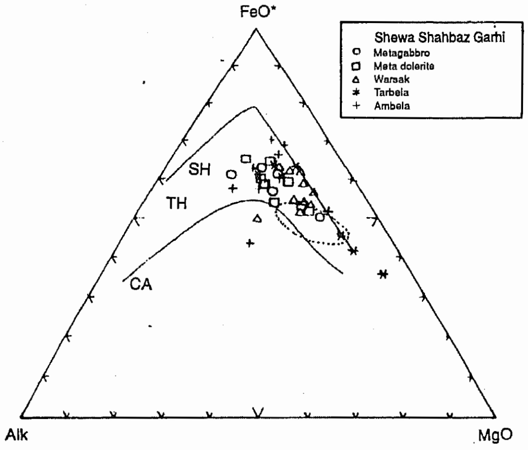
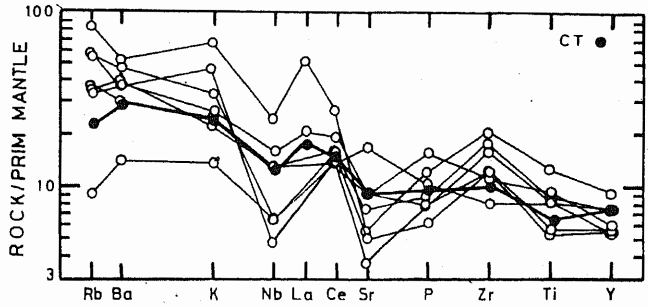


Fig. 6.7. MFA diagram for the mafic rocks of the PAIP. CA = boundary between calc-alkaline and tholeiitic series, after Irvine and Baragar (1971); SK = trend of Skaergaard liquid. All iron expressed as FeO*. Rocks containing excess alkalies fall off the trend of Skaergaard liquid, but occupy the tholeiitic field. Dolerites NW of Tarbela (Majid et al. 1991) plot in the dotted field. Source of data: Shewa-Shahbazgarhi–Ahmad et al. (1990), Ambela–Rafiq (1987), Warsak–Khan (1991), Tarbela–Jan (unpublished).

Fig. 6.8. Mantle-normalised spidergrams for the Panjal basalts. Normalising values after Wood (1979). These show features (such as negative slope, Nb trough and shape of the spidergram) similar to average continental tholeiite (CT) of Holm (1985). Source of data: Honegger et al. (1982).



The Paleozoic stratigraphy of southern Tibet (Lhasa block) is identical to that of Gondwanaland (Tapponnier et al. 1981a), raising the possibility that the Lhasa block may have been contiguous with, or not separated by a large open ocean from India during Permo-Carboniferous. This is further supported by the occurrence of volcanics of the same age and petrography as those of Panjal in the Karakoram plate. The eruption of Panjal Volcanics was followed by the evolution of an Atlantic-type continental margin in Triassic (Steckler and Watts 1978, Papritz and Rey 1989). The character of volcanism from tholeiitic to mildly alkaline during the Permo-Carboniferous changed to typical alkali-basalt series during the Triassic. Thus, Late Paleozoic rifting led to the breakup of microcontinents from Gondwana and opening of an ocean basin during Triassic and Early Jurassic. This rifting and magmatism may have been linked with a hot spot (cf., Gupta et al. 1982). From the above account it can be concluded that:

1) The three groups of rocks (PAIP, Panjal Volcanics and dolerite dykes) originated in the period Early Carboniferous to Permian/Early Triassic, with the Panjal volcanism reaching a peak during Late Carboniferous to Permian. The genetic relationships in the three groups of rocks are not clear, but the dolerites may have acted as feeders to flows in at least some cases. The onset of extensional tectonics, to which the rocks are related, is indicated by clasts in Early Carboniferous rocks north of Rustam.

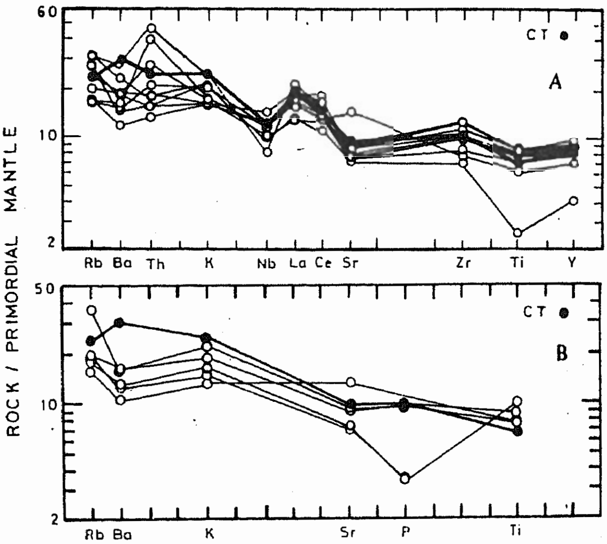


Fig. 6.9. Mantle-normalised spidergrams for the dolerite dykes in the Attock–Cherat Range (A) and Mansehra (B). Normalising values after Wood (1979). Their patterns closely match average continental tholeiitic basalt (CT) of Holm (1985), except for the segment Rb-Ba, a consequence ofmobility of these elements during alteration. (From Jan and Karim 1990).

2) Several of the PAIP complexes are bimodal (cf., Piccirillo 1987), the granitic rocks sharing the affinities of those from continental extensional regimes. The mafic rocks are mostly tholeiitic with traits of continental flood basalts. Despite petrographic diversity from mafic to silicic and tholeiitic to alkaline compositions, geochemistry and paleogeo-graphy clearly suggest that all the three groups of rocks originated within a continental plate during an extensional regime.

3) The three groups of rocks occur in a >600 km long and 150 km broad zone stretching along the northwestern edge of the Indian plate between the Indus Suture and the Khairabad Thrust. The dolerites occur in much of the area, whereas the PAIP and the Panjal Volcanics occur, respectively, to the west and east of the Hazara–Kashmir Syntaxis. Whether this means that the area to the west of the syntaxis is eroded to a deeper level or the volcanics did not extend that far cannot be ascertained at this stage. Additional work is required, especially in comparing the Swat amphibolites with the Panjal Volcanics.

4) Continental rifting is commonly associated with three types of igneous activity: intrusion of (a) basic dykes and (b) alkaline complexes, and (c) extrusion of lavas (Windley 1984). The PAIP–Panjal–dolerite trio in NW Himalayas is a perfect analogy.

JURASSIC - CRETACEOUS

The northeastern part of Balochistan shows several occurrences of alkaline rocks of Early Jurassic to Late Cretaceous age. These constitute some 130 km long belt trending NE-SW between Zhob and NE of Quetta. The presence of a variety of alkalic rocks in the western margin of the Indian plate has an important bearing on the geodynamic configuration of this part of Balochistan. The salient petrological aspects of the rocks are given in the following in a chronological order.

Alkalic rocks of Spangar–Kozh Kach area

The general features of the geology of the Spangar kimberlitic lamprophyre were presented by Ahmed and McCormick (1990), followed by further details by Ahmed

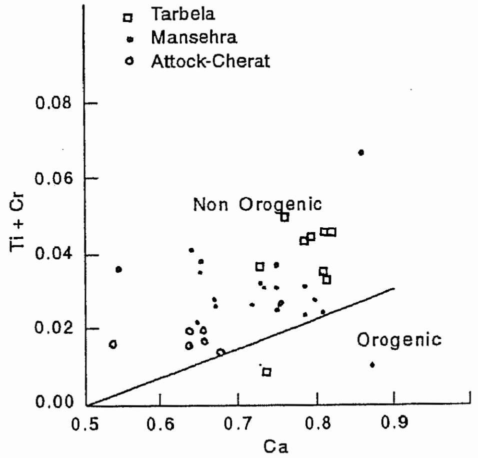


Fig. 6.10-A. Tectonic discrimination diagram, based on the composition of clinopyroxenes in the mafic rocks of the PAIP. Fields for orogenic and non-orogenic basalts are after Leterrier et al. (1982). Source of data: Mansehra–Ahmed (1985b), Attock-Cherat–Karim and Sufyan (1986), Tarbela–Jan (unpublished).

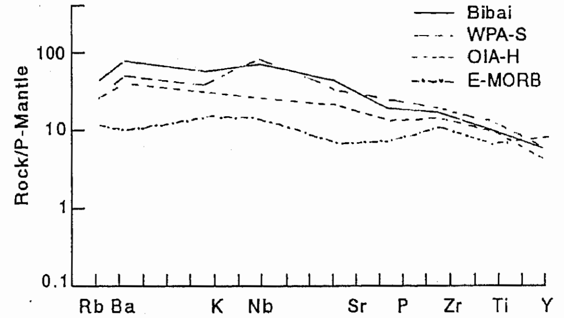


Fig. 6.10-B. A comparison of primordial mantle- normalised spider diagram for average Bibai alkali basalt, average Hawaii alkali basalt (OIA-H), average within-plate alkali basalt (WPA-S) and average E-MORB (Wood et al.1979). Adopted from Siddiqui et al. (1994).

(1991a). The lamprophyre forms a cluster of small linear dykes, sills, and lensoid, conical and pipe-like bodies and plugs. These intrude thin-bedded limestone (with minor shale component) of the Early Jurassic Shirinab Formation. Closely associated spatially and temporally with these lamprophyres are alkalic basaltic rocks. All the rocks are located in a 7.5 km long and 1.5 km broad area stretching E-W. Presence of E-W trending faults suggests that the lamprophyric outcrops may be fault-controlled. Another occurrence of basic (sodic alkalic) dolerites was reported near Kozh Kach Village (30° 37' N; 67°28' E), 20 km NE of Spangar. These occur in the limestone of lowermost Jurassic age (the Spingwar Member of the Shirinab Formation), but are lacking in the overlying Loralai Limestone of Early Jurassic (Ahmed 1991a).

The kimberlitic lamprophyre consists of olivine of two generations, phlogopite, perovskite, monticellite, garnet, chromian spinel, very rare pectolite, nepheline and, in one sill, melilite. Secondary chlorite and carbonate also occur. A few of the bigger outcrops contain mafic and felsic patches (with coarse phlogopite) indicative of some differentiation. Mineral chemistry of the rocks has been presented by Ahmed and McCormick (1990).

The alkalic basalt is porphyritic and amygdaloidal. It consists of kaersutite, plagioclase of variable composition, clinopyroxene, chlorite, biotite, calcite, chromian spinel, pyrite, chalcopyrite, ilmenite, quartz, epidote, talc, garnet, magnetite and titanomagnetite. Many of these minerals are secondary and the rocks show alteration to carbonate-talc/

chlorite assemblage. The clinopyroxene contains 0.6 to 3.6 wt% TiO_2.

Primary minerals in the sodic dolerites are clinopyroxene and plagioclase, but the rocks are commonly altered with carbonate + talc or chlorite as the extreme product. They also contain epidote, magnetite, ilmenite, anatase/rutile, and pyrite. Amygdules, where present, contain chlorite, calcite, and feldspar laths coarser grained than the host dolerite minerals. These rocks, like the alkali basalts, are olivine- and nepheline-normative. Their trace element geochemistry and the pyroxene composition in the alkali basalt suggest within-plate environment, and Ahmed (1991a) related them to horizontal extension and rifting in the western margin of the Indian Gondwanic Domain.

Tor Ghar Nepheline Syenite

Jadoon and Baig (1991) reported a circular, cone-shaped, undeformed alkaline intrusion from the Sulaiman fold-and-thrust belt. It occurs near Tor Ghar (30°18'25" N, 68°48'15" E), some 10 km ESE of Loralai. Measuring about 2 km across, it intrudes the Early Cretaceous Sember Formation, but has not been seen intruding Tertiary rocks. The rock is porphyritic and mostly made up of alkali feldspar, nepheline and biotite, with small amounts of clinopyroxene, apatite, muscovite and sphene; this mode classifies it as nepheline syenite. The intrusion is surrounded by an inner rim of finer grained basic composition and an outer one of hornfelses. No geochemical data are at hand. Jadoon and Baig (1991) suggest that the intrusion is a product of intra-plate magmatism over a hot spot, or rifting of the Indian plate through rebound relief tension during initial collision (compare Kempe and Jan 1980).

Bibai Volcanics

The upper part of the Cretaceous Parh Formation throughout NE Balochistan shows evidence of a widespread volcanic activity. The limestones in this formation contain ash beds, their lateritic derivatives, and fragments of lava and agglomerate (Kazmi 1984, McCormick 1985). In the Kach area, 45 km NE of Quetta, a thick succession of volcanic ash, tuffs, agglomerates and basaltic lavas, referred to as the Bibai Formation, overlies the Parh Formation (Barremian to Campanian) and underlies the Dungan Formation (Maastrichtian to Paleocene).

The petrography of the Bibai Formation, has been presented by Kazmi (1984). The flows are dark green to black, porphyritic, and amygdaloid, commonly exhibiting flow structure. They consist of a range of primary and secondary constituents: olivine, titaniferous- and aegirine augite, iron oxide, orthopyroxene, calcic plagioclase, celadonite, chlorophaeite, palagonite, lussatite, chalcedony, calite, zeolite, serpentine, and iddingsite Sawada et al. (1992) report that the flows consist of basanite, alkali basalt, trachy basalt, tephrite and phonolitic tephrite. They studied a large number of samples and noted that they consist of phenocrysts of olivine, titaniferous augite ± plagioclase ± opaque oxide in a groundmass containing these phases, and pargasite, phlogopite, apatite and glass, with analcine in the more evolved types. Overlying the flows and agglomerate of the lower zone is a thick sequence of conglomerates inter-bedded with ash beds and tuffs (Fig. 4.17). The tuffs are mixed with non-volcanic terrigenous sediments in the upper part (Kazmi 1984). The conglomerates, with a maximum thickness of 350 m, contain pebbles and boulders (ranging 3 to 40 cm in size), in a highly altered matrix which contains clinopyroxene. With the exception of rare pebbles of Parh limestone, all the pebbles and boulders in these conglomerates are those of the tuffs, agglomerates, basalts or andesites. Since andesites do not occur as flows, they may represent the upper level flows (of the lower zone) which were eroded away. According to Siddiqi et al. (1994), these volcanics form a 2 km wide and 1,200 km long belt stretching from Waziristan to Karachi.

Kazmi ((1984) suggested that the Bibai volcanism was initially submarine, as revealed by pillow lavas, followed by a subaerial phase. Based on limited clinopyroxene analyses from basalt and agglomerates, McCormick (1985) suggested that (1) the Bibai rocks are within-plate alkali basalt formed in an oceanic island and obducted onto the Mesozoic sediments on the western side of the Indian craton, and (2) ophiolites, occurring just to the west of these basalts and their interspersed marine sediments, have been tectonically obducted onto these from the west. On the basis of major and trace element geochemistry, Sawada et al. (1992) and Siddiqi et al. (1994) showed that the rocks are indeed similar to within-plate alkaline basalts and may represent hot spot magmatism related to the break-up of the Gondwanaland (Fig. 6.10). Major, trace and rare-earth element geochemistry suggests that the alkaline magma was produced by about 15% melting of an enriched garnet lherzolite source. The average Bibai alkali basalt is much enriched in Th, Ba, Rb and K relative to average oceanic island alkali basalt. This is taken to suggest contamination of the parent magma by partial melts of the Indian continental crust en-route to eruption. The occurrence of the associated coarse clastic rocks, laterite and ferruginous conglomerate in different areas has been attributed to change of depositional environments caused by doming related to a hot spot.

Porali suite

The Mor Range intrusive rocks and Porali Volcanics, which occur on the flank of the Bela Ophiolite, have been described by Sarwar (1981, 1992). These were emplaced before the Paleocene-Cretaceous obduction of the ophiolite and have been considered as probably related to Gondwana rifting and Deccan volcanism. The intrusive rocks consist of sills and dykes of ultramafic to mafic composition. The Porali Volcanics occur in small to large (km-size) outcrops adjacent to the Kanar Melange, but some may even be incorporated in it. They consist of volcanic agglomerate (debris flow ?) and minor flows, possibly deposited in thick (up to 1.5 km) piles around volcanic centres. They comprise aphyric to prophyritic, titanferous augite-bearing basalts which plot in the field of within-plate basalts on Hf-Th-Ta diagram. Their rather high Th content may suggest contamination by continental crust material (see Wood et al. 1979). The intrusives are of the same composition as flows.

Pir Umar Basalts

A minor structural slice of pillowed basalt occurs 24 km south of Khuzdar and 2 km north of Pir Umar rest house (27°39'30" N; 66°37'E). The neighboring rocks are entirely sedimentary and the basalt is emplaced in the upper part of the Jamburo Group limestone (HSC 1960) of Paleocene to Oligocene age. The basalt contains microphenocrysts in a pilotaxitic and intergranular matrix. It is made up of slender laths of fresh feldspar in a fine-grained chlorite matrix, abundant (over 15 vol.%) ilmenite+magnetite+ulvospinel, and minor calcite along fractures. It is a sodium-rich olivine- and nepheline-normative alkaline basalt with a high Nb content (62 ppm) like within-plate basalts. The geochemical characteristics, especially Zr/Nb and Y/Nb resemble those of oceanic island basalts. Ahmed (1990b) discussed the petrogenesis of the rock and suggested that the source magma of the rock was produced from some ancient oceanic lithosphere recycled into mantle. The magmatic event may have occurred when the Indian plate passed over a hot spot during its northward movement. The age of the rock is not known, but the model proposed by Ahmed (1991b) would suggest that it may be Cretaceous in age. The rock was probably emplaced tectonically in younger sediments.

"Twin Sisters" soda dolerite, Muslimbagh

Some 22 km southeast of Muslimbagh, Shams and Ahmad (1976) reported two soda dolerite dykes in a sequence of calcareous shales, flaggy limestone and sandstone of possible Cretaceous age. The dykes are 15 and 4 m thick and separated by a 2 m thick zone of rubble of dolerite. They are pyroxene (up to 2.5 cm) phyric and consist of plagioclase (>50%), clinopyroxene, biotite, relics of olivine (Fo_{80}), nepheline, chlorite, serpentine, actinolite, leucoxene, tiny grains of iron oxide, calcite ± green hornblende. The pyroxene is commonly replaced by chlorite, actinolite and serpentine, and the plagioclase is cloudy and albitised. There are substantial differences in the chemistry of the two samples analysed by Shams and Ahmed (1976), notably in their MgO, CaO, Na_2O, and normative olivine and nepheline contents. These features, the altered nature of the pyroxene and, especially, plagioclase, and late development of nepheline and calcite may have resulted from autometasomatism. Shams and Ahmed think that the rocks represent an alkaline magmatic activity in the area. Petrographically, the rocks are not very different than Pir Umar basalts.

Origin and geotectonic implications

Diapiric rise and melting of enriched mantle is considered as a plausable mechanism for the generation of magmas which can produce ultramafic-mafic alkaline rocks such as those of Balochistan. Ahmed (1991a) suggested that the Spangar rocks represent intracontinental magmatism related to Jurassic rifting and splitting of the western margin of the Indian plate away from its position adjoining the part of Afro-Arabian plate near Somalia. Although this magmatism ceased by mid Jurassic, the same extensional tectonics may have continued to provide channels for Bibai Volcanics of Maastrichtian age. These volcanics have been considered to be related either to a hot spot (McCromick 1985) or island arc magmatism (Kazmi 1984). The former veiw stems essentially from the composition of titaniferous augite, subsequently supported by the geochemistry of the rocks (Sawada et al. 1992), and the latter from the presence of andesitic flows. The volcanics occur in a belt of ophiolites and melanges of probable polygenic origin. Siddiqi et al. (1994) argue that the Bibai Volcanics are related to a hot spot underneath the Indian plate. Clearly, this region is geologically complex and requires further studies. The Pir Umar basalt shows oceanic basalt geochemistry. The age of this rock is not known but it may be an altered (sea floor metasomatism) equivalent of the Bibai Volcanics, emplaced tectonically in younger rocks. The Porali Volcanics have been considered as related to the Late Cretaceous Deccan volcanism of continental affinity (Sarwar 1981). The Tor Ghar nepheline syenite may be related to thinning of the continental crust during a regime of extension. In essence then , the alkaline magmatism may ultimately be related to hot spots and mantle plumes which produced a range of alkaline rocks during the Early Jurassic and Late Cretaceous in, possibly, both continental and oceanic settings. There are low-grade metamorphosed volcanic rocks (Koghozi Greenschist) of Jurassic age in Chitral. These will be described under Karakoram plate in a later section.

CRETACEOUS – TERTIARY

To this age belong ophiolitic complexes, commonly thought of as traces of plate boundaries (Coleman 1977). The description of these is followed by magmatism in Sindh, the Chagai and Kohistan arcs and Karakoram–Hindu Kush region. In the end a summarised account of the Himalayan leucogranites is presented.

Ophiolites

Ras Koh Range

The Ras Koh Range of west-central Balochistan is a tectonic block lying to the south of the Chagai magmatic arc and to the west of the Chaman Transform Fault (Photo. 15). The range is an E-W to NE-SW trending anticlinorium both limbs of which are complicated by numerous folds and faults that trend generally in NE-SW direction. The northeastern part of the range contains altered andesitic rocks of the Sinjrani Volcanic Group (HSC 1960, Fatmi 1977), possibly belonging to the Chagai domain and thrust over the Ras Koh Range. There also are large plutons generally poor in quartz. The rest of the range has been mapped as consisting mostly Upper Cretaceous or Paleocene to Eocene sedimentary rocks containing scattered masses of ophiolitic and intrusive rocks, especially along its northern margin.

The Sinjrani-type rocks in the Ras Koh Range form strips up to 13 km wide. They consist of agglomerate, lava flows, and tuffs, with subordinate limestone, tuffaceous shale and sandstone near top. Agglomerate beds are up to 16 m thick and consist of red to green fragments of lava and tuff in a matrix of dark green or maroon tuff and lava. The lava flows and flow breccia are up to 3 m thick andesite or basalt with minor rhyolite. They contain augite, brown hornblende, and plagioclase phenocrysts, and may be pillowed. Secondary epidote, carbonate, and chlorite occur along fractures and in the top and bottom of the flows (HSC 1960). Radiolarian fauna in chert interbedded with these volcanics suggest Middle Jurassic to Late Cretaceous ages. Thus they have been named Kuchakki Volcanic Group by HSC (1960) and Siddiqui et al. (1995) because of their older ages.

The oldest rocks in the rest of the Ras Koh Range are represented by small plutons (< 3 km at their maximum) of granodiorite/quartz diorite along its northern margin, such as those of the Belri Bargo area. They occur as inliers and their xenoliths have been reported in the volcanic members of the Paleocene Rakhshani Formtion (HSC 1960). Therefore, these plutonic rocks are older than the Paleocene. They are leucocratic and contain hornblende, clinopyroxene, magnetite and secondary chlorite, biolite ± amphibole ± calcite (Bakr 1963). The bulk of the Ras Koh Range is occupied by the Rakhshani Formation (Photo. 40), a sequence of interbedded shale, sandstone and limestone with subordinate intercalations of volcanic rocks (Ahmed 1961, Bakr 1963). The Rakhshani Formation conformably overlies the limestone of the Kharan Formation.

Volcanic rocks in the Rakhshani Formation consist of flows, agglomerate and tuff of basalt and rhyolite. This bimodal magmatism was contemporaneous, because (1) basalt and rhyolite are locally interbedded, (2) sills (and dykes ?) of rhyolite occur in basalt and vice versa, and (3) there are mixed basalt-rhyolite tuffs suggesting simultaneous eruption of mafic and silicic volcanics. The contemporaneous rhyolite-basalt volcanism was explained by Bakr (1963) as resulting from tapping of different levels of a zoned magma chamber with rhyolite top and basalt base. Detailed geochemistry is required to test rival hypotheses. In the Newberry Volcano of central Oregon, for example, rhyolite obsidian erupted simultaneously with basalt is considered as crustal melt, generated by heat associated with basalt magma (Linneman and Myers 1990).

The basalts (locally pillowed) are porphyritic and consist of augite- and hornblende (± quartz)-bearing varieties containing secondary epidote, chlorite and quartz. Some rocks contain glass in the matrix. Amygdules, where present, are composed of calcite and zeolite. The agglomerate is made up of basaltic fragments in a tuffaceous matrix. The fragments are well-rounded and polished, suggesting rolling and water transport, and the matrix contains shallow-water fossils. The tuff intercalations are locally welded and the mafic type may contain euhedral hornblende and augite.

The rhyolite contains quartz phenocrysts in a fine-grained to cryptocrystalline quartzofeldspathic groundmass with minor clinopyroxene partially replaced by chlorite, calclite and amphibole. The rocks range from white to grey, purple and green in colour. Spherulitic texture is common and some rocks may be rhyolitic to dacitic in composition. Dolerite dykes, striking parallel to the ESE trend of the Ras Koh Range occur at several places. They intrude the older granitoids and rhyolite flows but not the basalt flows. They are modally similar to the latter and may be their feeders.

The younger plutonic rocks comprise stocks and minor bodies apparently intruding the Rakhshani Formation. These range from gabbro and norite , through diorite and monzonite, to aplite. The bulk of these rocks occurs in a 12 × 8 km area around Belri in northwestern Ras Koh Range, but there also are many smaller bodies along the northern margin of the range. The larger of these bodies consist of diorite, and gabbro + norite. Some of the diorites pass gradually into a core of monzonite, but in some cases the latter intrude the former.

There are many bodies of ultramafic rocks, especially along the northern and western margin of the Ras Koh Range. According to McCormick (1985), these, together with the younger intrusions (i.e., gabbros, diorites. mozonites), constitute an ophiolitic suite. Most of the ultramafic bodies are small. Shearing and folding is common and the bodies are fault-bounded which suggests tectonic emplacement. The principal occurrence of the ultramafic rocks is seen in the western-most part of the range in a belt 40 km long and up to 5.5 km broad. The rocks consist of serpentinised dunite, peridotite (with small pods of chromite), and coarse-grained pyroxenites. The peridotites contain relics of olivine, pyroxenes and chromite in a serpentine matrix.

Asrarullah et al. (1979) suggested that the Ras Koh Range is a subduction-related complex probably formed during the Eocene. It was thrust southwards in a manner similar to the southward thrust of the Chagai magmatic arc over Ras Koh Range (Farah et al. 1984a). McCormik (1985) analysed clinopyroxenes from the Ras Koh basaltic volcanics and noted their affinity to within-plate alkali basalts. He proposed that the Ras Koh Range represents a mass of oceanic basaltic islands which collided with the Chagai arc. Geochemical data at hand are insufficient to elucidate the petrogenetic and tectonic setting of the magmatic rocks of the Ras Koh Range. It may be a collage of terranes the northeastern part of which may belong to the Chagai arc.

The detailed study of the western part of the range (Bakr 1963) shows that (1) there was a bimodal, contemporaneous basalt-rhoylite magmatism, (2) limestone and sandstone deposition took place in shallow to moderately deep waters, (3) intraformational conglomerates contain water transported clasts of volcanics, (4) the early sediments and volcanics were deposited on a basment of granitoids, and (5) at least some of the younger plutons display contact aureoles and are intrusive rather than tectonically emplaced. We, thus, are of the opinion that the bulk of the Ras Koh Range may be of continental affinity. The ultramafic rocks and some of the younger plutons (i.e., gabbros) may be ophiolitic (oceanic), but other possibilities (e.g., deep roots of the Chagai arc or a fore-arc basin) should not be summarily dismissed. There are schists and amphibolite in the western Ras Koh Range. These may represent metamorphosed wedges under the ophiolite, as at Muslimbagh (McCormick 1989). Ar-Ar dating of muscovite in an amphibolite from the metamorphic sole suggests an emplacement age of 110 Ma for the ophiolite (E. Gnos, personal communication, 1996).

The Bela–Zhob–Waziristan Suture Ophiolites

On the western margin of the Indian plate the suture is characterised by extensive belts of ophiolites and complex tectonic melanges emplaced onto the Indian plate during Paleocene. Some description of these has been given in Chapter 4. In the following, supplementary data are presented on these comlexes, from south to north.

Bela Ophiolite: This is the largest ophiolite complex of Pakistan (Fig. 6.11). It runs along the western boundary of the Indian plate for 400 km from northwest of Khuzdar to the coast of the Arabian Sea, west of Karachi. The ophiolite is mostly 10 to 15 km in width and forms a large nappe occupying the core of an open NNW-plunging syncline. Previously considered as an extension of the Deccan basalt province (Vredenberg 1909), it has recently been recognised as an ophiolitic complex (Stocklin 1977). The ophiolite has been described by DeJong and Subhani (1979), Gansser (1979), Ahsan and Akhtar (1984) and, especially, Sarwar and DeJong (1984) and Sarwar (1981, 1992).

The ophiolite is bounded on its eastern side by Jurassic and Cretaceous limestone and shale of the Mor Range. Its western contact is concealed under Quaternary deposits of the Bela embayment. In the north it is transgressed by the Nal Limestone of Oligocene age (Allemann 1979). West of the Mor Range, the ophiolite was thrust upon the Porali basaltic agglomerate and lava flows which appear to overlie the sedimentary rocks of the Mor Range without a major tectonic break. These rocks and the melange are cut by intrusions manifesting igneous activity unrelated to the ophiolite (Farah et al. 1984b).

The ophiolite is allochthonous; all the comprising rock-types occur commonly as fragments, blocks or slabs, generally a few tens of metres in diameter and displaying chaotic relationships. The ophiolite, thus, is a megamelange which, at Kanar, consists of ophiolitic and continental clasts in an argillaceous matrix. The Kanar Melange formed by a combination of complex tectonic and sedimentary processes during the Late Cretaceous-Paleocene. The melange horizons represent ophiolitic debris that was intermittently deposited on the irregular bottom of an ocean basin (Sarwar and DeJong 1984). Some blocks in the megamelange cover large areas. The largest of these, at Sonoro, is 116 km^2 in area and shows a complete ophiolite sequence from ultramafic tectonites at the base to pillow lavas and sediments at the top. Separated from each other by melange zones, the ophiolite

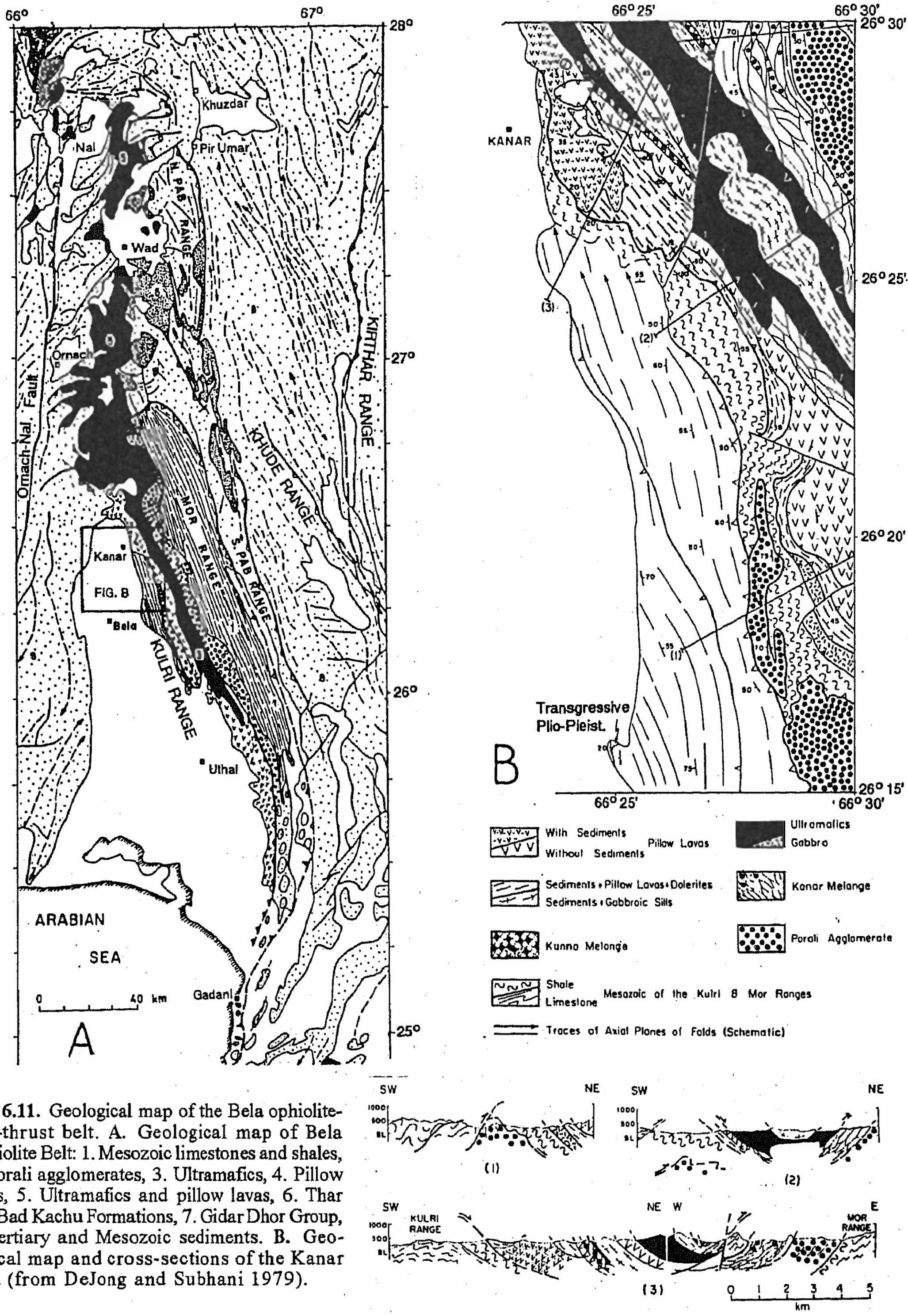


Fig. 6.11. Geological map of the Bela ophiolite-and-thrust belt. A. Geological map of Bela Ophiolite Belt: 1. Mesozoic limestones and shales, 2. Porali agglomerates, 3. Ultramafics, 4. Pillow lavas, 5. Ultramafics and pillow lavas, 6. Thar and Bad Kachu Formations, 7. Gidar Dhor Group, 8. Tertiary and Mesozoic sediments. **B.** Geological map and cross-sections of the Kanar area (from DeJong and Subhani 1979).

fragments contain the following rock groups:

1. Ultramafic rocks: These consist of (a) residual mantle harzburgite with minor dunite, and (b) cumulate dunite with subordinate wehrlite and clinopyroxenite, possibly occurring in cyclic repetition, but some clinopyroxenites form dykes (Ahmed 1990). Extensive alteration and small-scale tectonic intermingling between (a) and (b) have made it difficult to separate them on large-scale maps. The silicate phases in these are highly magnesian and the chromites generally have high Cr/(Cr+Al) ratios (Arif et al. 1996).

2. Mafic to silicic plutonic rocks: These include gabbros, diorites, quartz diorites, trondhjemites and granites. In some places, there are layered sequences, with olivine gabbros (containing anorthitic plagioclase) at the base, passing upward into gabbros, leucogabbros and diorites. Geochemical investigation by Ahmed (1993) shows that the plagiogranites are derived from basic magma through fractional crystallisation, whereas the potash-rich granitic rocks may have been generated by crustal anatexis by the residual heat of the ophiolite. U-Pb measurements on zircons yield a crystallisation age of 68 ± 3 Ma both for the trondhjemites and granites (Ahmed 1993). This is a little surprising; one would expect older ages for the plagiogranites since they would have most probably formed before the emplacement of the ophiolite onto continental crust.

3. Dolerite dykes: These are two types. One set, consisting of diabases, is restricted to the ultramafic rocks. The other occurs at the base of pillow lavas and consists entirely of dykes without intervening country rocks. Most dykes are up to 5 m in thickness and a few extend for more than a kilometre.

4. Volcanic rocks: These consist of pillow basalt, spilite, and keratophyre with associated pelagic sediments and mudstones of Aptian-Early Maestrichtian age (Sarwar 1992). The volcanic rocks are dominant in the southren part of the ophiolite.

The tectonic environments during the development of the ophiolite are yet to be clearly understood. DeJong and Subhani (1979) suggested that during the Late Cretaceous a volcanic island arc was developed on the western margin of the Indian plate. The chromite chemical data point to a more complex origin (Arif et al. 1996). This is further documented by the basaltic rocks. They have a composite "geochemistry, dominated by marginal-basin basalts. Presence of a proximal island arc is also indicated. At a few localities, oceanic-island type geochemistry indicates some diapiric upwelling as well" (Ahmed 1991c). Sarwar (1992) showed that the Wayaro area flows, although similar to the E-type MORB, differ in details such as La/Ta (9.6 to 18.0 as compared to 10 to 15 for the MORB). His stratigraphic, structural and petrological studies reveal that the ophiolite originated in a large oceanic fracture zone (leaky transform) which acted as a boundary between the Indian and the Neo-Tethys plates during the Cretaceous, analogous to the present-day Owen Fracture Zone. This ancestral fracture zone was destroyed when transform movement gave way to oblique convergence, culminating in the ophiolite obduction. Ar-Ar dates of 65-70 Ma were obtained on hornblende from the metamorphic sole of the ophiolite (Gnos et al. 1996). This timing is in perfect harmony with the U-Pb zircon age (68 Ma) of the granites related to the ophiolite emplacement (Ahmed 1993). For more information on the Bela Ophiolite, see Chapter 4.

Zhob Valley Ophiolites: The Zhob Valley ophiolite belt of northern Balochistan is the best-known occurrence of its type in Pakistan. It extends intermittently in a WSW-ENE direction for about 250 km between the Chaman Transform Fault to the west and the Indian shield to the east. The best exposures of the ophiolites are in the Muslimbagh-

Nasai area where some of the larger hills are entirely made up of these rocks (Bilgrami 1964). Because of their economic and tectonic significance, the rocks have been studied by many workers. Here we present a summarised account of the ophiolite in the Muslimbagh area, mainly after Moores et al. (1980). For additional references, the readers may consult Bilgrami (1963, 1968), Abbas and Ahmad (1979), Ahmad and Bilgrami (1987), Ahmed (1990), and Chapter 4.

The Muslimbagh Ophiolite consists of two principal tectonic massifs, the Saplai Tor Ghar–Nasai and the Jang Tor Ghar (Fig. 6.12), which are surrounded by melange containing blocks of pillow lava, diabase, ultramafics, limestone/marble, clastic sediments, and radiolarite. The ophiolite tectonically overlies Maestrichtian and older rocks and is overlain unconformably by shallow marine limestone of Eocene age. Ahmad and Abbas (1979), and Allemann (1979) consider that it was emplaced during the Paleocene. K-Ar ages of 81–82 Ma have been taken as indicative of time of formation and 67 Ma as time of emplacement of the ophiolite (Sawada et al. 1995). Ar-Ar dating of the metamorphic sole, and plastically deformed and recrystallised dolerite dykes also suggests an emplacement age of 65–70 Ma (Mahmood et al. 1995). Metamorphic rocks, such as hornblendite, pelitic schists (± garnet), and marble occur at the contact beneath the ophiolite at many places. The grade of metamorphism decreases rapidly away from the contact, suggesting that it owes its origin to residual heat in the ophiolite or frictional heat during its emplacement.

The ophiolite complex consists of ultramafic tectonites, ultramafic and mafic cumulates, a dyke complex, and a dolerite dyke swarm. Pillow lavas do not occur in the main mass; they are found as slivers in the melange. However, the ophiolite is complete to the south of Muslimbagh and contains all the lithologies, including pillow lavas (Ahmed 1990). The tectonites are serpentinised to different degrees, and consist of about 70% peridotite (lherzolite and harzburgite) and 30% dunite. The structures and layering of the tectonites generally plunge and dip southeastwards; their prime area of exposure is Jang Tor Ghar. The Saplai Tor Ghar–Nasai area comprises a large ultramafic complex dominated by dunite which may be thinly laminated and contains harzburgite layers (Photo. 41). Dunite also forms masses transgressing the layering and in pipe-like intrusive bodies.

Associated with the tectonites is a thick sequence of cumulate rocks with several cyclic units a few to several tens of metres thick. Commonly, these are olivine cumulates at the base and pass upwards through a thin layer of wehrlite into pyroxenite cumulate. The contact between the tectonite and cumulate ultramafic rocks appears to be planer and parallel to the attitude of the overlying cumulates. Plagioclase appears locally but in some of the uppermost cumulates there are gabbros as well as anorthosites (Photo. 42), which are well-layered and show an igneous lamination. Mineralogical data for the Muslimbagh area have been presented by Siddiqui et al. (1994). Olivine ranges from Fo_{94-91} in the tectonites to Fo_{92-72} in the cumulates. Orthopyroxene in the tectonites is En_{92-91}, and plagioclase ranges from An_{70-52} in the layered and foliated gabbros to An_{54-30} in the 1,000 m thick sheeted dyke complex and An_{30-8} in the plagiogranites.

The mafic to trondhjemitic dyke complex intrudes gabbro or diorite screens. A few trondhjemitic dykes cut all pre-existing rocks. A characteristic feature of the ophiolite is a large number of tholeiitic dolerite dykes. They reach up to 15 m in thickness, and display chilled margins. Some of them are composite. Sawada et al. (1992) noted that the sheeted dyke complex in Saplai Tor Ghar consists of dolerite, gabbro-diorite and

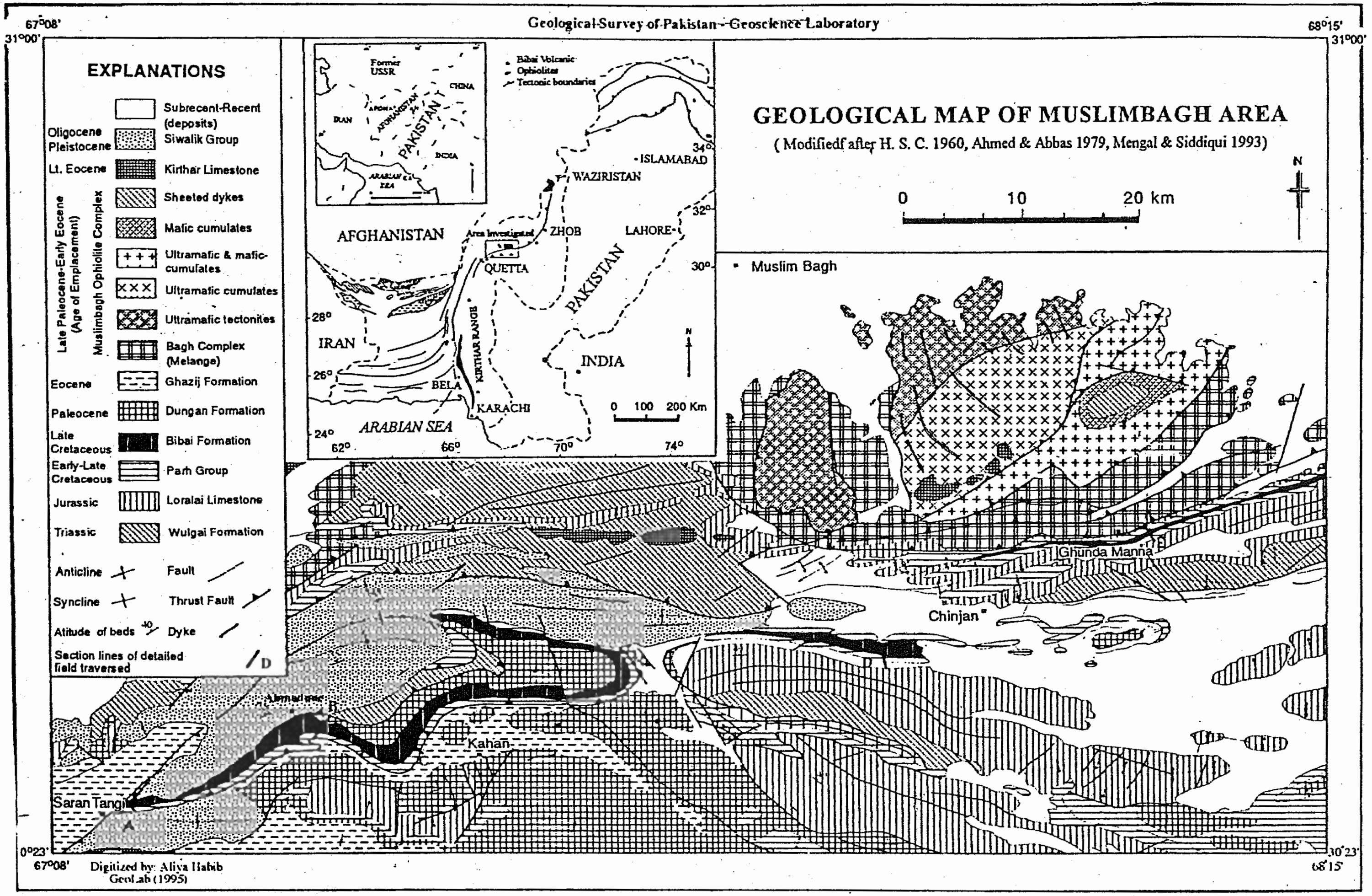


Fig. 6.12. Geological map of the Muslimbagh Ophiolite.

plagiogranite emplaced in three stages and displaying amphibolite to greenschist facies metamorphism and strong to weak foliation. The plagiogranites, according to them, were produced by wet melting of the metadiorites.

The melange zone was studied in detail by Kimura et al. (1993) who named it as the **Bagh Complex**. They divided it into six units consisting of sedimentary and igneous rocks representing the oceanic crust of the Neo-Tethys that was closed during the collision, a sedimentary sequence that was deposited on the western passive margin of the Indian plate, and a range of blocks derived from the ophiolite. Sawada et al. (1992) reported that the volcanic rocks in the melange can be divided into tholeiitic basalt of island arc or, probably, MORB affinity and alkali basalt (=Bibai type) derived from hot-spot magmatism related to break-up of the Gondwanaland. The sheeted dykes and pillow basalts in the melange zone of Muslimbagh Ophiolite, according to Siddiqui et al. (1994), are low-K tholeiites enriched in LILE and depleted in HFSE relative to N-MORB. Their trace and RE element geochemistry is consistent with a black-arc basin origin. The pillow basalts (tholeiites) and gabbro found towards the southeastern side of the melange zone show island arc-like signatures.

There has been a common belief that the ophiolite developed in a spreading centre and is oceanic in origin (e.g., Allemann 1979). But in the light of the recent views that many ophiolites may be related to island arcs or back-arc basins, other modes of origin cannot be ruled out. The generally high Cr/(Cr+Al) ratios in the chromites (Bilgrami 1968, Jan et al. 1984, Hoshino and Siddiqui 1993) are atypical of oceanic rocks (Dick and Bullen 1984). It has been suggested that the Muslimbagh Ophiolite may be a product of magmatism related to the break-up of the Gondwanaland under continental environments (Sawada et al. 1992) or it may have developed in a back-arc basin (Sawada et al. 1995). Mahmood et al. (1995) argue that the ophiolite represents a segment of ocean floor from the small and slow-spreading ocean branch of the Neo-Tethys, and shows a WSW-ENE-oriented obduction.

Waziristan Ophiolite: Occupying about half the Boya–Razmak area, the Waziristan Ophiolite is the third largest ophiolite of Pakistan after Bela and Muslimbagh. It consists of broken blocks and thrust slices which have over-ridden Jurassic to Cretaceous sediments. Eocene sediments, mostly limestone and shales with a red-oxidised zone towards their base, overlie the ophiolitic rocks unconformably. Thus the ophiolite was emplaced by the Paleocene (Khan et al. 1982, Jan et al. 1985). The age of emplacement is similar to that of the Bela and Zhob Ophiolites described in the previous pages.

The ophiolite displays a complex structure. The rocks are intensely folded, faulted and, in places, brecciated and granulated. Deformation has dismembered the ophiolite which now consists of a chaotically arranged stack of thrust slices and blocks. A complete normal-order sequence is nowhere preseved but all the members of a typical ophiolite (ultramafic rocks, gabbros, sheeted dykes, pillow lavas, pelagic sediments and plagiogranites) are present in different localities. The ophiolite is interesting from an economic point of view as it contains deposits of copper, chromite, manganese and magnesite (Khan et al. 1982, Badshah 1985).

The ophiolite occupies an area of 30 × 25 km. The ultramafic rocks include harzburgite, dunite (commonly displaying cataclastic fabric and serpentinisation), and pyroxenite. The dunite occurs mostly in layers and lenses in the harzburgite, and the pyroxenite as dykes and (?)layers. Feldspathic rocks, mostly gabbros, locally grading to feldspathic gabbros and anorthosites, are of limited extent and at places strongly altered. The sheeted dyke swarm is best developed to the north of Datta Khel. The dykes, fine- to medium-grained and mostly 1 to 10 m in thickness, may display asymmetrical chilled margins like those of the Troödos Ophiolite on Cyprus (Gass 1980). They consist of labradorite, augite, brown to green pleochroic hornblende, and ilmenite, with variable amounts of secondary/metamorphic chlorite and amphibole. Some dykes contain abundant actinolite, chlorite, albite, epidote, quartz, carbonate, and opaque grains (Jan et al. 1985).

Volcanic rocks make up more than half the area of the ophiolite. They include pillow lavas, tuffs and agglomerates/breccias containing fragments of volcanic, ultramafic, doleritic and dioritic rocks, jasper and limestone (Badshah 1985). The volcanic rocks are commonly altered and contain secondary chlorite, actinolite, epidote, calcite and iron oxide/hydroxide. Whitish dykes, commonly associated with copper mineralisation, traverse the volcanics. Some of these appear to be trondhjemitic and some rhyolitic to rhyodacitic in composition. Badshah (1983b) has also recognised calc-alkaline plutonic and volcanic rocks in the area.

Field studies (Beck et al. 1993), supplemented by limited data on the chemistry of rocks (Hamidullah 1994), chromite (Jan et al. 1985) and pyroxene (Ahmed and Hamidullah 1987), suggest that the ophiolite may be an island-arc type. In the adjacent Khost (Afghanistan), Kurram and Thal areas, there are small bodies (< 5 km^2) of folded and sheared ultramafic rocks, serpentinites and pillow basalts, intercalated with and overlain by bedded chert and radiolarite. These are enclosed tectonically in Cretaceous limestone, shale and tuff of the newly named Kahi Melange covering >15,000 km^2 area (Beck et al. 1995). Detailed stratigraphic and structural data, supported by radiometric dates, show that thrusting of glaucophane-bearing accretionary prism onto the NW Indian plate began after 66 Ma. By 55.5 Ma, this segment of the Neo-Tethyan ocean was closed and the accretionary prism and the 100 Ma ophiolite rocks were thrust over India (Beck et al. 1995).

The Indus Suture Ophiolites

The Indus Suture in northern Pakistan demarcates the Kohistan magmatic (island) arc terrane to the north from the Indian plate to the south. Much of the Kohistan terrane in the immediate vicinity of the suture consists of an extensive belt of amphibolites with a variety of other rocks ranging from ultramafic to granitic in composition. There are at least three prominent mafic-ultramafic complexes sitting on the suture: Tora Tigga in Dir, Jijal along the Indus, and Sapat along the Kaghan–Kohistan watershed to the NE of Naran. These have been considered as cumulates related to the Kohistan magmatism, therefore, their description is deferred to a following section. In this section we restrict ourselves to ophiolitic rocks which occur within the suture zone proper, either as components of tectonic melanges, or as uprooted ultramafic bodies.

Bajaur–Utmankhel: The suture zone in this area is apparently made up of a stack of thrust slices and blocks (Badshah 1979). The zone is a tectonic melange consisting of blocks ranging from a few tens of metres to several kilometres, and derived from Kohistan, Indian plate and the Neo-Tethys. The Kohistan lithologies are represented by amphibolites, and mafic to silicic plutons a few of which might be post-dating the suturing. The Indian plate rocks consist of low-grade metamorphosed pelites, calcareous rocks and quartzite; the pelites contain abundant veins of quartz and dykes of leucogranitic pegmatites. The Neo-Tethyan rocks comprise serpentinised peridotites, greenstones, and manganiferous sediments (piemontite schists). The ultramafic rocks are particularly abundant in the

Chamarkan area bordering Afghanistan.

In Umbar Utmankhel, a slice of highly fossiliferous limestone of Eocene age is closely associated with volcanic rocks. Both of these units may have been thrust from the north onto the Indian plate (Badshah, per. comm.). In this region there are several occurrences of economic mineral deposits, e.g., chromite, emerald, manganese, soapstone, and abundant marble.

Skhakot-Qila Ophiolite Complex: This ophiolite, also named as Dargai or Harichand complex, constitutes a southerly arcuate mass of ultramafic rocks that stretches E-W for 30 km, with an average width of about 3.5 km (Fig. 6.13). Several small bodies occur to the north, south and west of the principal mass (Badshah 1979, Hussain et al. 1984). It over-rides metamorphosed pelitic, calcareous and carbonaceous rocks of presumably Precambrian age. Gravity and magnetic studies suggest that it is an ultramafic klippe, less than 1.5 km deep, over-riding the Indian plate some 30 km to the south of the Indus Suture (Malinconico 1982). Hussain et al. (1984) have described it as part of an extensive melange zone.

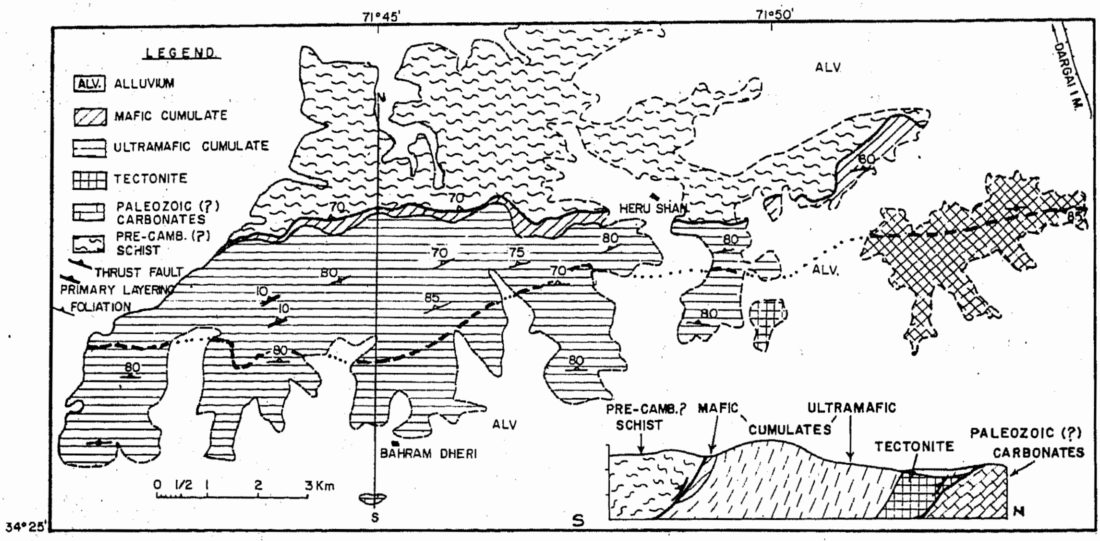


Fig. 6.13. Simplified geological map and section of the eastern part of the Dargai ultramafic complex (from Rossman and Abbas 1970, Ahmed 1984). Thick dashed line separates chromite with Cr/(Cr+Al) > 0.61 (south) from those with Cr/(Cr+Al) < 0.61 (north).(From Ahmad 1984).

The Complex has been considered as a dismembered ophiolite consisting of ultramafic tectonites, ultramafic cumulates, and mafic cumulates (Rossman and Abbas 1970, Asrarullah et al. 1979). Metadolerite dyke swarms containing Ti-bearing brown hornblende have been described by Ahmed (1984) near the ultramafic contact with gabbros. Extensive outcrops of greenstone and minor amphibolite, reported in the melange zone by Hussain et al. (1984), may represent the volcanic component of the ophiolite.

The ultramafic rocks consist dominantly of harzburgite, followed by wehrlite and dunite. Although tectonitic harzburgite is more dominant in the southern part and gabbros and dolerites occur in thin bands in the northern part, the contact between the ultramafic tectonites and cumulates is ill-defined. The complex is locally traversed by pyroxenite dykes containing highly magnesian minerals, barring a few ferro-websterites. Serpentinisation and steatisation are extensive but generally not intensive. Shear zones and small ophiolitic bodies in the melange have been converted into talc-carbonate ± fuchsite assemblages. Some of these, in the western part of the complex, host emerald mineralisation (Rafiq and Jan 1985). Qaiser et al. (1970) and Ahmed (1988) have given details of rodingite bodies. Podiform chromite deposits, some forming several tens of metres long seams, are mostly confined to minor, irregular dunite bodies, although pyroxene-bearing rocks may occur in direct contact with segregated chromite (Ahmed 1982).

Analyses of olivine, pyroxene, and amphiboles from the complex have been presented by Ahmed (1987a,b). Olivine ranges from Fo_{97} to Fo_{91} in chromitites and Fo_{93} to Fo_{88} in ultramafic rocks. In a few peridotites and several pyroxenites, however, olivine is less magnesian, down to Fo_{74}. Orthopyroxene has a constant composition (En_{91-90}) in harzburgite, orthopyroxenite and clinopyroxenite, but in websterite it is En_{90} to En_{85} and in Fe-websterite En_{79-78}. With rare exceptions, the wollastonite content of the orthopyroxene is below 5 mole%. The clinopyroxene is diopsidic, with higher Fe content in wehrlite and ferro-websterite. Several types of calcic amphiboles occur in the rocks.

Ahmed (1988) discussed the petrology of the ophiolite with the help of major- and trace element analyses of 100 rocks, along with rare earth elements for eleven of them. The chondrite normalised patterns of the gabbros and dolerites compare well with those of other ophiolites and Mid-Atlantic Ridge gabbros. He concluded that the assemblage of the ophiolite could have been cogenetically formed after moderate degree of partial melting in the lherzolitic source mantle.

The compositional variation of chromite has been related to petrogenesis (Ahmed 1982, 1984). The chromites have been stratigraphically divided into a lower (southern) zone with Cr # [Cr/(Cr+Al)] >0.61, and an upper (northern) zone with Cr # <0.61. The high-Cr ores are thought to have formed in an environment similar to stratiform deposits, whereas the high-Al chromites formed under "alpine-type" conditions. The full range of Cr # classifies the complex as Type II alpine-type peridotite according to the subdivision proposed by Dick and Bullen (1984). To this type belong the largest alpine-type peridotites such as those of New Caledonia, Samail, Bay of Islands, and Josephine. According to Dick and Bullen (1984), most Type II peridotites represent complex multi-stage melting histories not found at mid-ocean ridges. Such petrogenesis may include areas where a young island arc was constructed on older oceanic crust, or sections across the transitions from arc to oceanic lithosphere, from lithosphere formed at the earlist stage of arc or continental rifting to more typical oceanic lithosphere in a maturing ocean basin, or across small aborted intra-arc rift basins preserved within paleo-arcs.

In the southern part of the Malakand pass, there are several small lenses consisting of carbonate, quartz, talc ± fuchsite, embeded in Precambrain pelitic schists. These have a high Ni content (about 1,000 ppm) and are probably derived from ultramafic rocks (Jan and Shah 1987). They may be related to the main Sakhakot-Qila ophiolite, occurring about 8 km to the WSW. But the lenses show similar metamorphism and deformation as the surrounding schists and the possibility of an ancient suture merits further investigation.

Mingora–Shangla Ophiolite Melange: A series of tectonic melanges containing ophiolitic rocks characterise the Indus Suture zone in Swat (Fig. 6.14). This melange group has been divided into: 1) Mingora Ophiolitic Melange, 2) Charbagh Greenschist Melange, and 3) Shangla Blueschist Melange (Kazmi et al. 1984). These are a chaotic assemblage of rocks apparently derived from oceanic crust, volcanic arc, trench, and continental margin

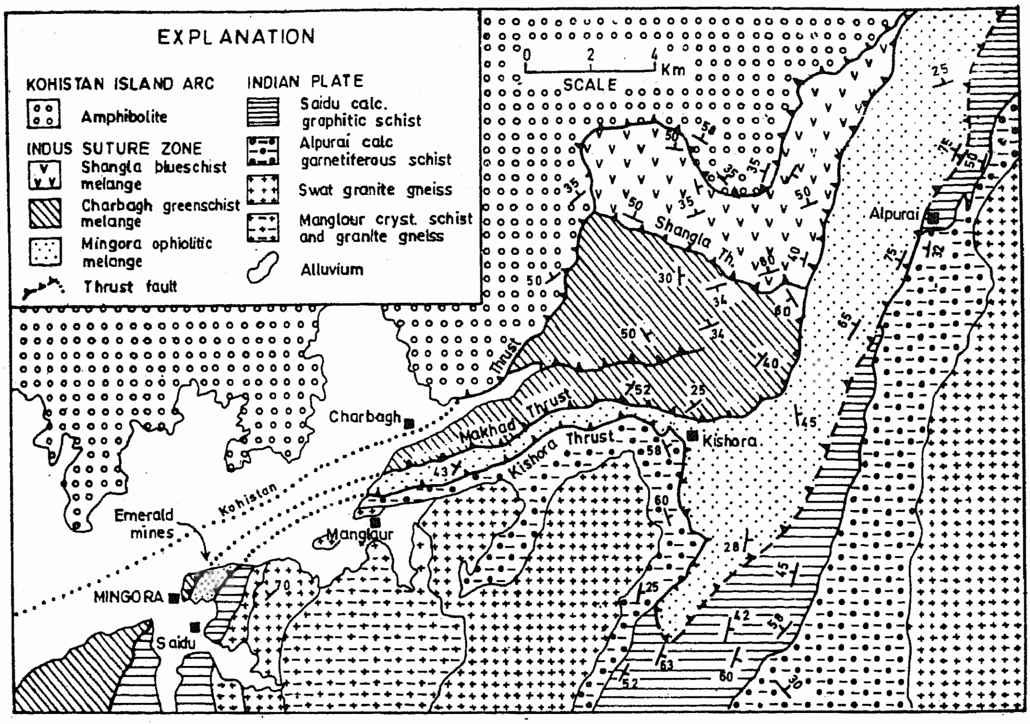


Fig. 6.14. Geological map of the Mingora–Alpurai area (from Kazmi et al. 1984).

(Jan and Jabeen 1990). Lithologies include altered ultramafic blocks, greenstone metabasalt, blueschists, greenschist metapyroclastics, metagabbros, plagiogranites and dolerites, metagraywake, metacherts, marbles and calcareous schists, quartz-mica-chlorite schists, and piemontite schists. Limestone blocks contain fossils of Jurassic to Middle Cretaceous ages (Kazmi et al. 1984), whereas the blueschist facies metamorphism has been dated at about 85 Ma (Shams 1980, Maluski and Matte 1984). Ultramafic bodies occur throughout the melange group, but they are more common in the ophiolite melange. Gabbro bodies are less abundant and commonly altered. The ultramafic mass east of Shangla (Khan and Humayun 1980, Chaudhry and Ashraf 1986, Arif and Jan 1993) contains relics of olivine, pyroxene, chromite, and small lenses of chromitite. The rocks have locally altered to dolomite, talc and/or quartz ± chlorite ± fuchsite ± epidote, especially on the eastern margin of the body.

The ultramafic rocks are mostly harzburgite, but some are dunite, lherzolite, websterite and diopsidite. Serpentinisation and metamorphism in greenschists and lower amphibolite facies make their demarcation difficult in the field as is the distinction between the ultramafic tectonites and cumulates. Chaudhry and Ashraf (1986) presented 30 chemical analyses of the ultramafic rocks. The high NiO content (>0.18%) is regarded by them as suggestive of mantle origin for the rocks. The molar 100 Mg/(MgO+FeO*) contents of most of the analyses fall between 87 and 91, with an average of 89 for the entire suite. Such values are typical of mantle peridotites (Loney et al. 1971, Himmelberg and Loney 1973, Dick 1977). The primary silicates are also similarly magnesian; some olivines were exceptionally enriched in magnesium ($Fo_{96-98.5}$) due to magnetite release during metamorphism (Arif and Jan 1993).

The gabbroic rocks, plagiogranites and albitites from the Mingora Ophiolitic Melange have been studied in some detail by Barbieri et al. (1994). These display low- to medium-grade metamorphic mineral assemblages. Gabbros contain albite, clinozoisite, calcic amphibole ± clinochlore ± sphene ± biotite, with local relics of augite. Plagiogranites are weakly foliated and consist of albite ($An_{0.2}$), quartz, actinolite, pargasite, and epidote. Albitites are foliated and contain accessory quartz, blue-green amphibole, monazite, zircon and allanite. Major-, trace-, RE elements and isotope data have also been presented for these rocks. Figure 6.15 shows the normalised plots of REE and trace elements in these and pillow lavas. The $^{87}SR/^{86}Sr$ values range from 0.70505–0.70674 for gabbroic rocks to 0.70527–0.70620 for plagiogranite, 0.70620 for pillow lava, and 0.70830–0.70841 for albitites. Barbieri et al. (1994) conclude that (1) the Na, K, Rb, and Ba contents of these oceanic rocks may have been modified due to interaction with sea water, (2) the very low contents of high field strength and rare earth elements (HFSE and REE) found in gabbroic

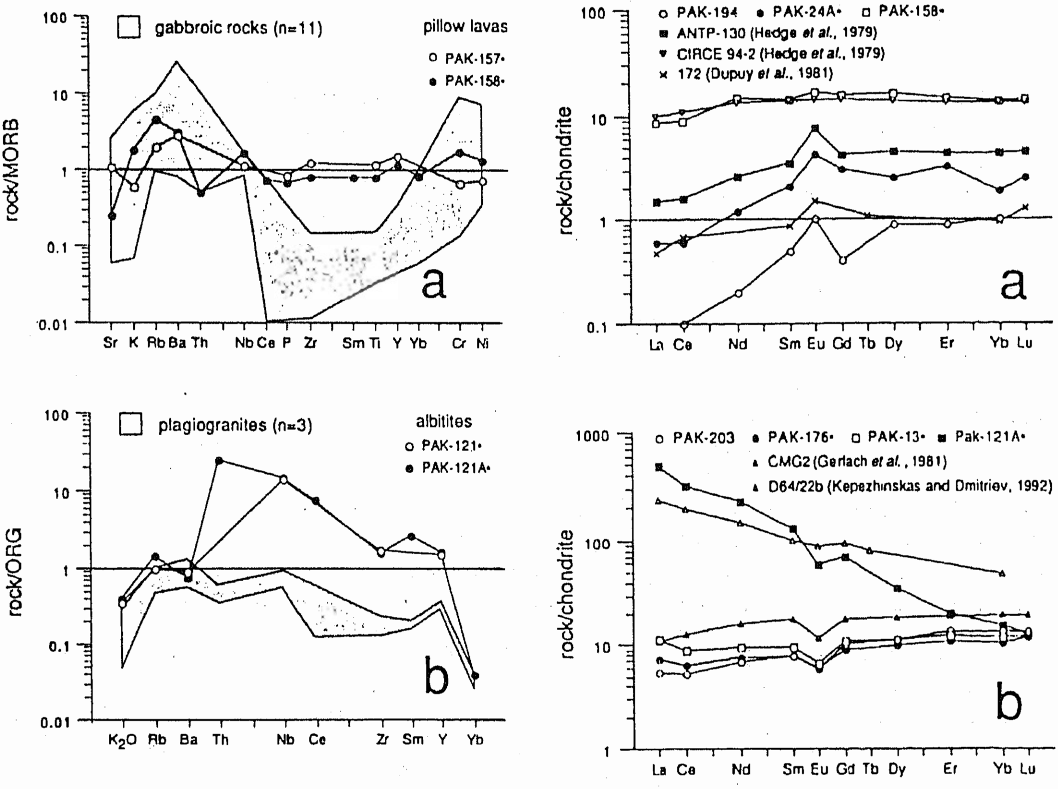


Fig. 6.15. Left. Spider diagrams for mafic and felsic rocks from the Mingora ophiolitic melange. Normalising values after Pearce (1983) and Pearce et al. (1984). **Right.** REE patterns for selected samples from magmatic rocks of the Mingora ophiolitic melange; (a) gabbroic rocks and pillow lava, (b) plagiogranites and albitite. The patterns of one lava (CIRCE 94-2), two gabbros (ANTP 130, 172), one plagiogranite (CMG2) and one granite (D64/22b), from other ophiolitic sequences or oceanic settings, are shown for conparison. Normalising values after Boynton (1984). (From Barbieri et al. 1994).

rocks and plagiogranite may be pointing to a cumulate origin from a refractory parental magma, (3) pillow lavas have N-MORB chemistry, and (4) the anomalously high HFSE, LREE and high $^{87}Sr/^{86}Sr$ ratios (very close to sea water) in the albitites may be due to a more complex origin of the albitites.

The lavas, locally interbedded with sediments, range from basic to intermediate in composition and may exhibit porphyritic and vesicular textures, but most are highly altered and, like the gabbros, contain abundant epidote, amphibole, chlorite and albite. Some are spilitic and represented by blueschists formed approximately at 7 kbar and 400° C (Jan et al. 1981d). Dykes occur only locally (Shams 1980) and a sheeted dyke complex has not been found. The absence of sheeted dykes has been related to high spreading rate (Arif 1994). Based on mineralogic data, especially chromite chemistry, Arif and Jan (1993) suggested that the ophiolitic rocks of Shangla may have a complex origin and developed under conditions transitional between those of island arc and oceanic settings. During their Early Tertiary obduction, the ultramafic rocks were first serpentinised and then metamorphosed under transitional greenschist–amphibolite facies conditions.

Allai Ophiolitic Melange: This melange occurs 30 km NE of Shangla between the Kohistan terrane and Indian plate. Like that of Mingora–Shangla, this has also been divided into ophiolitic-, blueschist- and greenschist melanges separated by a series of faults (Baig 1989). The ophiolitic melange consists of blocks and lenses of greenschist metapyroclastic, greenstone metabasalt, metagabbro, clinopyroxenite, dunite, peridotite, pillow lava, limestone, metachert and amphibolite in a sheared matrix of greenschist, serpentine or talc-carbonate. The blueschist melange consists of blocks of glaucophane-bearing metagraywacke, greenschists, phyllite, pelitic schist, metagabbro and minor lenses of talc, serpentinised peridotite, metachert and limestone in a sheared matrix of greenschist and phyllite. The greenschist melange contains blocks of greenstone, greenschists, amphibolites and metasediments in a sheared greenschist matrix. The melanges may be tectonically interleaved on their south with rocks of the Indian plate. All have gone through three phases of deformation.

The molar 100 Mg/(Mg+Fe*) ratios of peridotites (89-90) match those of mantle peridotites, but the pyroxenites have slightly lower ratios. Shah and Majid (1992) suggest that these represent a dismembered ophiolite suite. The lavas consist of olivine tholeiites and quartz tholeiites of island arc tectonic setting (Shah and Majid 1985), whereas the greenschists appear to be metamorphosed oceanic rocks (Shah 1985). The abundance of diopsidites over peridotites in the melange (Shah and Majid 1992) may be suggestive of polygenetic set up for the ultramafic rocks. Diopsidites are common in the Jijal Complex (considered as arc cumulates by Jan and Windly 1990) which borders the melange on its west.

Babusar Pass Melange: There is a >1 km broad melange zone containing highly mylonised and deformed rocks. In this zone there are two types of melange. The Kohistan rocks are thrust over a melange containing volcanics, with minor plutonic rocks and calc-schists. This is thrust over another melange containing beds and blocks of felsic and mafic amphibolites and marbles within a calc-schist matrix. This melange, in turn, is thrust over rocks of the Indian plate. The melange units represent mixtures of Kohistan and Indian plate rocks in a wide thrust zone and all the four units show three phases of deformation (Chamberlain et al. 1991). Spencer (1988) suggests that the volcanic melange formed during an obduction stage and the calc-schist melange during an earlier subduction stage.

An "alpine-type" peridotite body, measuring 2.5 × 0.8 km, was reported from the amphibolites in the suture zone by Ahmed and Chaudhry (1976). Ghazanfar et al. (1991) have reported additional occurrences of ultramafic rocks in the suture here. These consist mostly of serpentinised wehrlite and subordinately of dunite, but there are large masses of coarse-grained pyroxenites further west. We regard all of these as arc-related cumulates belonging to the Sapat Mafic-Ultramafic Complex (Jan et al. 1993).

Burzil Pass–Dras Ophiolitic Melange: The Indus Suture zone in Dras area is made up of Mesozoic foreland deposits (Lamayuru unit) of the Indian continent in the south, Shergol Ophiolitic Melange in the middle, and Ladakh island arc volcanics and associated sediments of Cretaceous age in the north. There are at least two prominent melange zones to the north and south of Dras, but ophiolitic slices are also incorporated tectonically within the arc volcanics (Honegger et al. 1982). A similar picture emerges from the study of the geological map of Dras–Burzil Pass area to the southeast of Nanga Parbat (Casnedi 1976, Desio 1978). Sharma (1990, 1994) suggested that the Dras arc witnessed a major tectonic activity during the Late Cretaceous when slices of oceanic basement were incorporated into the arc near arc-trench interface. This was followed by metamorphism and Paleocene collision and uplift, causing southward thrusting of the ophiolitic slices onto India. In Spongtang, a large ultramafic thrust sheet occurs some 30 km to the south of the suture zone like the Skhakot–Qila Ophiolite in northern Pakistan. The melange contains serpentinised peridotites, dunites, gabbros, dolerites, basalts (some pillowed), Upper Jurassic radiolarian cherts, siliceous schists, mica schists, and gneisses. Intercalated pelagic foraminiferal limestone assign Upper Cretaceous age for part of the ophiolites in Pushkyum–Mulbekh section. The volcanic rocks have MORB-like geochemistry, but some represent tectonic slices of Dras (Frank et al. 1977, Honegger et al. 1982). Locally, the melange also contains imbricated slices of blueschists showing oceanic island or a transitional MORB-type tectonic setting. They were metamorphosed during the Middle Cretaceous at 350–420° C and 9–11 kbar (Honegger et al. 1989).

The Shyok Suture Melange

The Shyok Suture marks the closure between the Karakoram plate to the north and the Kohistan–Ladakh magmatic arcs to the south. Tectonic melanges and olistostromes are associated with the suture, but they have a subordinate proportion of ophiolitic components, raising the possibility that only a narrow stripe of ocean (or marginal basins) may have separated the Karakoram plate from the magmatic arcs. The melange is up to 3 km wide in the Shishi Valley in the southwest and reduces gradually only to a 150 m wide fault zone in Naz Bar (Chitral). But further east in Hunza it attains a thickness of 4 km.

In Chitral the melange consists of slate with interbedded conglomerate and sandstone, and blocks of volcanic greenstone, limestone, red shale and serpentine. It has a strong planar fabric defined by orientation of blocks and slaty cleavage. Clastics include sandstone, quartzite, limestone, volcanics and diorite. Blocks in the melange are of relatively few types. Limestones (some of which are Cretaceous in age and up to 200 × 50 m in size) and greenstones are the most common (Pudsey et al. 1985). A melange of more or less similar character occurs in Yasin and Ishkoman Valleys.

Near Chalt the melange in Hunza Valley consists of lenses of limestone, sandstone, conglomerate and mafic to ultramafic rocks in a matrix of chloritoid slate. Some blocks of sedimentary rocks are several kilometres in length. The basic rocks may be brecciated and there are greenschists rich in epidote, chlorite and actinolite. Schneider (1957) has also reported andesite. The ultramafic lenses consist of serpentinite, chlorite-carbonate, talc-chlorite-carbonate, and chlorite with magnetite octahedra. Fragments in the conglomerate may reach half a metre in length, some consisting only of limestone (Coward et al. 1982). Further east in the Rakaposhi area, greenstone may be quite abundant, judging from the fragments in the glaciers. No geochemical data are available, but at least the ultramafic fragments may have been derived from the Tethys or marginal/back-arc basins since the lithologies both to the north and south of the melange are impoverished in ultramafic components.

The petrographic and structural features of the ophiolitic melange in the Skardu–Khaplu area were described by Brookfield (1981). In the Skardu area, Zannetin (1964) reported serpentinites, chlorite schists (with or without epidote or actinolite), marbles, amphibolites and basalts. The basal contact of the melange here is a southwesterly directed thrust. Below this occur muscovite-chloritoid schists showing isoclinal folding with axial planes dipping steeply NE (related to the southward emplacement of the melange zone).

On the western side of Hushe Valley, a thick zone of north-dipping greenstones and greenschists with microdiorite are succeeded by a thick series of banded quartzofeldspathic schists and gneisses showing isoclinal folds, with axial planes dipping steeply southwest. On the northeastern side of the confluene of Hushe and Shyok Rivers, the section starts with a typical ophiolitic melange, containing fragments of serpentinite, pyroxenite, marble, red and black chert, porphyritic basalt, dacite, rhyolite and gabbro. These pass upwards into a metamorphosed distal flysch sequence in which the grade of metamorphism increases northwards, culminating in interlayered biotite-plagioclase gneisses and pelitic schists with thin marble layers and zones of serpentinite melange. Dips gradually change from NE to SW. Further north are thrust slices of dolomitic marble, pyroxenites and amphibolites and, finally, quartz diorite which is thrust steeply over the Karakoram Batholith to the north; but the thrust slices are cut by younger intrusions of the batholith. The melange shows two phases of deformation and metamorphism (Brookfield 1981).

Tirich Mir Ultramafics

In northwestern Chitral, some 30 km NW of the Shyok Suture, a series of lenticular and elongate igneous bodies follow the Tirich Fault for over 125 km. These consist of fine- to coarse-grained gabbro, which may be foliated and partially amphibolitised, and minor serpentinite. In the Garam Chashma area, one of these bodies is up to 2 km thick and 20 km long. It consists of medium- to coarse-grained gabbro and diorite, locally "altered" to very coarse hornblendite. The Tirich Fault juxtaposes low-grade metamorphic rocks of the Lutkho Formation against medium-grade rocks of Arkari Formation, both of which are probably Middle Paleozoic. According to Leake et al. (1989), it is possible that these basic and ultrabasic bodies were tectonically emplaced along the Tirich Fault, which would suggest that this is a more fundamental structure than had previously been recognised.

Deccan Trap, lower Sindh

The Deccan Traps may once have covered an area in excess of 1.5 million square kilometres in the Peninsular India and adjoining regions (Krishnan 1956). There is no doubt that the westernmost extension of the traps lies beyond the present course of the Indus River in southern Pakistan. Very limited information is available on the trap occurences in

Sindh, whereas those in India have been studied in detail from various angles by a very large number of scientists (cf., Subarao and Sukheswala 1981). Crookshank (1952) reported a 12 m thick basalt flow interstratified with sandstone of Upper Cretaceous age at Bor Hill. Some 35 km from Ranikot to Jakhmari, he reported another bed of basalt, 12 to 27 m thick. This overlies Cardita beaumonti (Danian to Maestrichtian) beds and underlies fresh water beds of Lower Ranikot (Thanetian) age. One of the flows can be traced for 35 km (HSC 1960). The bulk of the trap in Sindh, as in adjacent Kutch and Kathiawar, is regarded as subaerial.

The Burmah Oil Company bore holes in the area have revealed even greater thicknesses of the trap. At Lakhra three zones of basalts (14, 76 and 27 m thick) interbedded with sediments were found at the base of the Ranikot Formation. Near Thatta (24 km SE) atleast eleven flows of basalt ranging in thickness from 1 to 34 m and interbedded with tuff and clay were encountered. The presence of 130 m of volcanic rocks in bore holes 80 km apart cannot be taken as an isolated event. The Deccan Trap may have covered much of southern Sindh, and flows may be common under the younger cover. There are reports of basaltic flows in lower part of the Ranikot Formation in offshore drill cores (Shuaib 1976) and the Porali volcanics in the Mor Range have been considered to belong to the Deccan magmatism (Sarwar 1992). These may well be a westward continuation of those of Kutch.

There is a tight paleontological control over the age of the trap in Sindh, i.e., very Late Cretaceous, which is in harmony with the rest. The Deccan magmatism has been dated at 67 to 62.5 Ma, although some have suggested that it may have spanned a considerably long period, from about 100 to 30 Ma (Subarao and Sukheswala 1981, Verma 1991). Murthy (1981) stressed that the Deccan magmatism is one of a series of events of global dimension which require carefull study and synthesis. He attributes magma tectonics to cymatogeny and states, "the vulcanicity has been global, related to crust-mantle imbalance that came into being during the Mesozoic which is exogenically and endogenically manifested by high heat flow, eruptive activity, redistribution of land and sea, breakup of the Gondwana supercontinent, rise of mantle along crustal fractures, significant changes in the biota and perhaps first record of true morphogeny. The manifestations of this stupendous vulcanicity have been little understood.....".

The Chagai Magmatic Arc, Balochistan

The Chagai assemblage occurs in a broad plateau or arc bordered on the south by an abrupt, deep fore-arc basin. It consists of calc-alkaline magmatic, and sedimentary rocks which extend N-S for 150 km and E-W for 400 km in northwestren Balochistan of Pakistan and neighbouring Afghanistan and Iran. The magmatic belt appears to be a part of a several thousand kilometres long andesitic arc developed on the southern margins of Gondwanic microcontinental blocks in Iran and Afghanistan (for more details see Stocklin 1977, Lawrence et al. 1981). The Chagai arc is considered to have developed in response of northward subduction of the Arabian oceanic plate (Stonely 1974) or Neo-Tethys under the southern margin of the Dasht-i-Margo basement block as a continental margin of Andean-type (Sillitoe 1978, Sillitoe and Khan 1977), or as an intraoceanic island arc (Siddiqui et al. 1986, 1987).

Magmatic rocks: Most descriptions of the Chagai magmatic rocks are based on field observations and little geochemical and geochronology data are available to throw light on the petrological evolution of the arc. It appears, however, that the Chagai arc magmatism predominantly andesitic in character. This is a little strange because many island arcs are characterised by basaltic magmatism during the early stages of their growth. Whether this is related to the growth of the arc over a continental margin is not clear.

Volcanic rocks: The Chagai arc contains abundant flows, agglomerate, tuff and volcanic sand in different formations (see Chapter 5). Most of these have been regarded as andesitic, with subordinate basalt, dacite and rhyolite. But the Sinjrani Formation, which contains the main bulk of the arc volcanics, is metamorphosed and the volcanics are frequently replaced by albite, epidote and actinolite. Therefore, a correct evaluation regarding the preponderance of andesite should await detailed geochemistry. [In the central part of the belt, for example, the early stages of volcanism are basaltic and the later stages andesitic to dacitic (Siddique et al. 1986)]. The Sinjrani agglomerates occur in up to 16 m thick layers of jumbled and unsorted masses of volcanic debris. The fragments are up to a metre in size and consist of red to green and black porphyritic andesite. Ash and tuff in many colours are interlayered with coarse clastic, mostly as thin beds. The flows, brecciated to homogeneous, are 3 to 6 m thick andesite containing phenocrysts of augite, brown hornblende, plagioclase and, locally, orthopyroxene; but some are basaltic with olivine phenocrysts, quartz andesite or even more silicic (Vrendenberg 1901, HSC 1960). Some of the volcanics are mineralised. Sillitoe and Khan (1977) have reported Kuroko-type Zn-Cu massive sulfide and Manto-type mineralsation associated with a Late Cretaceous dacitic horizon 60 km NW of Saindak.

Plutonic rocks: These appear to consist predominantly of quartz diorites/tonalites, but subordinate amounts of mafic diorite, granodiorite, adamellite, granite, aplite and pegmatite have also been reported (Shcheglov 1969). They are divided into two groups: (1) the Chagai intrusions, and (2) the Shorkoh intrusions. The former occur mostly in the Chagai area close to Afghanistan, and the latter occur mostly in the western part of the arc in the Saindak–Robat region. There are numerous dyke swarms throughout the arc; these have been grouped with the Shorkoh intrusions (HSC 1960).

The Chagai intrusions comprise numerous bodies many of which are batholitic in dimentions (Photo. 43). The geological map of the Chagai region (Fig. 4.47) shows that they cover an area in excess of 2,000 km^2. They are considered as shallow-level intrusions, possibly joining into a huge batholith at depth (HSC 1960). There are many dyke-like offshoots in the country rocks. Preliminary petrography shows that they consist of hornblende-quartz diorite grading to diorite (with local orthopyroxene) and biotite granite, but along the southern margin of the Chagai Hills there is a greater range, including small pegmatites and aplites. Texturally they range from fine-grained biotite granite east of Manzil to coarse, micropegmatitic quartz diorite at Malik Noro. The large intrusions are porphyritic at margins. Granitic pebbles in conglomerate at the base of Maestrichtian Humai Formation suggest that at least some of these intrusions were emplaced during the Late Cretaceous (HSC 1960). In a 40 × 10 km area along the Afghan border to the northwest of Chagai, Siddiqui et al. (1986) noted three phases of intrusive activity: an early phase of composite hornblende gabbro-diorite stocks, a middle phase of large batholiths of adamellite and granodiorite, and a late phase of small tonalite intrusions. The petrography of these rocks and their associated volcanics has been described, together with major element analyses for 14 of them. The rocks plot as tholeiite on MFA diagram.

The intrusive rocks from western part of the arc have been described in some detail (Ahmed et al. 1972, Sillitoe and Khan 1977). These consist of several groups of different ages.

A group of very Late- or post-Cretaceous sills (with some lenticular masses and dykes), several hundred metres thick, form many high ridges in the Saindak area because of their resistant nature. Named as Tanki sills, they stretch NW-SE for some 55 km and can be used as a marker horizon (HSC 1960). They are made up of andesite porphyry containing abundant phenocrysts of plagioclase and augite in a very fine-grained to microlitic groundmass. The sills may be vesicular near the top, and columnar structure is well-developed at places. Chloritisation and saussuritisation have variably affected the rocks.

An earlier group of tonalites, quartz diorites and diorites was emplaced about 20 to 21 Ma ago (K-Ar hornblende and biotite ages: Sillitoe and Khan 1977), and intrudes the nearly horizontal beds of Amalaf Formation near Saindak. The intrusions form many small (<1 km^2) stocks and associated dykes and sills. The Saindak copper deposit is associated with three closely-spaced tonalite porphyry stocks around a parent stock (20 Ma age). These possess abundant phenocrysts of quartz, medium plagioclase, biotite, and up to 10 mm long hornblende, in a fine-grained groundmass of the same minerals and, locally, minor K-feldspar. The hornblende porphyries are commonly altered and contain secondary albite, epidote, chlorite ± calcite. There also are uniformly fine-grained quartz diorites containing andesine, hornblende, and 5–10% quartz. Around some of the stocks, there are several-kilometres broad zones of albite-epidote (locally hornblende) facies hornfelses (Ahmed et al. 1972).

A large number of dykes and sills of hornblende andesite porphyry are associated with the stocks but there also are a few quartz diorite porphyry dykes that post-date the early phase of copper mineralisation. Dykes of other compositions have been described: aplites cutting the tonalite porphyry predate weathering and mineralisation, and dacite porphyry containing quartz, hornblende and plagioclase emplaced during mineralisation. Very interesting in the Saindak area are local occurrences of near vertical dykes up to 12 m thick and trending NNE to N. To the north of the Saindak Fault, the swarms occur in a N-S zone measuring 2,500 × 250 m (Fig. 6.16: Sillitoe and Khan 1977). There appears to have been a long pause in the magmatic activity until Quaternary or, possibly, Late Pliocene, discussed at the end of this chapter.

The Spinatizha segment

West of the Chaman Transform Fault near Spinatizha (33°33'N:66°23'E), a terrane of crystalline rocks is exposed that links the Chagai arc with the Kandahar magmatic arc in southern Afghanistan. According to Lawrence et al. 1981, this area is an eastern continuation of the Chagai calc-alkaline terrane dragged by oroclinal flexing into the Chaman transform zone, and connects with the Kandahar arc to the north. However, such rocks have not been reported from elsewhere in the region. Four major rock units have been recognised in the Spinatizha area: (1) Spinatizha Metamorphic Complex, (2) Bazai Ghar Volcanics, (3) Khawaja Amran Intrusive Series, and (4) sedimentary sequence (Fig. 6.17).

The metamorphic complex includes orthogneiss, greenschist, amphibolite, metavolcanics, marble, and foliated muscovite granite. The volcanics include flows of andesite and basalt. The pyroclastic material is, however, more abundant. There are abrupt changes from rocks with well-preserved volcanic textures to those showing strong cataclastic foliation. Much of the deformation during metamorphism occurred along discrete deep-crustal shear zones. The foliation strikes subparallel to the Chaman Fault, and is vertical or steeply inclined, commonly dipping towards the fault. The volcanic rocks have been

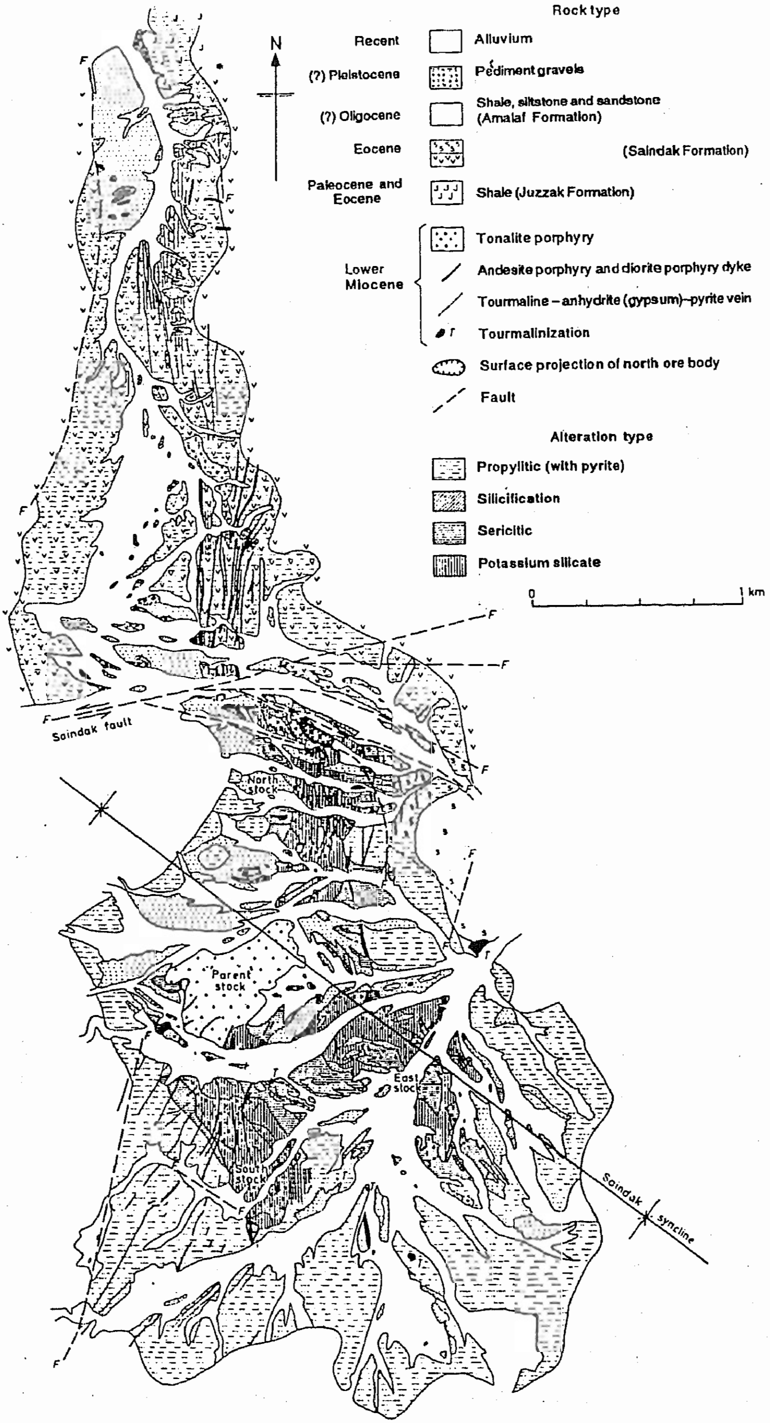


Fig. 6. 16. Geological map of the Saindak porphyry copper deposit (from Sillitoe and Khan 1979).

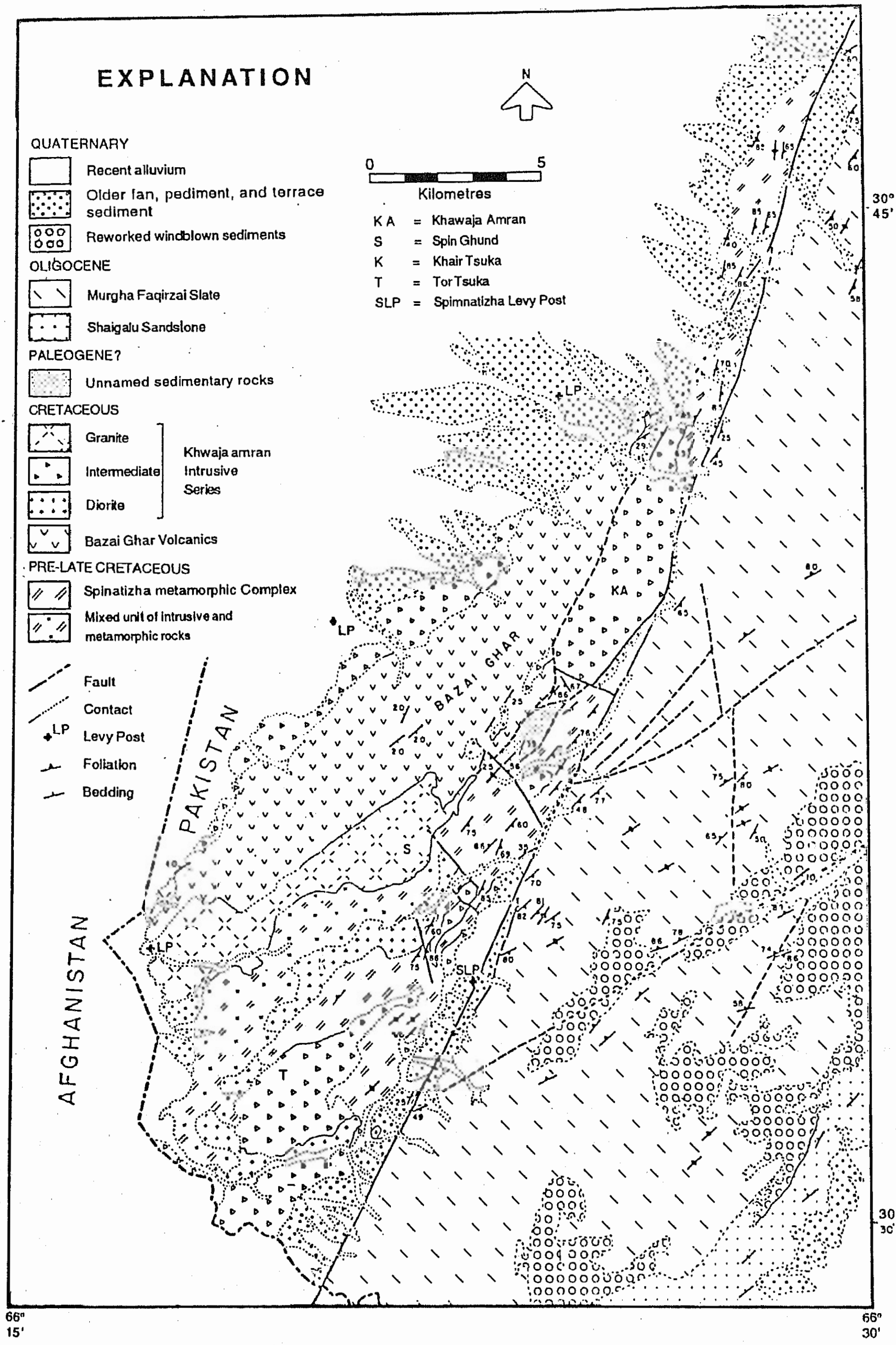


Fig. 6.17. Reconnaissance geological map of the Spinatizha area (from Lawrence et al. 1981).

metamorphosed in upper greenschist facies and consist of albite, actinolite, epidote, opaques ± chlorite ± biotite ± muscovite ± prehnite ± sphene ± calcite. The intrusive rocks are pre-tectonic and display cataclastic texures. They include granite, granodiorite and some diorite.

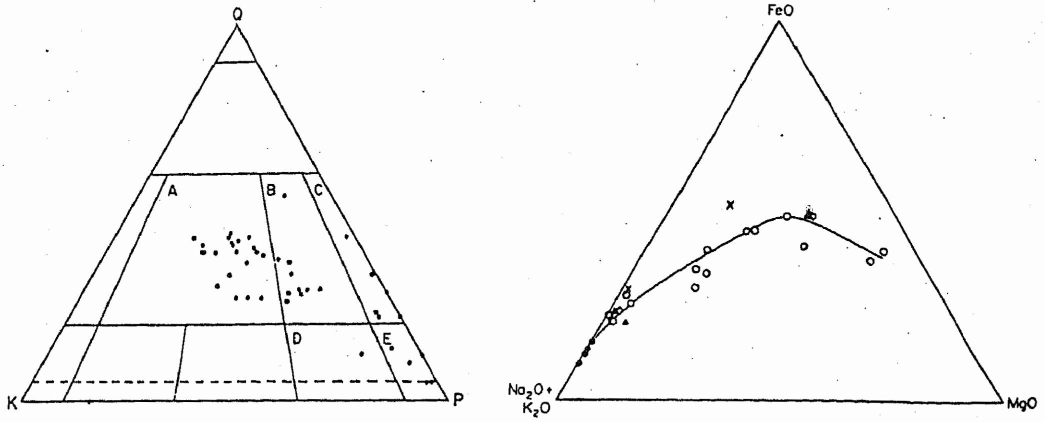


Fig. 6.18. (a) Modal compositions of the Khawaja Amran intrusive series. A–granite, B–granodiorite, C–tonalite, D–quartz monzodiorite, E–quartz diorite (IUGS classification 1973). (b) AFM diagram of chemical analyses from the Spinatizha area; solid circles= Spin Ghun granite, open circles= other rocks of the Khawaja Amran intrusive series, solid triangles= rocks of the Spinatizha metamorphic complex, crosses= Bazai Ghar Volcanics. Total iron as FeO. (From Lawrence et al. 1981).

The Bazai Ghar Volcanics, possibly younger than those of the metamorphic complex, are exposed as a single block along the western edge of the area mapped by Lawrence et al. (1981). These consist of flows, tuffs, breccias and ignimbrites, mostly andesitic. A three-fold subdivision may be possible: the bottom unit is volcaniclastic, the middle part mostly andesitic flows and flow breccia, and the top unit volcaniclastic, with a few interbedded andesite flows, and rare limestone and ignimbrite. These volcanics are also recrystallised: albite grows after plagioclase, chlorite and actinolite after the mafic minerals, and epidote is common in the matrix. These rocks can be correlated with the Sinjrani Volcanics of the Chagai arc, and those of the neighbouring Kandahar arc.

The Khawaja Amran Intrusive Series occurs in four large stocks and many smaller plutons. Modal analyses of 50 samples classify the rocks mostly as granite and granodiorite with some tonalite and quartz diorite, but 19 chemical analyses suggest that some gabbros or mafic diorites (SiO_2 45.8 to 51.8 wt%) and trondhjemite may also be present. The large plutons are composite, whereas the younger ones are more silicic, a common character of the Himalayan batholiths (Jan et al. 1981b). On MFA diagram (Fig. 6.18) the plots of the analyses display a typical calc-alkaline trend. Lawrence et al. (1981) suggest that they may represent a group of genetically related I-type rocks and may be an eastern extension of the Arghandab batholith in the Kandhar arc. This batholith and its satellites have radiometric ages of 100 to 110 Ma (Weippert et al. 1970, Tapponnier et al. 1981b, Afzali et al. 1979). In our opinion, the Khawaja Amran Intrusive Series may be younger because it has not been affected by the Early Tertiary tectono-metamorphic episode seen in the Chagai arc. The Khawaja Amran intrusions are cataclacised only along fault-related shear zones and are apparently free of the metamorphic overprint seen in the other two units of the Spinatizha area.

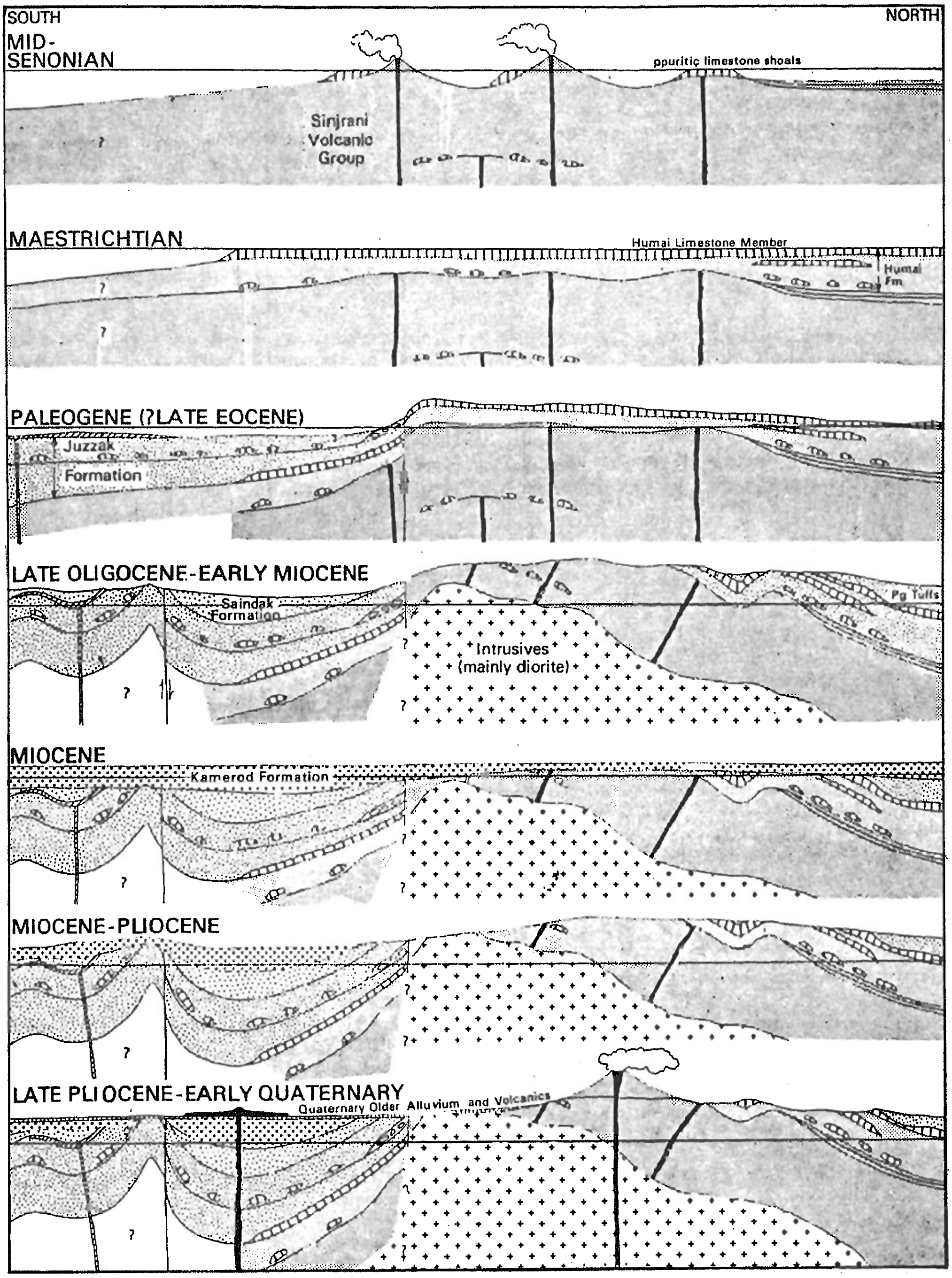


Fig. 6.19. Schematised sections across the Alamreg–Mashki Chah area (Chagai District), showing the interpreted geological evolution from Mid-Senonian times to the Quaternary. The principal volcanic and tectonic events are portrayed. Length of sections about 60 km. (After Arthurton et al. 1979).

The sedimantary sequence consists of coarse- and fine-grained clastic rocks (conglomerate and sandstone). Its relations to the other rocks are unknown, but the clastics contain volcanic, granitic and cherty material. Lawrence et al. (1981) thought that it could represent a small fragment of the fore-arc basin. The crystalline rocks, according to them, are an old portion (roots) of the Kandahar volcanic arc brought to the surface by vertical motion adjacent to this segment of the Chaman Fault.

Evolution of the Chagai Magmatic Arc

Field, petrographic, and limited geochemical data suggest that the rocks of the Chagai magmatic belt may represent an assemblage of an island arc proximal to continental margin. Most workers consider that the arc developed in response to northward subduction of an oceanic lithosphere underneath the Afghan microcontinental block. Initiation of magmatism in the arc corresponds in time with collision of the Afghan block with Asia in the Panjaw region of Afghanistan. Thus, collision in the north activated subduction in the south (Tapponnier et al. 1981b). Intermittent exposures of "andesitic" arc extend for many thousand kilometres from western Iran through southern Iran, Balochistan, Kandahar, Kohistan–Ladakh and, possibly, beyond in southern Tibet (Tapponnier et al. 1981a, Shackleton 1985). These coincide with, and lie to the north of, the assumed subduction zones beneath a series of Gondwanic microcontinents welded to Asia during the Middle Mesozoic. But the situation in Kohistan and Ladakh is far more complex; these may be a collage of accreted masses of oceanic regime, of mid-oceanic island arcs, Andean-type margins, and marginal/back-arc basins (Khan et al. 1993, 1995).

Several magmatic, tectonic, uplift and denudation episodes have been recognised in the Chagai arc since its initial growth during the pre-Maastrichtian (Table 4.9). An early major episode of deformation, uplift and erosion occurred approximately at the Cretaceous-Paleocene boundary (Fig. 6.19). This may be related to collision between the Chagai arc and the Ras Koh block. This may have resulted in 1) metamorphism of the volcanic arc, 2) obduction and uplift of the arc, and 3) emplacement of the Ras Koh ophiolite (fragments of the intervening oceanic crust or, less likely, lower levels of the arc itself). Another uplift episode occurred at 35 Ma (zircon fission-track data). This may be related to further obduction of the arc or to isostatic response of the arc to continued underthrusting.

Early in the Neogene, a major episode of deformation and uplift (20 Ma apatite fission-track data) established continental conditions lasting to the present day. Many calc-alkaline plutons were also emplaced at about 20 Ma (K-Ar hornblende, biotite ages). Yet another episode of important deformation occurred at about the Miocene-Pliocene boundary. The former deformation has been linked to the collision between the Lut and Afghan blocks, and the latter to collision between the Arabian and Iranian plates (Arthurton et al. 1979). Continued subduction resulted in renewed andesitic magmatism during the Late Pliocene-Pleistocene.

The Kohistan Magmatic Arc

The Kohistan and Ladakh terranes or tectonic zones (Desio, 1974) have been considered as intra-oceanic island arcs developed in response to northward subduction of the Neo-Tethyan oceanic lithosphere (Jan 1977, 1980, Tahirkheli et al. 1979, Klootwijk et al, 1979, Thakur and Sharma 1983). It appears that the arcs were welded to the Karakoram plate 85 to 95 Ma ago, and thereafter they became an Andean-type margin before collision with India 55 to 65 Ma ago (Pettersan and Windley 1985, Powell 1979, Patriat and Achache 1984, Klootwijk et al. 1992).

The Kohistan tectonic zone in northern Pakistan is characterised by a complex interplay of magmatism, deformation, metamorphism, uplift and erosion (Tahirkheli and Jan 1979, Bard et al. 1980, Butt et al. 1980, Bard 1982a, 1982b, Coward et al. 1982, 1986, Windley et al. 1985, Treloar et al. 1989b, Khan and Coward 1990). It is made up of a variety of Cretaceous and Tertiary igneous and subordinate sedimentary rocks which can be grouped in several distinct units on the basis of age and lithological characteristics. Each of these units extends for tens to hundreds of kilometres along the E-W length of the arc (Fig. 6.20). Here we present the principal petrological aspects of these units which show the following rough disposition from north to south: Yasin Group, Chalt (and other) Volcanics, Kohistan Batholith, Jaglot Group, Chilas Complex, Southern Amphibolites, Jijal Complex.

Volcanic sedimentary and magmatic sequences

Yasin Group: Described in considerable detail by Ivanac et al. (1956), Desio (1963), Matsushita and Huzita (1965), Tahirkheli (1979, 1982) and Pudsey and others (1985, 1986), this group consists of sedimentary and volcanic rocks forming a long and narrow belt immediately to the south of the Shyok Suture. Locally, blocks of these rocks also occur within the Shyok Suture Melange (? olistostrome). In the type locality, Pudsey (1986) has divided the group into: (a) upper 2,000 m of red, purple, green and grey shales with interbedded greenstones, probably tuffs, (b) middle 500 m of grey slates with some distal turbidites, and (c) lower >500 m volcanic-lithic conglomerate and sandstones, tuffs, slates and rudist limestones. The group contains Aptian-Albian fossils (see Chapter 5).

Lateral lithological variations are common and some sections are devoid of the volcanic components. The Hunza section contains mostly clastics and volcaniclastics in the form of distal turbiditic slates with massive greywacke. The Ishkoman section is devoid of carbonates and consists of slates with some quartzite coarsening to pebbly conglomerate towards the Shyok Suture. In Chitral, the equivalent Purit Formation (< 1 km thick) comprises red calcareous shales overlying limestone. On the northeastern edge of the Nanga Parbat–Haramosh Massif, arenaceous slates, conglomeratic schists, conglomerates, phyllite and Pakora limestone of the Turmik Formation are the possible easterly equivalent of the Yasin Group (Desio 1963, 1964, Zanettin 1964, Desio et al. 1985, Le Fort et al. 1995). The group may have been deposited in an intra-arc and/or back-arc basin (Pudsey 1986, T. Khan et al. 1995).

Chalt Volcanics: Underlying and locally interbedded with the Yasin Group, these volcanics form an arcuate belt more than 350 km in length. They have been referred to as the Greenstone Complex (Ivanac et al. 1956), Turmik Formation (Desio 1963, 1964, Desio et al. 1985, Zanettin, 1964), Rakaposhi Volcanic Group (Tahirkheli 1982), and Chalt Volcanic Group (Petterson et al, 1990a). The volcanics comprise a diverse group of rocks. They range from basalt to andesite, dacite and rhyolite flows, conglomerate, breccia and tuff. They exhibit low grade metamorphism characterised by chlorite, epidote, amphibole, sodic plagioclase, quartz, with or without carbonate, iron oxide and mica. These minerals (± zeolite?) also occur in amygdules and veins; quartz veins are common in some places. Modal variations result in a variety of colours: green, grey, brown, buff and white, locally mottled.

There appears a regional variation in the composition of the volcanics. Those of the eastern part (Chalt, Jaglot, Turmick) consist mainly of metmorphosed flows and tuffs of

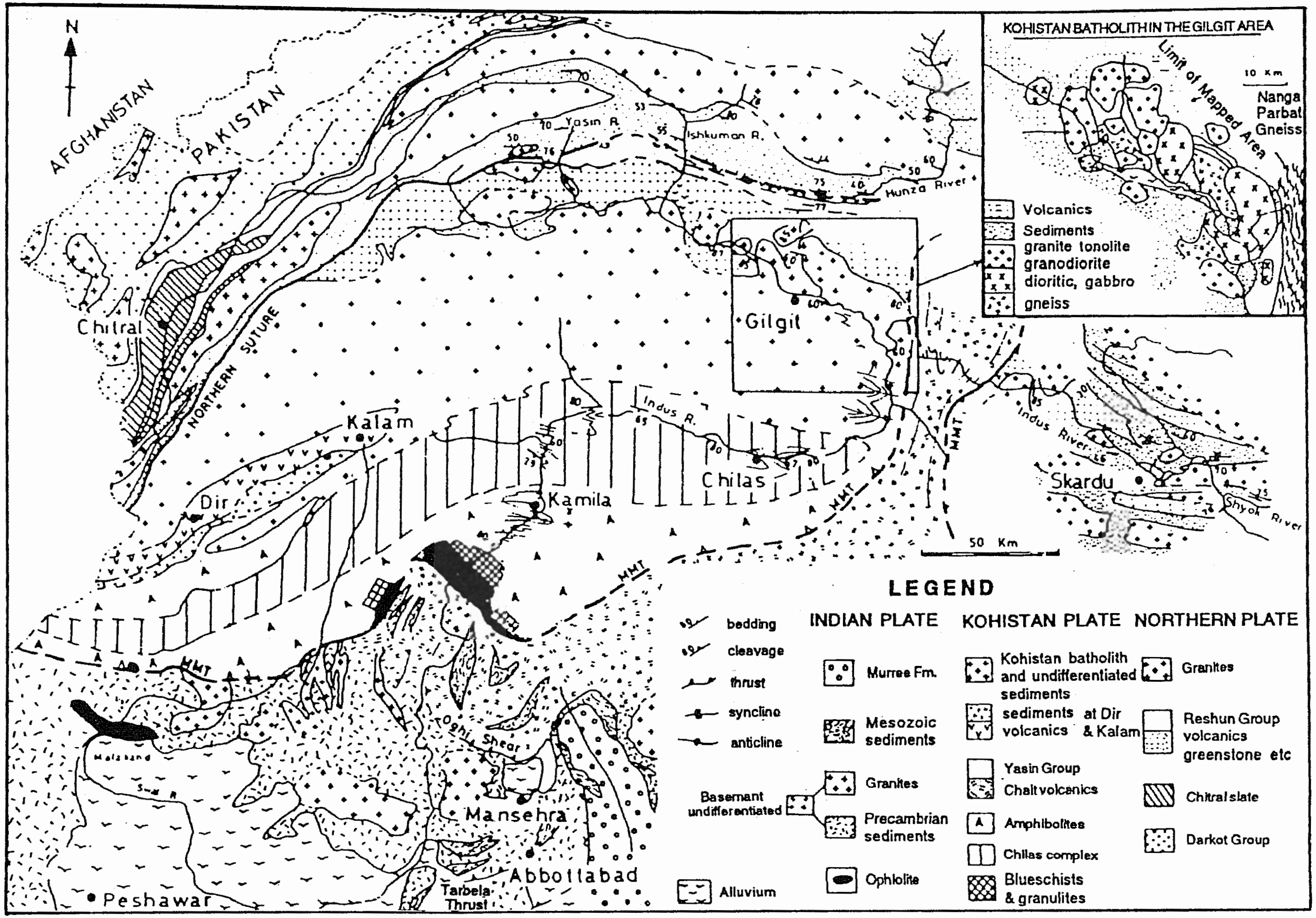


Fig. 6.20. Lithological and structural map of Kohistan. Inset shows details of plutons in the Trans-Himalayan Kohistan Batholith near Gilgit. MMT= Main Mantle Thrust/Indus suture, Northern suture=Shyok suture. (From Coward et al. 1986, Tahirkheli and Jan 1979).

basaltic to andesitic composition with subordinate dacite and rhyolite. The tuffs may be rich in crystals while the flows are massive to pillowed (Photo. 44). The pillows are amygdaloidal, locally enriched in epidote, and their cores are coarser grained than crusts and matrix. They are commonly deformed into elongated lenses. Copper mineralisation is locally seen in the volcanics. The proportion of volcaniclastics increases westwards where hornblende-pyroxene-rich tuffs are interbedded with basaltic-andesitic lavas between Gilgit and Ishkoman (Petterson et al. 1990a). A high proportion of volcaniclastics and silicic lavas occurs in Yasin and further west (Ivanac et al. 1956). The eastern volcanics are strongly deformed and show tight isoclinal folding.

Preliminary geochemistry suggests that the Chalt Volcanic Group can be divided into two distinct types (Petterson et al. 1990a). The eastern (Chalt proper) volcanics have a tholeiitic trend and high MgO (6–15 wt%), Cr and Ni contents with compositions typical of boninites and, in extreme cases, basaltic komatiites (Cameron et al. 1979). They have negative Nb anomalies, low concentrations of incompatible trace elements such as Ce, Y, Zr, Ti, Rb and K, and are similar to low-K tholeiites (Gill 1981). The western volcanics display a typical calc-alkaline (basalt-andesite-rhyolite) trend of differentiation. They are enriched in low field-strength (Rb, Ba, Sr, K) and light rare earth elements (e.g., Ce) relative to high field-strength and heavy rare earth elements (e.g., Zr, Ti, Y). Figure 6.21 shows the chemistry of the two types of volcanics.

We are of the view that the high Mg-type eastern volcanics were probably erupted before the deposition of the Yasin Group (Aptian-Albian) in a back-arc basin (T. Khan and others, 1993, 1995). The western volcanics were possibly generated in an Andean-type margin at a later date. Petrographically, these are similar to the Late Paleocene-Early Eocene Shamran and Utror Volcanics discussed in the following.

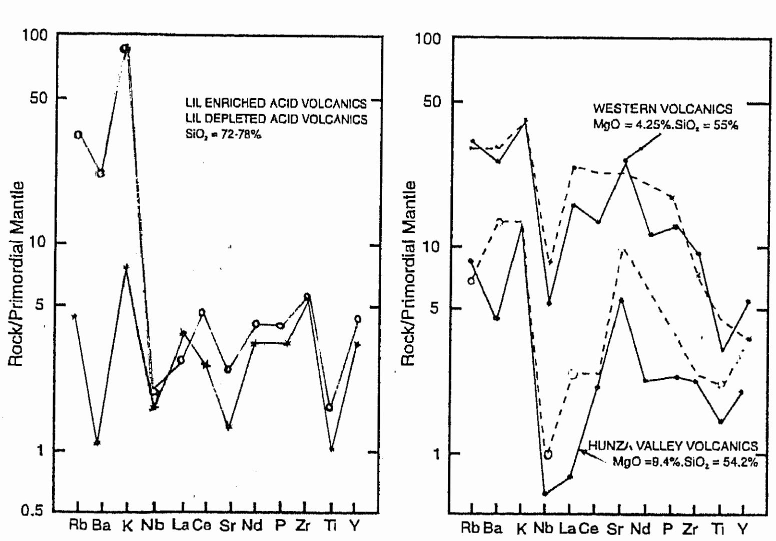


Fig. 6.21. Spidergram for high-Mg and calc-alkaline volcanics. Circle in right-side represents average low-K tholeiitic andesite.(From Gill 1981. Petterson et al. 1990a.

Other volcanic occurrences in Kohistan: Several more occurrences of volcanic rocks have been recorded in the Kohistan magmatic arc, e.g., Swat–Dir–Chitral, Thak Valley, Skardu–Deosai. Although they do not fit into the north to south traverse being followed here, it will be appropriate to mention these for the sake of continuity and comparison with the Chalt Volcanic Group.

Utror, Drosh and Shamran Volcanics: The Utror Volcanics comprise over 3,000 m thick red, green, grey, white, at places mottled, rocks. They form a NE-trending belt extending from eastern Afghanistan through Bajaur, Jandul, Shringal and Utror, to Paloga and beyond (Jan and Mian 1971, Kakar et al. 1971). The volcanic activity was predominantly explosive and fragmental rocks (breccia, agglomerate, tuff) are much more common than flows; there are ignimbrites reminiscent of glowing avalanches. The volcanic stratigraphy is complex and eruptive centres have not been located; however, there are some feeder dykes and at least three volcanic necks, one to two kilometre across, near Afghan border. Sullivan et al. (1993) reported that the volcanics accumulated in a predominantly subaerial ring-plain or flanking facies distal to the main focuss of volcanic activity.

The Utror Volcanics are predominantly silicic with up to 79 wt. % SiO_2; but andesite and basalt are common in Dir. They have been affected by low-grade metamorphism leading to the growth of chlorite, epidote, iron oxide and micaceous minerals in many rocks. Some are quite fresh, including rare welded tuffs and perlite. Geochemical studies (Majid and Paracha 1980, Majid et al. 1981, Hamidullah et al. 1990, Sullivan et al. 1993, Shah et al. 1994) confirm the calc-alkaline character of the rocks. Kakar et al. (1971) and J. Khan (1979) reported the occurrence of *Lockhartia sp.* and *Discocyclina/Aktinocyclina sp.* in arenaceous and calcareous rocks sparingly interbedded with the volcanics. Treloar et al. (1989) reported an ^{40}Ar-^{39}Ar hornblende age of 55 ± 2 Ma for a basaltic andesite, confirming Late Paleocene-Early Eocene age for the volcanics. They show a thrust contact with the underlying Late Cretaceous/Paleocene sedimentary rocks consisting of basal conglomerate passing upward into deep-water turbiditic sandstone and siltstone with rare limestone, deposited in a fault-controlled rapidly subsiding, elongate basin within the magmatic arc (Sullivan et al. 1993). The sediments overlie the arc basement and, alongwith the Utror Volcanics, have been intruded by younger plutons of the Kohistan Batholith (Jan and Mian 1971).

The Drosh Volcanics occur in a thin belt to the south of the Shyok Suture in southwestern Chitral. They are well-exposed on the road south of Drosh and in the lower Shishi Valley. Overlying the Purit Formation, they are a sequence of thickly bedded porphyritic andesites with phenocrysts of plagioclase, hornblende and pyroxene. Some are less phyric and contain only small plagioclase phenocrysts. There are thin, red shales interbedded with the volcanics. The lavas are strongly epidotised in many places; some are highly vesicular and a few are autobrecciated. Copper mineralisation is locally associated with them.

In the section to the north of Lowari Pass, a series of steeply dipping basic to acidic metavolcanics (greenschist to ?amphibolite facies), with intercalating metasediments including marbles, is intruded by a number of foliated diorite bodies 2–3 km across. These rocks have a linear fabric and are cut by deformed mafic dykes. Further north are greenschist facies tuffaceous rocks with some andesite lavas. Intruded by small diorite and granodiorite bodies these are unconformably overlain by the Purit Formation (Yasin Group). The stratigraphic position and presence of deformed plutons suggests that these particular volcanics may be the equivalent of the Chalt Volcanics. Immediately south of the Shyok Suture in the Shandur Pass area are amphibolitic gneisses with some mica-rich metasedimentary bands. These are similar to, but rather more migmatitic than, the metavolcanics north of Lowari Pass. These amphibolites were metamorphosed prior to injection of tonalitic veins, muscovite pegmatite sheets and, lastly, some undeformed dykes.

Photo. 44.

Photo. 45.

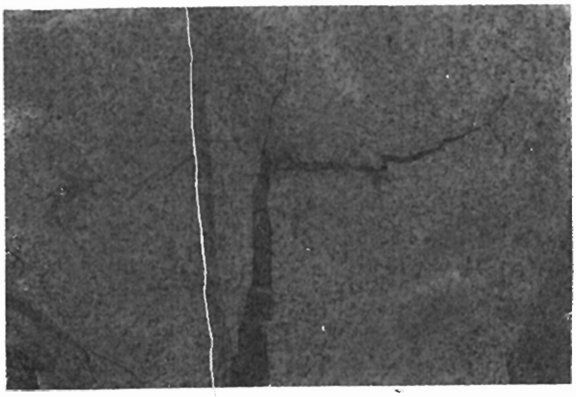


Photo. 46.

Photo. 47.

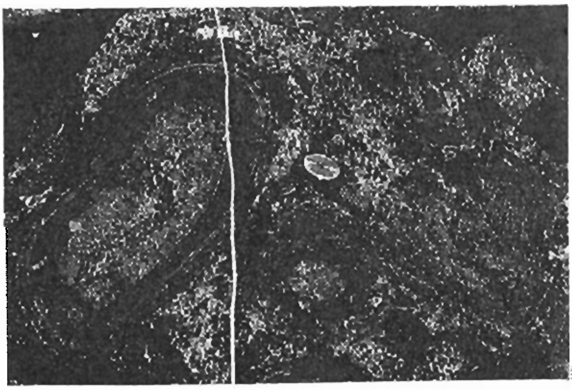


Photo. 48.

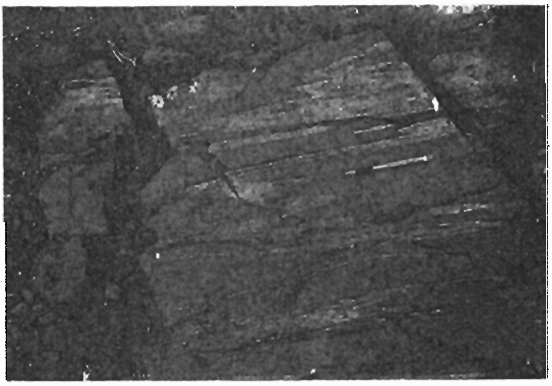


Photo. 49.

Photo. 44. Deformed and metamorphosed pillow lavas intruded by a granite dyke, in a road cut south of Chalt. (Photo: *M. Q. Jan*).

Photo. 46. Kohistan Batholith south of Matum Das. 102 Ma trondhjemite cut by Late Cretaceous mafic dyke. (Photo: *M. Q. Jan*).

Photo. 48. Large orbicules, some containing more than a dozen concentric shells, in a gabbro-norite (Kohistan Batholith) near Deshai, Swat Kohistan. (Photo: *M. Q. Jan*).

Photo. 45. Kohistan Batholith near confluence of Indus and Gilgit Rivers. Metavolcanics (amphibolites) cut by composite aplite-pegmatite dykes. (Photo: *M. Q. Jan*).

Photo. 47. Network of granitic veins in dioritic rock (Kohistan Batholith), near confluence of Indus and Gilgit Rivers. (Photo: *M. Q. Jan*).

Photo. 49. Isoclinally folded granitic layers in amphibolite (metavolcanic) host, Sai Nala. (Photo: *M. Q. Jan*).

Photo. 50.

Photo. 51.

Photo. 52.

Photo. 53.

Photo. 54.

Photo. 55.

Photo. 50. Peridotite of the Chilas complex with a layer of cumulus chromite, NE of Chilas. (Photo: *M. Q. Jan*).

Photo. 52. Graded anorthosite-gabbro showing syndepositionsl faulting. On the left there is a faulted block with pegmatite but no graded layers. Chilas complex west of Thak–Indus confluence. (Photo: *M. Q. Jan*).

Photo. 54. Deformed layers of serpentinised dunite in diopsidite of the Jijal Complex, north of Jijal town. (Photo: *W. Hamilton*).

Photo. 51. Gabbronorite boulder containing peridotite which shows both concordant and discordant relations. From Chilas complex in Buto Gah, south of Chilas. (Photo: *M. Q. Jan*).

Photo. 53. Broken blocks showing rhythmic layers, grading from ultramafic to anorthositic composition, in an anorthosite matrix. Chilas complex, Buto Gah. (Photo: *Ian Davison*).

Photo. 55. Garnet granulites of the Jijal complex showing layers rich in garnet and/or diopside, 4 km north of Jijal. (Photo: *M. Q. Jan*).

To the east of Shandur Pass there are gently northward-dipping, low-grade metavolcanics termed as the Shamran Volcanic Group (Pudsey et al. 1985). These share many characterisitics with the Utror Volcanics including a high quality Ar-Ar plateau age of 58 ± 1 Ma on a hornblende-bearing basaltic andesite (Treloar et al. 1989). The Shamran sequence contains flows of basalt, andesite, and abundant rhyolite. Explosive activity is reflected in abundant fragmental rocks (breccia, agglomerate, tuff) and ignimbrite. Shaly and sandy sediments occur locally. Most volcanics are epidotised and greenish grey, but some are reddish to purple due to iron oxide. Like the rest, these are also cut by younger plutons and andesitic dykes.

Southeastern Kohistan Volcanics: Lithologies in the hanging wall of the Indus Suture in southeastern Kohistan belong to the Sapat Mafic-Ultramafic Complex (Jan et al. 1993). Between the latter and the Chilas Complex, the terrane is occupied by E to W-trending stripes of volcanic and plutonic (diorite to granite) rocks (Ahmed and Chaudhry 1976, Khan and Thirlwall 1988, Ghazanfar et al. 1991). The volcanic stripes, named differently by different authors, range from mafic to intermediate in composition and locally contain small bodies of gabbro, tonalite, granodiorite and granite. The volcanics are locally deformed internally and may display foliation. Pillow structures are preserved in some andesitic(?) lavas near Dalpur (Buto Gah).

Although some of the volcanic rocks appear to be fresh and contain phenocrysts of feldspar and hornblende, most have been affected by low-grade metamorphism and consist of actinolite, epidote, chlorite, albite, quartz and sphene. Some of the amphibolites (containing hornblende, plagioclase ± quartz ± garnet) may be higher grade equivalents locally developed. The southeastern Kohistan metavolcanics range from green to grey basalt to andesite and are intruded by dykes of porphyritic dacite containing garnet. Geochemically, the volcanics can be divided into two distinct groups (Khan and Thirlwall 1988, Khan pers. comm.). The Niat Volcanics display ocean-floor chemistry, whereas the Sumal Volcanics are calc-alkaline and island-arc type. The former may, thus represent the foundation (oceanic crust) for the growth of the island arc partly represented by the Sumal Volcanics.

Skardu–Deosai Volcanics: The Skardu area of Ladakh–Kohistan arc has been investigated by Desio (1963, 1964, 1978), Casnedi and Ebblin (1977) and Hanson (1989). Three major formations, intruded by a range of granitic and mafic rocks, have been described. The Bauma–Harel Formaton in the north-eastern and northwestern part of the area adjacent to the Shyok Suture consists of volcaniclastic metasediments (chlorite-epidote greenschist), interbedded with slates, phyllites, minor calcareous rocks and multicoloured conglomerates. Containing poorly preserved Cretaceous fossils, it has been correlated with Chalt Volcanics and Gawuch Formation further west. Coeval with, or younger than, this formation is the Katzarah Formation covering large areas to its south. This unit is represented by complexly folded high-grade (up to sillimanite-K-feldspar zone) gneisses, calc-silcates, marble and metavolcanic amphibolites. To the south of Skardu occur the Burji Formation phyllite, slate, chlorite-epidote metavolcanic and limestone containing upper Cretaceous fossils.

The Askore amphibolites of the Katzarah Formation occurring on the eastern flank of Nanga Parbat consist of garnet amphibolites and amphibole (± biotite ± epidote) gneisses. Southwards they become more quartzofeldspathic with some quartzopelitic horizons. The conglomeratic layers contain amphibolitised pyroxenite, quartzfeldspathic and amphibolitic blocks. In the Askobar Valley, the amphibolites have an overlying cover of pink quartzite and white marble. The Askore amphibolites have been considered by Le Fort et al. (1995) to be derived from Dras Volcanics (probably Dras-I volcanics: Reuber 1989, Robertson and Degnan 1994). The Turmik Formation in this area consists of arenaceous slates (± epidote ± chlorite), conglomeratic schists and conglomerates containing clasts of acid to intermediate volcanics, amphibolite, marble, quartzite, minor serpentinite, and numerous limestone horizons. The formation is considered an equivalent of the Chalt Volcanics (Le Fort et al. 1995).

In the Deosai Plateau to the south, large areas of volcanic rocks, containing ultramafic masses and intruded by abundant granitic rocks have been reported. These consist of flows, agglomerates, tuffs, and intercalated sediments with Cretaceous (Orbitolina) limestone and slate (Wadia 1937). The volcanics range from basalt and andesite to rhyolite, metamorphosed under greenschist facies. A 125.4 ± 6 Ma Ar-Ar hornblende age on an andesite (Hamidullah et al. 1992) confirms the previously suggested Late Jurassic-Early Cretaceous age for these volcanics which are a continuation of those of Dras (Frank et al. 1977, Honegger et al. 1982, Dietrich et al. 1983, Sharma 1990b).

Kohistan Batholith: This is a principal unit of the Kohistan magmatic arc and constitutes a 300 km long and up to 60 km broad belt to the west of Nanga Parbat. It forms the western part of the great Transhimalayan Batholith (Gansser 1964) and continues on the eastern side of Nanga Parbat in Deosai, Ladakh and beyond (Desio 1964, Brookfield and Reynolds 1981, Sharma 1990c, Thakur 1992). The Kohistan Batholith is composite and consists of numerous large to small plutons, plugs, dykes and sheets (see inset in Fig. 6.20) emplaced over a time span of some 75 million years. A wide range of rocks has been reported to constitute the batholith: gabbros, hornblendite, diorites, quartz diorite, adamellite, granodiorite, granite, tonalite, trondhjemite, aplite, pegmatite, etc (Photos. 45 and 46). Locally, as many as five pulses of intrusions can be observed within a few tens of metres, commonly showing a decrease in mafic constituents with time (Photo. 47). Multiple intrusive activity is best seen in the Indus Valley section where dykes and sheets of leucogranites, <1 to 10 m thick, occur in swarms and networks, locally making up to a third of the exposures.

Depending upon age, composition, cooling history, volatile content, depth of emplacement, size, etc., the intrusions display a variety of textures and structures: fine-grained to pegmatitic, idiomorphic to allotriomorphic, non-porphyritic, equigranular to megacrystic, undeformed to strongly deformed (foliated, gneissose, or banded) and unmetamorphosed to metamorphosed. Some of the composite aplite-pegmatite bodies to the east of Gilgit display igneous lamination/layering. Orbicular structures and comb-layers occur in a few places (Photo. 48). Those of the Swat Kohistan are spectacular with some orbicules over 30 cm in length and consisting of more than a dozen shells (Jan and Mian 1971, Symes et al. 1977).

Space does not permit to go into the petrographic details of this huge batholith. Interested readers may consult Jan et al. (1981b), Shams (1983), and Petterson et al. (1990b) for regional perspective. Local descriptions have been given by Badshah (1979) for Bajaur ; Chaudhry et al. (1974) and Butt et al. (1980) for Dir ; Jan and Mian (1971), Khalil and Afridi (1979) and Jan and Asif (1983) for Swat ; Ivanac et al. (1956), Casnedi et al. (1978), Blasi et al. (1980), Petterson (1984), George et al. (1993) and Khan (1994) for Gilgit ; Wadia (1937), Zanettin (1964), Desio (1963), Hanson (1989) and Reynolds and Brookfield (1983) for the area east of Nanga Parbat.

Following preliminary geochemistry by Blasi et al. (1980), Majid (1979), and Jan and Asif 1983), the Kohistan Batholith in the Gilgit area was studied in detail by Petterson and Windley (1985, 1986, 1991), Petterson et al. (1990b, 1993), and Khan (1994). On the basis of field relations, petrology and geochronology, they divide the batholith into three formative stages:

i) An early, bi-modal suite of gabbro-diorites and trondhjemite formed within an island arc setting between 110 and 85 Ma; e.g., Matum Das and, possibly, Deshai (Swat) and Lowari (Dir).
ii) The main bulk of the batholith formed within an Andean-type plate margin between 85 and 40 Ma. This stage comprises an earlier gabbro-diorite suite (85 to 60 Ma) and a later granitic suite (60 to 40 Ma) (Treloar et al. 1989); e.g., Gilgit, Shirot, and possibly Laikot, Donger (Swat) and Warai (Dir).
iii) Leucogranitic sills and dykes within a post-collisional setting (30 to 26 Ma with one age of 23 ± 1 Ma: Petterson 1984, George et al. 1993); these include the Indus-Gilgit confluence and Pari acid sheets.

Most rocks of the batholith display a typical calc-alkaline trend on MFA diagram (Fig. 6.22), but, some rocks are tholeiitic, sub-alkaline or even alkaline (Blasi et al. 1980, Debon et al. 1987, Le Fort et al. 1995). Major, trace, and rare-earth element geochemistry and some Sr and Nd isotopic data have been presented for the Matum Das (Hunza Valley), Gindai (Yasin) and Indus-Gilgit confluence plutons by Petterson et al. (op. cit.), Debon et al. (1987) and George et al. (1993). Trace element and REE patterns for selected rocks are shown in Figure 6.23. The batholith has a fairly complex petrogenetic history. The earliest phases in the batholith were either derived from a juvenile metavolcanic crust or modified mantle wedge above the subduction zone. The early post-suturing gabbro-diorites have a dominantly mantle-derived isotopic signature. With time, however, there is evidence for an increasing crustal input as the immature crust of the lower arc began to melt. The Pari acid sheets (typical garnet-muscovite leucogranites) have distinctly lower ENd and higher ESr, suggesting derivation from a metasedimentary source with a high Rb/Sr ratio and a crustal residence time of about 70 Ma.

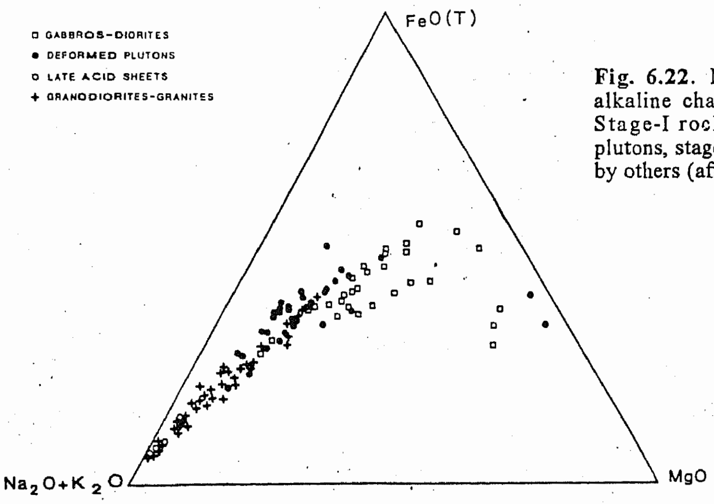


Fig. 6.22. MFA diagram showing the calc-alkaline character of the Kohistan Batholith. Stage-I rocks are represented by deformed plutons, stage III by late acid sheets, and stage II by others (after Petterson and Windley 1986).

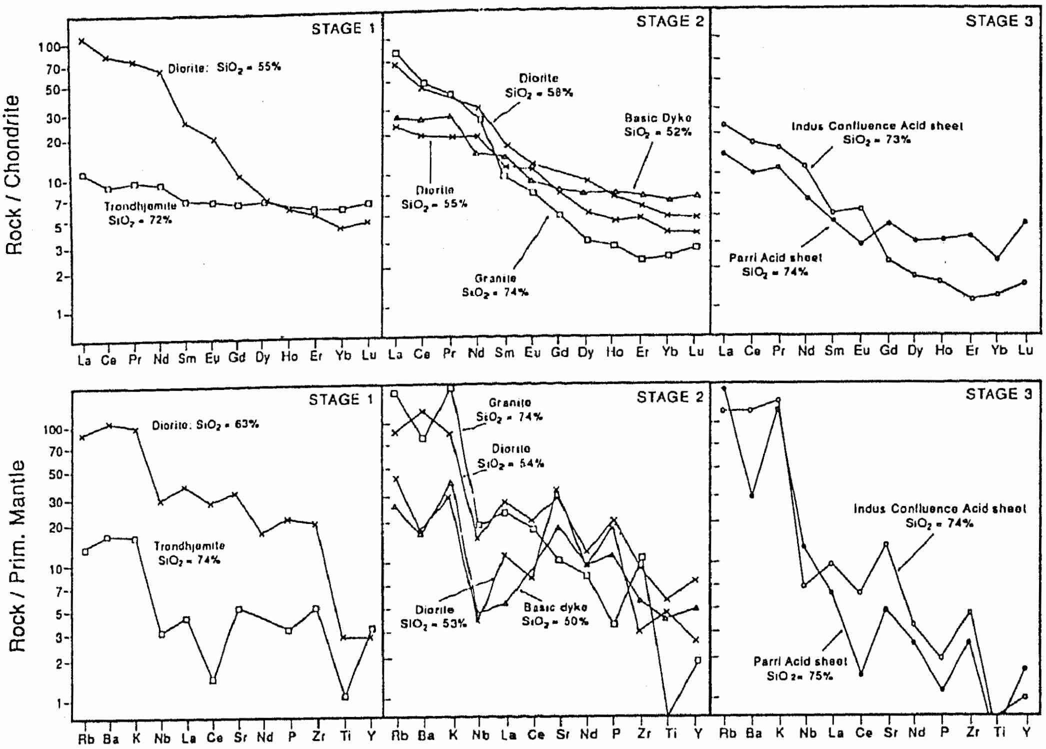


Fig. 6.23. Spidergrams and chondrite-normalised rare-earth element diagrams for the three stages of the Kohistan batholith. Crosses–gabbros and diorites; squares–felsic plutons (over 68–70% SiO_2); triangles–stage 2 basic dykes (from Petterson and Windley 1986).

Mafic dykes: The Kohistan Batholith is intruded by low-medium-K basalt, basaltic andesite and trachyandesite dykes, up to 4 m thick and locally forming swarms. These also cut the Chalt Volcanics, Jaglot Group rocks, and the fold strucutures and penetrative fabric associated with the collision between the Kohistan arc and the Karakoram plate. Orientation data along the Hunza River suggest they intruded in response to NW-SE to N-S tension. Petterson and Windley (1992) report a 75 Ma average Ar-Ar age; but Khan et al. (1992) classified them into two age groups, the first group being metamorphosed to amphibolites and the second group not showing metamorphism. Petrographic and geochemical details of the dykes have been presented by Petterson and Windley (1992) and Khan et al. (1992). They are considered to have formed by partial melting of a mantle source metasomatised by subduction-related processes.

Jaglot Schist Group: The Yasin sediments and Chalt Volcanics are repeated to the south by the Gilgit anticline and Jaglot syncline (Photo. 49). In the Sai Nala and adjacent valleys, T. Khan et al. (1994, 1995) noted that the Chalt Volcanics overlie a more than one kilometre thick sequence of metasediments and interbedded metavolcanics that they termed as the Gilgit Group. Because of the volcanic component, Treloar et al. (1996)

prefer to call them as the Jaglot Schist Group. The sedimentary rocks in the group consist of psammites, pelites, and calc-silicates that may be thin-bedded and graded-bedded. The interbedded volcanics may be pillowed or non-pillowed massive basalts. Both groups of rocks show strong deformation; some pillows are stretched >10:1 and look like banded amphibolites (Jan 1988). Most of the interbedded volcanics are only up to a few metres thick.

The Jaglot-type metasediments have been reported in Dir and Swat areas also (the Kalam Group), and Treloar et al. (1996) suggest that the group may form a thin band extending right across the Kohistan arc. The rocks occur to the north of the Chilas Complex which intrudes them. On their north the metasediments are intruded by the Kohistan Batholith. The Jaglot Schist Group shows metamorphism from biotite to sillimanite and K-feldspar grade. The calc-silicate rocks contain diopside, tremolite ± garnet, and the volcanics are amphibolitised. Evidence of partial melting and migmatisation is seen in K-feldspar schist/ gneisses near Jaglot, Sai Nala and south of Kalam. Some of the Sai Nala gneisses look like Nanga Parbat-type remobilised basement gneisses. But they seem to conformably underlie the Chalt Volcanics and are probably Early Cretaceous. The sediments are similar to graywacke dominated lithologies (i.e., turbidites) deposited in a medium to deep water marine (back-arc) basin.

Chilas Complex: The Chilas Complex is a large body of mafic-ultramafic rocks extending from Nanga Parbat to eastern Afghanistan. Similar rocks also occur on the eastern flank of Nanga Parbat (Misch 1949) and in Kargil, Ladakh (Rai and Pande 1983). This Complex has been studied in some detail, at least petrographically (Jan et al. 1984, Shams 1975, Khan et al. 1989, 1993, Khan and Jan 1992). Much of the Complex is made up of gabbronorites (mostly uniform with local layers of pyroxenites and anorthosite) and subordinate quartz-two pyroxene diorites. These have been referred to as the main gabbronorite association and contain abundant labradorite or andesine. Ductile shearing results in local high-temperature recrystallisation (Jan and Howie 1980), but the rocks have also been hydrated to amphibolites along lower temperature shear zones.

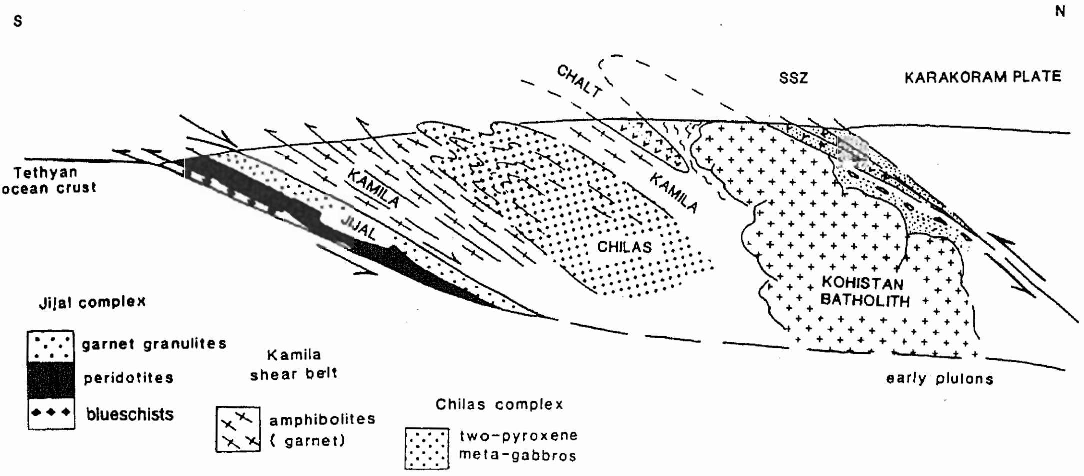


Fig. 6.24. Schematic cross-section of Kohistan at around 80 Ma, after collision of Kohistan with the Karakoram plate and before collision with Indian plate (from Treloar et al. 1989b, Searle 1991).

The main gabbronorites contain several small bodies (commonly less than 5 km^2 in area) of ultramafic-mafic-anorthositic composition (UMA association). These bodies are more common near Chilas (Fig. 6.24) and consist of dunite, peridotite, pyroxenite, troctolite, olivine gabbros, gabbronorite and anorthosite characterised by more magnesian pyroxenes and hornblende, and much more calcic plagioclase (An_{83-98}) than the main gabbronorites (Photos. 50 to 53). The UMA rocks display excellent depositional structures such as modal layering, graded and/or cross bedding, slump folding, synsedimentary faulting, and so on. The mutual relations of the two groups have been differently described. Khan et al. (1989) favoured a later phase of magma generation for the UMA rocks rather than considering them cumulates to the main gabbronorites. The rocks of the two associations are cut by a suite of basic (amphibolite) dykes. These dykes may be contemporaneous with the 70 Ma Jutal dykes in the Kohistan Batholith (Petterson and Windley 1992).

The age of the Chilas Complex is not clearly known, particularly with respects to the Shyok Suture. It intrudes the Jaglot Schist Group to its north and the Kamila Amphibolites to its south. These relations and the presence of Jutal-like dykes suggest a >70 Ma Cretaceous age. A much lower degree of deformation in the Chilas, when compared to early arc rocks, has been taken to suggest that the Complex was emplaced after the suturing of the Karakoram plate and Kohsitan (Treloar et al. 1996). U-Pb ages on zircon separated from a gabbronorite in Upper Swat fall on a concordia of 84 Ma (Zeitler et al. 1980), which would be consistent with a post-suturing emplacement of the Chilas Complex.

Details of mineral chemical data and major, trace and rare-earth elements geochemistry have been presented by Jan et al. (1984, 1991) and Khan et al. (1989, 1993). The Chilas Complex shares several traits with island arc plutonic rocks. On an MFA diagram, it displays a typical calc-alkaline trend (Fig. 6.25). Trace element patterns of the rocks (Fig. 6.26) have marked negative Nb anomalies, positive Sr, Ba, and P anomalies, low Rb, and $(LREE/HREE)_N$ >1.0. The trace element features of the mafic dykes are similar to their host rocks but, like the UMA rocks, they have tholeiitic rather than calc-alkaline signature of the main gabbronorites. Khan et al. (1989, 1993) suggested that the Chilas Complex was generated from a mantle diapir in the early stages of arc rifting or back-arc spreading. Earlier magma (main gabbronorites) was calc-alkaline. Further melting of this source region produced picritic (UMA association) and tholeiitic melts (mafic dykes).

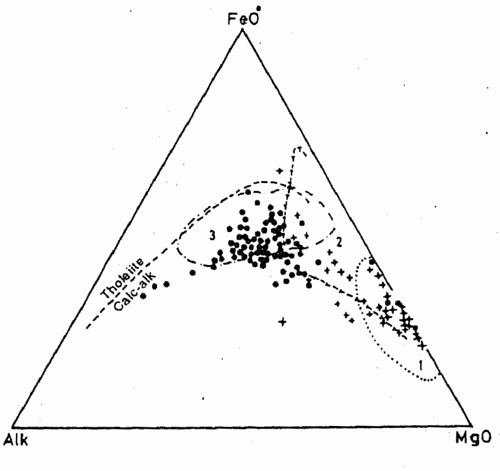


Fig. 6.25. AFM plot of the rock analyses from the Chilas Complex. Dots = main norites and retrograde amphibolites, crosses = analyses from the UMA association (data source: Hamidullah and Jan 1986); 1–ophiolitic cumulates (Coleman 1977), 2–island arc cumulates, and 3–non-cumulates (Beard 1986). Tholeiitic-Calc-alkaline boundary after Barker and Arth (1976).

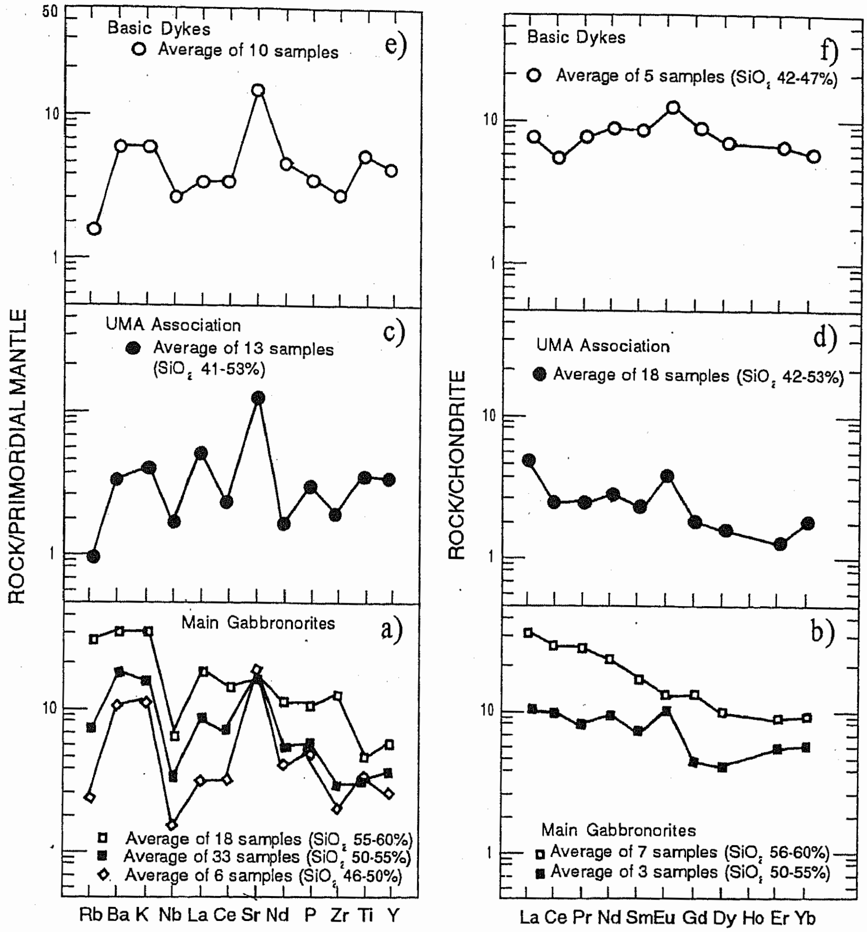


Fig. 6.26. Multi-element trace- and rare-earth patterns for the three phases of magmatism in the Chilas Complex. Normalising values for the primordial mantle are from Wood et al. (1979) and chondrite from Nakamura (1974). (From Khan et al. 1993).

Southern or Kamila Amphibolite Belt: This is a complex belt of amphibolites and a variety of other rocks occupying the southern part of the Kohistan arc. It has been described by Jan (1979a, 1988, 1990), Bard et al. (1980), Treloar et al. (1990) and Khan et al. (1993). It consists predominantly of amphibolites and subordinately of hornblendites, hornblende gneisses, diorites (± quartz), granitoids (including plagiogranite), with minor pegmatites and metasediments. The amphibolites can be divided into (a) fine-grained, commonly banded, and (b) medium- to coarse-grained, homogeneous to gneissose, locally banded/layered types (Jan 1979a, 1988, Treloar et al. 1990). The former are considered as mainly volcanic and the latter plutonic in origin. Relics of pillow lavas (Thak Valley, Upper Swat) and gabbroic rocks (Indus gorge, Swat, Dir) support this genetic subdivision. Much of the amphibolite belt has been strongly sheared, complexly folded and locally made up of imbricate thrust slices. Therefore the relations between the metavolcanic and metaplutonic precursors have been frequently obliterated. A lot of granulation and banding can be related to shearing (Jan 1990).

The relative proportions of the metavolcanic and metaplutonic amphibolites vary along the strike of the arc. The belt exposed in the valleys to the south of Chilas is composed essentially of fine- to medium-grained amphibolites of volcanic parentage. Westwards in the Indus, Swat and Dir valleys, more than two thirds of the belt is made up of metamorphosed plutonic rocks initially intrusive into volcanic rocks. In many places (e.g., Pattan–Kayal, Khwaza Khela, south of Fatehpur, and southern Dir), the relics are gabbronorites similar to those of the Chilas Complex. Some of these are layered and contain ultramafic-mafic-anorthositic components. But some amphibolites appear to be derived from tholeiitic plutons (Shah et al. 1992, Jan et al. 1993).

The amphibolite belt shows upper greenschist to amphibolite facies conditions of metamorphism (550–680° C, 4.5–6.5 kbar PH_2O). Garnet and diopside have abundantly developed locally (e.g., Lilauni, north of Shangla), suggesting higher PT conditions. Bard (1982b) and Treloar et al. (1990) think that operating pressures may locally have exceeded 8–10 kbar. This is entirely consistent with the occurrence of kyanite + zoisite-bearing high-pressure assemblages in a few places (Jan and Karim 1995). In some places the rocks are migmatised and partially melted. It appears that some of the trondhjemitic and granitic rocks in this belt and the Kohistan Batholith are extracted from these amphibolites (Jan 1988).

The volcanic rocks of the southern amphibolite belt are probably the oldest in the Kohistan arc, although their relationship to the Jaglot Schists is not known. Earlier geochemical studies (Jan 1988, 1990, Shah et al. 1992, Khan and Thirlwall 1990) show that the amphibolites generally have island arc-type calc-alkaline to tholeiitic chemistry. More recently, Khan et al. (1993) showed that the metavolcanic amphibolites range from basalt, through basaltic andesite, to andesite, and can be divided into HFS-enriched E-type and HFS-depleted D-type (Fig. 6.27).

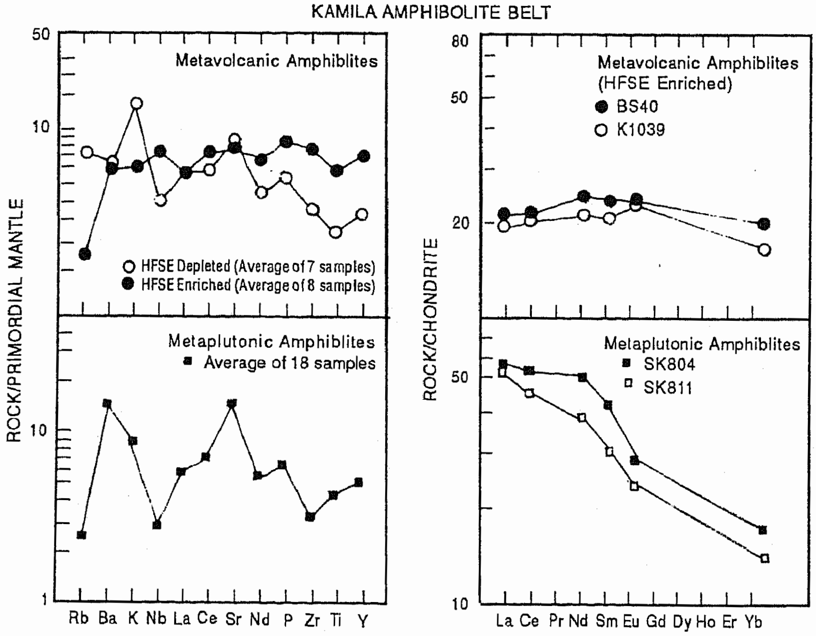


Fig. 6.27. Multi-element trace- and rare-earth patterns for the basic rocks in the Kamila amphibolite belt. Normalising trace-element data for the primordial mantle are from Wood et al. (1979) and rare-earth data for the chondrite are from Nakamura (1974). (From Khan et al. 1993).

A few rocks appear to be transitional type. The metamplutonic amphibolites are calc-alkaline and somewhat similar to the D-type. They show positive spikes for P, Sr and K, and negative spikes for Nb and Zr. Combined with $(\mathrm{LILIE/HFSE})_N$ ratios of more than one, these may be the product of the same phase of subduction-related magmatism in early to mature stages of arc growth. The E-type amphibolites represent an early phase of magmatism with an involvement of E-MORB source in their generation. Treloar et al. (1996) think that the E-type, with flat REE and incompatible trace element pattern, may have developed in oceanic plateau environments. If so, these may represent the oceanic basement on which the arc was built.

***Jijal Complex*:** Sitting on the Indus Suture along the southern fringe of the Kamila amphibolite belt are a series of mafic-ultramafic, seemingly cumulus, complexes. Of these the Sapat, Jijal, and Tora Tigga Complexes have been studied in some detail. The Sapat Complex in southeastern Kohistan consists of basal ultramafic cumulates more than a kilometre thick. These consist of dunite, peridotite, pyroxenite, and chromite-rich layers, passing upward into troctolite/metagabbros (Photo. 54). Jan et al. (1993) suggested that these rocks represent island arc cumulates in a chamber of tholeiitic magma. The Tora Tigga Complex in western Dir (Fig. 6.28) consists of olivine- and pyroxene-rich rocks, hornblendites, metamorphosed gabbronorites and plagiogranites. Preliminary geochemistry of these rocks suggests cumulate origin to a possibly calc-alkaline magma (Jan et al. 1983, Jan and Tahirkheli 1990).

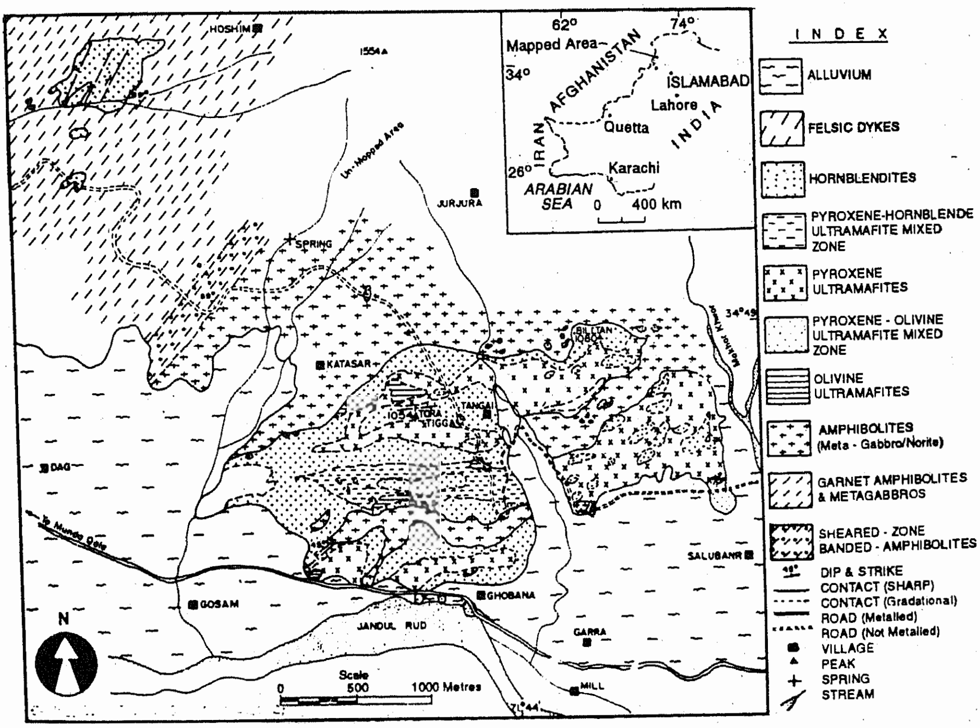


Fig. 6.28. Geological map of the Tora Tigga complex, Munda area, Dir (from Jan et al. 1983).

The Jijal Complex is a tectonic block covering about 150 km^2 area. It consists of garnet granulites in the north and ultramafic rocks in the south (Jan 1979b, Jan and Howie 1981). The ultramafic rocks are devoid of garnet and consist of abundant diopsidite and some dunite, peridotite, websterite and chromitite. Layering is seen in many places and Miller et al. (1991) have presented a cumulus stratigraphy of the Complex. Jan and Windley (1990) described the mineral chemistry and showed that the high Cr/(Cr+Al) ratio of the chromite was suggestive of arc environment. The granulites are dominated by garnet-clinopyroxene-plagioclase-quartz-rutile paragenesis, with some hornblendites (± garnet ± clinopyroxene ± rutile) and "garnetite", and minor garnet pyroxenites (Photos. 55 to 57). These rocks also display local layering.

Jan and Howie (1981) suggested that both groups of rocks were metamorphosed in granulite facies (see Chapter 7). Radiometric dates of 91 ± 6.3 to 114 ± 39 have been reported for the complex (Coward et al. 1986, Yamamoto and Nakamura 1996, Sano et al. 1996), but it is not clear whether these reflect metamorphism or magmatism. Trace element pattern of the granulites is shown in Figure 6.29. The overall geochemical abundance of the granulites, especially the Nb and Zr troughs and Sr peak, are suggestive of subduction-related magmatism. Jan and Windley (1990) proposed that the ultramafic rocks are cumulates related to a primitive arc tholeiitic magma, whereas the granulites have been derived from a more evolved (calc-alkaline) magma. The two suites may represent a continuous sequence of cumulates, the granulites having formed from an evolved liquid after the separation of large quantities of magnesian pyroxenes and olivine. Alternatively, they may have formed from independent batches of arc-related magmas. The field relations of the two have been differently interpreted.

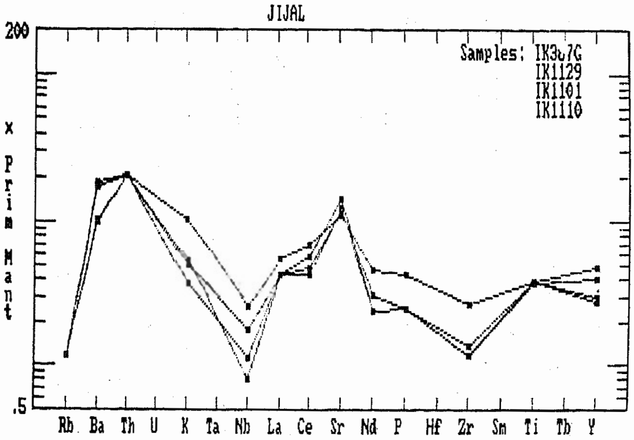


Fig. 6.29. Trace-element spidergram for garnet granulites of the Jijal Complex. Primordial mantle values are from Wood et al. (1979). (Jan and Weaver, unpublished data).

Tectonic Model for Kohistan

Since 1968, the status of the Kohistan zone has changed from the least known to one of the best studied areas of the Himalayas. There have been numerous publications and many opinions about the petrological evolution of this region. Early models envisaged that the Kohistan zone is made up of the upper, middle and lower crust of an island arc turned on end during the Himalayan Orogeny. As new data gathered, it became obvious that such a simplified model is inadequate for this very complex area. Thus, suggestions ranging from two arcs to frontal arc-back arc, and oceanic crust-island arc collage have been proposed. In the following we summarise a model currently postulated by Treloar, Petterson and Jan (1996).

It appears that only the southern amphibolite belt contains rocks which developed in an intra-oceanic island arc in strict sense. These early stages of arc growth occurred in response to northward subduction of the Neo-Tethyan oceanic lithosphere and were located on a magmatically thickened oceanic crust. Once subduction was initiated, extension in the back-arc led to development of a basin in which Jaglot sediments were deposited and Chalt Volcanics were extruded, followed by further sedimentation and volcanism of the Yasin Group (Khan et al. 1994). This phase culminated with melting of the lower arc crust or underlying mantle, generating the stage I plutons of the Kohistan Batholith prior to suturing of the arc to the Karakoram plate. Thickening of the arc accompanied suturing, as did pre-80 Ma deformation and partial melting in the Kamila shear zone (Treloar et al. 1990).

Subsequently, the arc experienced tensional rifting during which the Chilas Complex was emplaced (Khan et al. 1989). The initiation of stage II plutonism at about 85 Ma (Petterson and Windley 1985, Treloar et al. 1989) started synchronously with the emplacement of the Chilas Complex at about 85 Ma (Zeitler et al. 1980). The emplacement of the huge Chilas body, possibly as a result of asthenospheric upwelling, may have supplied the heat for widespread regional metamorphism and initiation of deep arc melting, contributing to subsequent stage II plutonism. The cessation of this period of extension is dated by erosive exhumation of the stage II plutons in Swat and Dir. Renewal of extension at about 60 Ma, involving fore-arc collapse, resulted in deposition of Baraul Banda Formation and extrusion of Utror and Shamran Volcanics in an extensional basin during Paleocene (Sullivan et al. 1993). The younger plutons may be related to melting in a thickened arc crust following collision with India during Early Tertiary.

Thus the development of Kohistan was largely accomplished within an extensional framework interrupted by two periods of compression. One of these may be related to the suturing of the arc with the Karakoram plate >85 Ma ago, and the other to the suturing of the arc with Indo-Pak some 55–60 Ma ago. Extensional regimes have long been advocated in island arcs (e.g., Hamilton 1989). The magmatic growth history of Kohistan can be described within the bounds of a retreating subduction zone (Royden 1993).

Karakoram Block

Much of the magmatism in the Karakoram block is granitic in nature. Mafic plutons are rare and lamprophyre dykes are small and sparse. There also are some poorly defined Jurassic volcanic rocks. Jan et al. (1981b) pointed that the granitic rocks of the Karakoram block occur in two distinct belts, the Karakoram granitic belt and the Khunjerab-Tirich Mir granitic belt, separated by a major fault, the Reshun Fault. Searle (1991) has given detailed accounts of the petrography, geochemistry and petrology of the igneous rocks, together with many radiometric dates, from the Karakorams. In view of this readily available source of information, we provide only a summarised account of the magmatism in the Karakoram block.

Jurassic(?) volcanics

Underlying the Chitral Slates there is an up to 4 km broad belt of volcanogenic rocks metamorphosed in greenschist facies (chlorite-epidote-albite-quartz). The rocks are mostly intermediate in composition, with some thin acidic layers containing abundant quartz. According to Calkins et al. (1981) they may consist of both flows and tuffs. Named as Koghozi greenschist, these rocks become finer grained and thinner in the southwest and northeast.

There is a local sheen due to development of chlorite; illite is absent. Associated with the greenschist are Cu-Pb-Fe sulfide disseminations and veins (Pudsey et al. 1985).

In the northern Karakoram region of Hunza, Gaetani et al. (1990a, 1993) report the occurrence of faulted blocks of volcanic and associated sedimentary rocks, overlain by the Tupop Formation of Cretaceous age. The volcanics are pillowed and considered to be anorogenic basalt. The Tupop Formation consists of conglomerates interbedded with sandstone and unfossiliferous nodular limestone. The conglomerate contains up to 2 m blocks, some of which are rhyolitic in composition. There is no further information published on these rocks. This part of the Karakoram has been regarded to have been rifted away from Gondwana during the Permian, but there seems a general lack of volçanic rocks (Gaetani et al. 1990a). The northeastern Karakoram in Ladakh, on the other hand, contains (?) rift-related volcanic rocks (Gergan and Pant 1983).

Khunjerab–Tirich Mir Granitic Belt

In western Chitral and the area to the south and west of Baroghil Pass, several granitic intrusions have been mapped by Calkins et al. (1981), Pudsey et al. (1985), Buchroithner (1980, 1985), and Leake et al. (1989). These intrude a Jurassic-Devonian tectonostratigraphic sequence thrust SSE, along the Reshun Fault, over the Cretaceous Reshun Formation. Desio (1979) calls this fault as Chitral Fault and according to him it is continuous with the Upper Hunza Fault. The latter is viewed by Desio as the northern boundary of the Karakoram. The Reshun Thrust, therefore, serves as a major tectonic break between the Khunjerab–Tirich Mir granitic belt to the north and the Karakoram granitic belt to the south (Jan et al. 1981b). Eastwards, granitic rocks of comparable tectono-stratigraphic position occur in Baroghil Pass, Sost, Giraf, Khunjerab and, possibly, beyond. The Wakhan granites probably also belong to this group (Buchroithner and Scharbert 1979). Not much petrological data have been published on the rocks of this belt.

Of the several plutons in western Chitral, those of Tirich Mir, Garam Chashma, and Kafiristan are the largest and of batholithic dimensions. Poorly constrained by radiometric dates, Leake et al. (1989) divide these into two groups. The Tirich Mir and Kafiristan plutons consist of hornblende-biotite granodiorite gneiss containing up to 5 cm long plagioclase crystals. Desio et al. (1964) have reported a 115 ± 4 Ma Rb-Sr biotite age for the Tirich Mir pluton. Slightly younger ages (93 to 86) have been reported for the Zebek intrusion to the northwest. But in view of recent discovery of pre-Ordovician granites in Yarkhun Valley, it would not be surprising if some other plutons in this belt are also of this older event. The Bumburet pluton, for example, is equated with those of Yarkhun (Le Fort et al. 1994). The second group of intrusions (e.g., Garam Chashma) are typical leucogranites (containing two micas, tourmaline and garnet) associated with the highest grade of metamorphism. K-Ar biotite ages of 48 ± 2 Ma for Kafiristan and 20 ± 1 and 19 ± 1 Ma for the Garam Chashma pluton indicate younger cooling history (Searle 1991), but Leake et al. (1989) thought that the leucogranites are similar to the 28–12 Ma High Himalayan leucogranites (Le Fort 1981, Searle and Fryer 1986). The Tirich Mir pluton has an associated band of gabbroic rocks along the Tirich Fault. These have been described in a previous section.

In Giraf area of upper Hunza Valley, Desio and Martina (1972) described a 53 Ma (K-Ar) quartz syenitic and monzonitic pluton comprising perthite phenocrysts in a matrix of feldspar, quartz, hornblende, pyroxene and biotite. It is cut by dykes of granodiorite, aplite and porphyries. A number of small plutons occur near Sost, 50 km north of the Karakoram

Batholith. These range from amphibole-biotite-quartz diorite (± clinopyroxene) to granite. One pluton, a pink granite, is rich in sphene and Fe-oxide and characterised by large phenocrysts of pink K-feldspar. Ogasawara et al. (1994) report K-Ar hornblende and biotite ages of 105 ± 5 Ma and 95.0 ± 4.7 Ma for the Giraf Syenite, and 84.2 ± 4.2 Ma and 85.9 ± 4.3 Ma K-Ar biotite ages for the plutons north of Sost. Debon et al. (1996) report similar K-Ar dates: 112 ± 8 (hornblende), 95 ± 5 (biotite) for Giraf, and 112 ± 3 (hornblende) and 87 ± 4 (biotite) for Shachkan plutons north of Sost.

A large mass of plutonic rocks occurs along the Pakistan-China border near Khunjerab. Intruding Paleozoic-Early Mesozoic sediments, these consist mostly of hornblende-biotite granodiorite quite similar to those of the Karakoram Batholith. There also are some adamellite, granite and clinopyroxene-bearing (?)quartz diorite. K-Ar biotite ages range from 96.9 ± 4.8 Ma, to 107 ± 5, and hornblende ages from 105 ± 5 to 115.6 ± 0.8 Ma (Treloar et al. 1989c, Ogasawara et al. 1994, Debon et al. 1996). Since all the three occurrences of the uppermost Hunza Valley are almost undeformed, these ages may well record the age of formation of the rocks. Recently, Zanchi (1993) has mapped stocks of granodiorite composition in Shuijerab–Shimshal Pass. These appear to be a continuation of those of the Khunjerab area, with a K-Ar biotite age of 104 ± 3 Ma (Debon et al. 1996). Gaetani et al. (1990b) prefer to group the K2 gneisses with the Khunjerab-Tirich Mir granites rather than with the Karakoram Batholith. If so, this belt would also stretch for several hundred kilometres. However, Desio (1979) has raised doubts whether the granites near the Khunjerab and Kilik Passes and those of Tirich Mir belong to the same igneous belt.

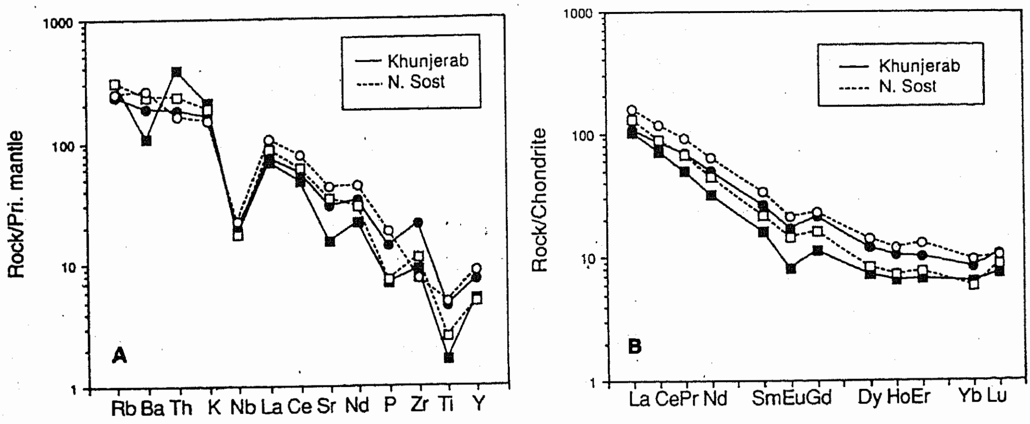


Fig. 6.30. A- Primordial mantle-normalised trace-element variation diagrams for granitoids from the Khunjerab and Sost plutons. B- Chondrite-normalised rare-earth element (REE) patterns for granitoids from the Khunjerab and Sost plutons. (From Searle 1991).

The geochemical data for the granitic rocks from Khunjerab and Sost (Fig. 6.30) suggest they are related to subduction processes (Searle 1991) and possibly linked with the collision between the Karakoram plate and Kohistan arc (Debon et al. 1996). The upper Hunza region plutons range from Mg-K metaluminous granitoids with biotite and amphibole to two-micas peraluminous type. The metaluminous ones may have originated from a mantle source with little crustal input, whereas the peraluminous granitoids are indicative of an increasing involvement of the continental crust. For further details on geochemistry and tectonic modelling of these granitoids, Debon et al. (1996) may be consulted.

Karakoram Granitic Belt

Also known as the Karakoram (Axial) Batholith, this belt extends for 600 km from southwest of Drosh to west of Pangong Lake and attains a width of up to 30 km. This is a composite batholith with many plutons, dykes and veins of variable composition emplaced over a period exceeding 100 m.y. (Photos. 58 and 59). Multiple intrusions can be observed in many places and, like the Kohistan Batholith, young leucogranites may occur in swarms. Searle et al. (1989) and Crawford and Searle (1992, 1993) have proposed a broad subdivision of the Batholith into pre-collision and post-collision units. The pre-collision rocks include the Hushe Gneisses and (?) Kande plutonic unit (Jurassic), the Darkot Pass, Hunza, K2, Broad Peak, Muztagh Tower units and early phases in Yasin Valley (Cretaceous). The post-collisional (Tertiary) ones comprise Batura, Baltoro, Mango Gusar, Masherbrum and Sumayar units, Hunza and Korophon dykes, and lamprophyres (Fig. 6.31). The Batholith consists of large plutonic units displaying major differences in age, chemical and mineralogical composition, and tectonometamorphic history (Debon et al. 1987). Therefore, we follow Searle (1991) and describe it from the west, central, northern and eastern sectors of the Karakoram Range in Pakistan.

Western Karakoram: The western-most exposure of the Karakoram Batholith, the Kesu-Bunizom pluton, extends from Mastuj to Kesu and beyond, a distance of more than 100 km. This is the least studied part of the batholith, with fragmentary information given in Pudsey and others (1985, 1986). The batholith here consists of deformed (locally migmatitic) diorite, quartz diorite and granodiorite intruded by younger, underformed granitic dykes and pegmatites. Ar-Ar hornblende age of 111 ± 5 Ma (Pudsey 1986) suggests that some units of the pluton were emplaced in Paleozoic sediments before the formation of the Shyok Suture. However, there are younger intrusions that apparently cut the suture. The fresh biotite granite of Bunizom may also be post-collisional.

To the west of Ishkoman the Karakoram Batholith splits into Darkot Pass (northern) and Ghamu Bar (southern) plutonic units. These have been studied in some detail by Debon et al. (1987). The Darkot plutonic unit intrudes the upper Paleozoic Darkot Group and consists of a wide range of rocks from quartz-monzodiorite to leucogranite. The most abundant rocks are medium- to coarse-grained, porphyritic adamellite and quartz-monzodiorite. Chemically these rocks compare well with subalkaline and calc-alkaline granitoids from other areas (Fig. 6.32). The porphyritic type yields a 111 ± 6 Ma Rb-Sr age whereas leucogranites have an isochron age of 109 ± 4 Ma (Debon et al. 1987). The main intrusion may thus be Aptian-Albian, but there is a great range in K-Ar hornblende-biotite ages (199 to 44 Ma) and, according to Searle (1991), the Darkot Pass unit may be Jurassic-Cretaceous.

The Ghamu Bar pluton is also emplaced in the Darkot Group and contains abundant septa of the sedimentary rocks. Much of the granitoid, particularly in the Umalsit area, is medium to fine-grained adamellite containing up to 5 × 18 cm megacrysts of K-feldspar. Northwards the porphyritic texture disappears progressively and the pluton becomes a heterogeneous mixture of adamellite, granodiorite and leucogranites. The Ghamu Bar pluton is traversed by mylonitic zones, and alteration is common. Major and RE element geochemistry supports their calc- alkaline nature. Debon et al. (1987) report poorly constrained Late Paleozoic Rb-Sr

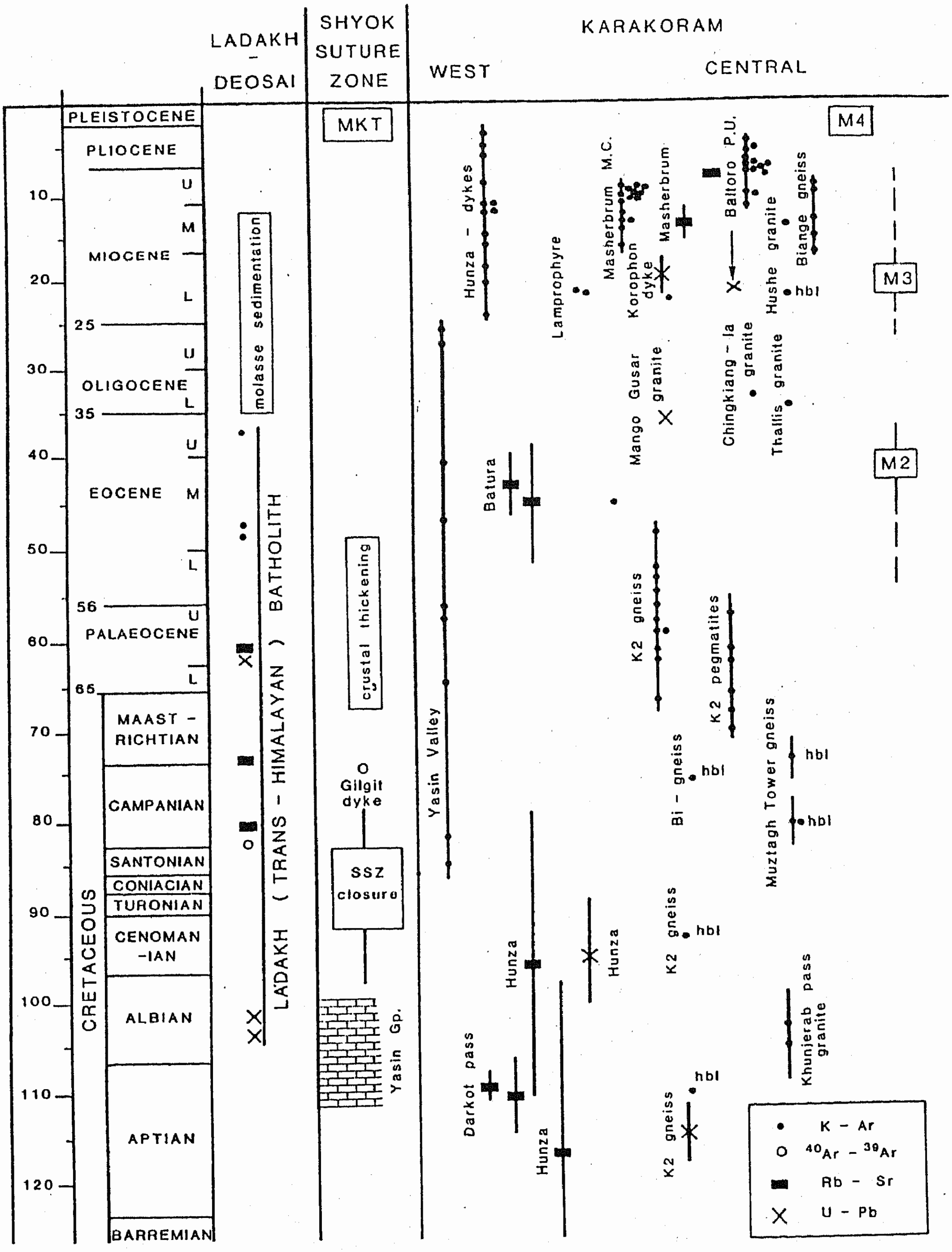


Fig. 6.31. Compilation of Cretaceous-Tertiary radiometric ages of the central Karakoram and the Ladakh sector of the Kohistan–Ladakh terrane (from Searle et al. 1989).

ages for the pluton.

Along the Karambar Valley, the Karakoram Batholith comprises four plutonic complexes: (1) the mid-Cretaceous, calc-alkaline Hunza plutonic unit, (2) a stock of subalkaline, porphyritic granitoids, (3) a composite group of fine-grained granitoids, and (4) the Koz Sar alkaline complex. These have been described in detail by Debon and Khan (1996). The alkaline rocks are unusual and merit some comments. They constitute a ~5× 20 km pluton and have a Rb-Sr isochron age of 88 ± 4 Ma. The most common members consist of metaluminous to slightly peralkaline monzonite (± quartz), granite and leucogranite with Fe-rich mafic silicates including clinopyroxene and calcic amphibole. The alkaline character of these rocks testifies to the development of extensional tectonics, a process compatible with an oblique collision and/or with the decrease of the convergence velocity. This late orogenic pluton appears to be genetically related to the nearby subalkaline granitoids and is originated from the same mantle source with a small crystal contribution (Debon and Khan 1996).

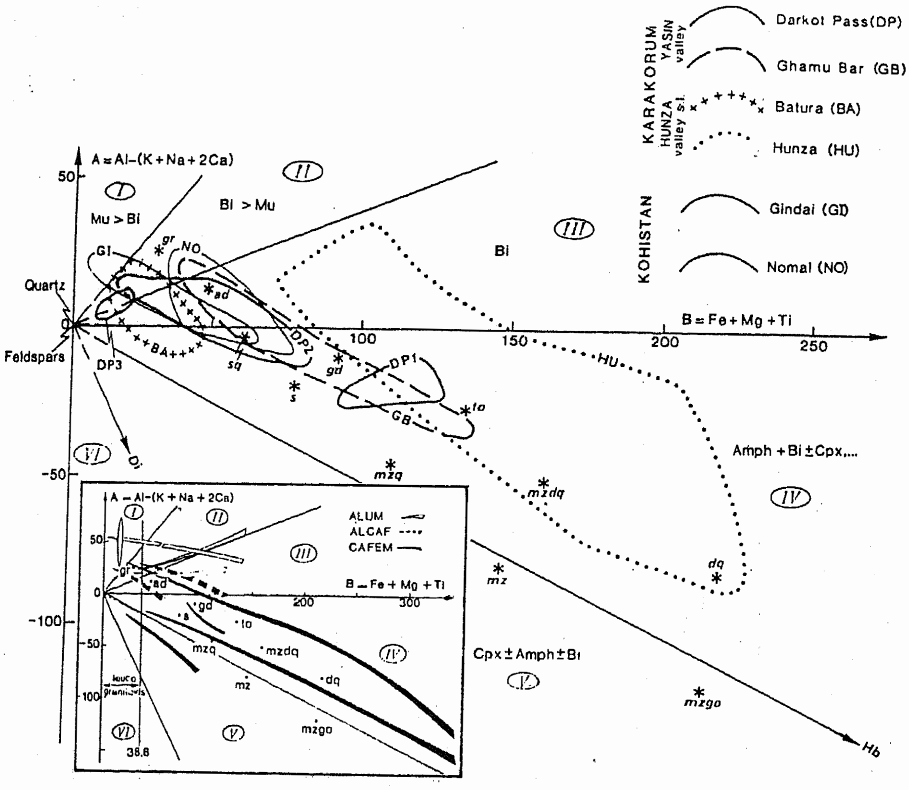


Fig. 6.32. Distribution of the six plutonic units in the "characteristic mineral diagram." The parameters are in gram-atoms × 10^3 in 100 g of rock or mineral; A is the "aluminous index" and B is proportional to the dark mineral content (La Roche 1964). This diagram separates peraluminous rocks or minerals with positive value of A, e.g., biotite (Bi), muscovite (Mu), from metaluminous ones, e. g., amphibole (Amph, Hb), clinopyroxene (Cpx, Di). Each of its six sectors, numbered from I to VI, corresponds to a specific mineralogical composition; to a first approximation: I–rocks with Mu > Bi (by volume), II– Bi > Mu, III–Bi, IV–Bi + Amph ± Cpx, V–Cpx ± Amph ± Bi. The trends of the three main types of magmatic associations (see the inset) are distinguished: ALUM = aluminous, ALCAF = alumino-cafemic, CAFEM = cafemic (from Debon et al. 1987).

Hunza Karakoram: Four distinct plutonic units constitute the batholith in the Hunza Valley section. These are the Hunza Plutonic Complex, the Batura and Sumayar plutons and the leucocratic dykes. The Hunza Plutonic Complex is reliably dated at 95 ± 5 Ma (U-Pb zircon) with two K-Ar dates of 46 Ma (hornblende) and 26 Ma (biotite) (Le Fort et al. 1983). It contains a range of rocks, from quartz diorite to granite, but granodiorite is predominant. The southern part is well foliated and metamorphosed, with growth of garnet. Here also occurs a migmatite consisting of several lithologies ranging from mafic to leucocratic. Large blocks of mafic quartz diorites in granodiorite and interfingering relations between the two near Gulmit may suggest that the quartz diorites were emplaced slightly earlier than the granodiorite. A dense anastomosing network of late dykes consists of a co-magmatic biotite-aplite monzogranite, two-mica granites, and garnet-muscovite pegmatitic leucogranites with and without aplitic central part. Such dykes, as well as those of granodioritic composition, also occur in the metasediments to the south of the main batholith (Searle 1991). In Nagar and Aliabad there are two, over 100 m thick, sheets of garnet-muscovite leucogranite cutting regional foliation. There are granodiorite sheets at Hasanabad and Bar. These parallel the regional folition and are probably related to the Hunza unit, but there also are younger granitic bodies near Hasanabad.

The Sumayar pluton intrudes staurolite-grade metasediments to the south of Nagar. This 4 km wide body of leucogranite contains two micas, garnet and tourmaline, and intrudes the deformation/metamorphic fabric of its host rocks. The Batura plutonic unit consists of undeformed biotite-hornblende granodiorite and granites, two samples of which have Rb-Sr isochron age of 63 ± 2 and 42.8 ± 5.6 Ma (Debon 1996, Debon et al. 1987) This subalkaline and ferriferous association is dominated by light-coloured metaluminous and slightly peraluminous rocks with low initial $^{87}Sr/^{86}Sr$ ratio (0.7050–0.7056). It has apparently originated in a mantle-source with small crustal contribution, through a two-stage melting process (Debon 1996). East of Pasu along the Shismal Valley, there are two very long and narrow dykes of lamprophyre containing hornblende phenocrysts. These resemble those of the Gasherbrum area (Searle 1991, Desio and Zanettin 1970).

Northern Karakoram: This includes the area adjacent to the Chinese border, hosting some of the highest mountain peaks in the world. The K2 Gneiss consists of interlayered ortho- and paragneisses of the middle crust uplifted along the hanging wall of a large-scale thrust (Searle et al. 1989, 1990). The orthogneiss contains K-feldspar megacryst, biotite and hornblende, and the paragneisses comprise clinopyroxene-hornblende psammites and garnet-diopside marbles. U-Pb zircon data show that the orthogneisses were intruded at 115–120 Ma, whereas K-Ar hornblende cooling ages are 111–94 ± 3 Ma. Leucogranitic pegmatite dykes, consisting of quartz, two-feldspars, garnet, two-micas and tourmaline, are presumed to be pre-collisional, with K-Ar biotite cooling ages ranging from 70 to 58 Ma (Searle 1991, Crawford and Searle 1992).

Large-scale intrusions of quartz diorite, post-dating Lower Cretaceous host sediments, are closely associated with the K2 Gneisses in Broad Peak-Gasherbrum area. These porphyritic rocks have associated ash-flow/tuffs both of which are strongly altered. The Broad Peak quartz diorites are very similar to the K2 Gneisses and have been correlated with them and with the Muztagh Tower unit, all of which have typical calc-alkaline chemistry and were probably related to an Andean-type magmatism.

Several dark coloured lamprophyre dykes intrude the Gasherbrum sediments, Doksam sequence and the K2 gneisses. One lamprophyre also occurs in the Hushe Gneiss; the two near Pasu have been mentioned. These occur in 1.5 m or thinner dykes, clearly post-dating deformation, and are classified as amphibole-rich vogesites and biotite-rich minettes belonging to the calc-alkaline shoshonitic group. These have K-Ar biotite ages of 22 ± 0.7, 24 ± 1.0 and 46 ± 2.0 Ma (Searle 1991).

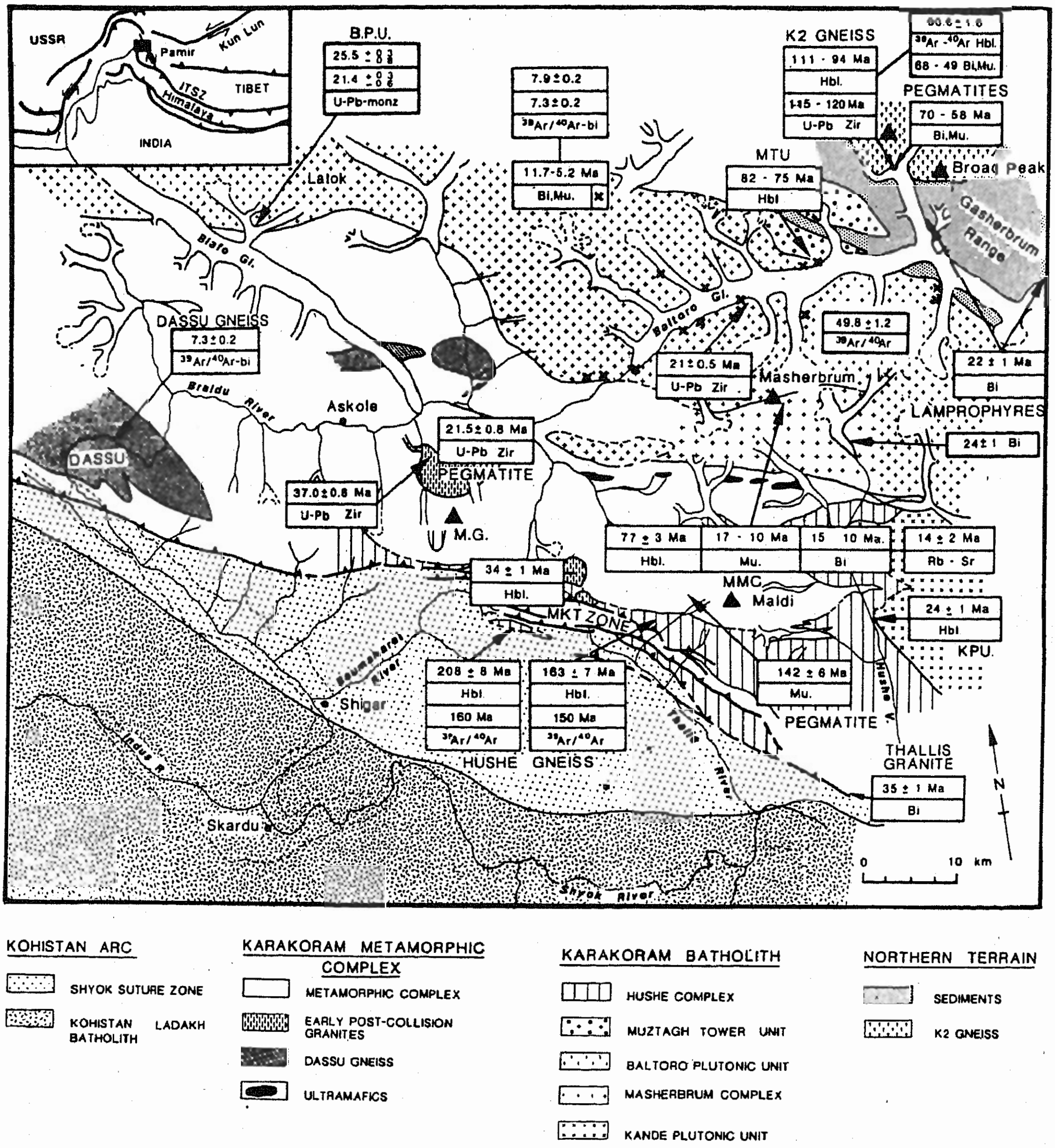


Fig. 6.33. Geological sketch map of the Biafo-Baltoro-Hushe area of the central Karakoram showing all radiometric age data and the location of samples dated. U–Pb data are either from zircon (zir) or monazite (monz). K-Ar data are from hornblende (hbl), muscovite (mu) or biotite (bi), MKT–Main Karakoram Thrust, BPU–Baltoro Plutonic unit, MTU–Muztagh Tower unit, MMC–Masherbrum complex, MG–Mango Gusar granite (from Searle 1991).

Photo. 56. Jijal Complex south of Pattan. Garnetite relics in hornblendite cut by an earlier set of garnetiferous granulite veins and a later set of epidote veins. (Photo: *M. Q. Jan*).

Photo. 57. Jijal garnet granulite cut by a pegmatite containing garnet porphyroblasts. The light margin consists of zoisite. (Photo: *M. Q. Jan*).

Photo. 58. Close up view of the Karakoram batholith north of Hunza. Granite dykes of many generations intrude the Late Cretaceous granodiorite. (Photo: *M. Q. Jan*).

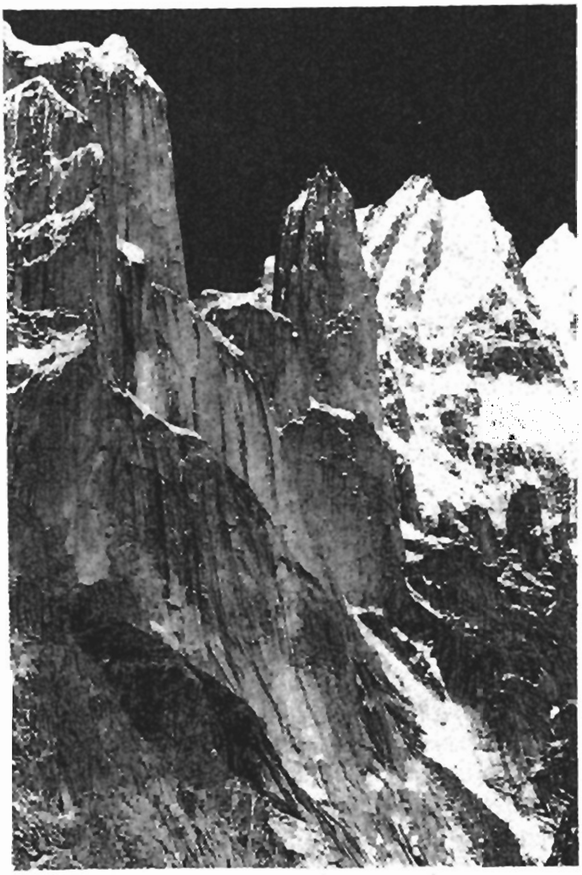


Photo. 59. View of the Great Trango Tower (6,500 m), showing outcrops of biotite monzogranite and garnet-two mica leucogranite-part of the Baltoro plutonic unit. (Photo: *M. P. Searle*).

Photo. 60. Orbicular granite in the Baltoro pluton, Trango Towers. Orbicules of biotite rim K-feldspar in monzogranite. (Photo: *M. P. Searle*).

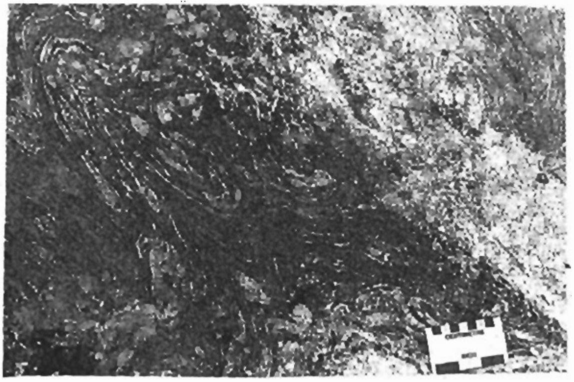


Photo. 61. Leucogranite dyke (21 Ma) intruding and cross-cutting folded sillimanite gneisses with melt pods, south of the southern margin of the Baltoro granite, Braldu Valley. (Photo: *M. P. Searle*).

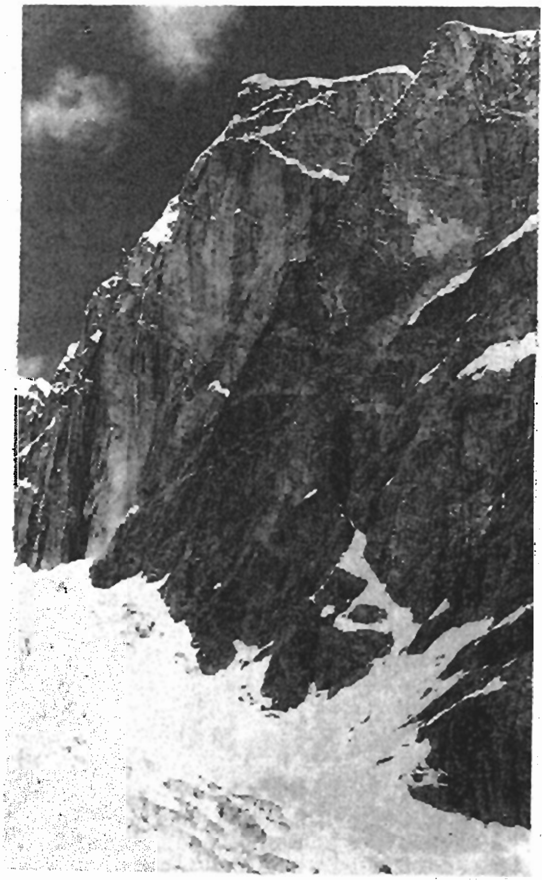


Photo. 62. Baltoro leucogranite (21 Ma) intruding older dioritic amphibolite of Masherbrum on the west bank of Yermanandu Glacier. (Photo: *M. P. Searle*).

Photo. 63. Biotite leucogranite enclosing rafter of dioritic amphibolite, Baltoro granite pluton (Masherbrum peak). (Photo: *M. P. Searle*).

The Baltoro–Biafo–Hushe areas: This area stretches from Snow Lake-Hispar in the west to the Siachen Glacier and Khapalu in the east. Several large and small plutons have been reported from this area (Fig.6.33). East of Hushe Valley the Karakoram Batholith is represented by the Kande plutonic unit. It is a composite unit, the earliest phases of which are deformed diorites that may be coeval with those of Hunza and Muztagh Tower. A regional andalusite-grade metamorphism is associated with this unit. There are tonalite to granodiorite lithologies similar to K2 and Masherbrum. The youngest phase is a garnet-two-micas leucogranite forming a complex of dykes and veins. These may be related to the Baltoro plutonic unit.

The Muztagh Tower unit crops out immediately north of the Baltoro plutonic unit. It consists of deformed and strongly foliated tonalitic to granodioritic gneisses with K-Ar hornblende cooling ages of 82–75 ± 3 Ma. Garnet-bearing two-micas leucogranites of Miocene age intrude the southern margin of this unit (Crawford and Searle 1992).

The Baltoro plutonic unit consists of massive, co-magmatic rocks of monzogranite to leucogranite composition stretching for 100 x 10–20 km in the central Karakoram. Rare orbicular biotite granites occur on the Trongo Towers (Photos. 59 and 60). The petrology, trace element, REE and isotope geochemistry of this unit have been presented in detail by Searle (1991). There are garnet + biotite ± muscovite leucogranitic dykes emanating in the country rocks (Photo. 61). Numerous sillimanite-grade aligned xenoliths occur along the southern margin in the Biafo Glacier area, but the core of the pluton is massive and devoid of xenoliths. Parrish and Tirrul (1989) reported a U-Pb zircon age of 21 ± 0.5 Ma on samples from Urdukas on the Baltoro Glacier. The Baltoro monzogranites have high K, Th (up to 85 ppm), Ba (up to 2,500 ppm), Sr (up to 800 ppm), LIL/HFS ratios, and negative Nb anomaly. It has been suggested that they may be lower crustal melts promoted and contaminated by a mantle input.

The Masherbrum Complex consists of rafts and blocks of biotite-hornblende orthogneiss, biotite-rich paragneiss and layered calc-silicate marbles. It is intruded by massive leucogranite sheets and garnet-two micas pegmatite-aplite leucogranitic dykes (Photos. 62 and 63). K-feldspar megacrystic monzogranites are also present on Masherbrum. The schists have a K-Ar hornblende age of 77 ± 3 Ma (which is the same as for the Muztagh Tower unit) and the pegmatites have a Rb/Sr isochron age of 14.1 ± 2.1 Ma.

Several plutons occur outside the main granitic belt. In addition to the granitic rocks at Nagar and Aliabad, Sumayar, Bar and Hasanabad already commented upon, two plutons merit mention. The Mango Gusar pluton is a medium-grained stock of two-micas leucogranite. This undeformed pluton cross-cuts syn-metamorphic deformation fabrics and has a U-Pb zircon age of 37 ± 0.8 Ma. The smaller Chingkang-La pluton is a pyroxene-hornblende-sphene-biotite ± muscovite granite with fluorite in miarolitic cavities. This has a K-Ar hornblende age of 34 ± 1 Ma.

A Model for the Karakoram Batholith: The batholith has been the site of magmatism for over 100 Ma since at least the middle of Cretaceous. The concentration of large-scale magmatic activity in a rather narrow and linear belt over a long period of time points to large-scale tectonic control on plutonism. Geochemistry (Figs. 6.34 and 6.35) and isotopic data all point to subduction-related magmas intruded along the southern margin of the Karakoram plate (Searle 1991). This has been attributed to the northward subduction of the Neo-Tethyan lithospheric branch to the north of the Kohistan arc (Jan and Asif 1983, Le Fort et al. 1983). Following the intrusion of the Cretaceous plutons, closure between the

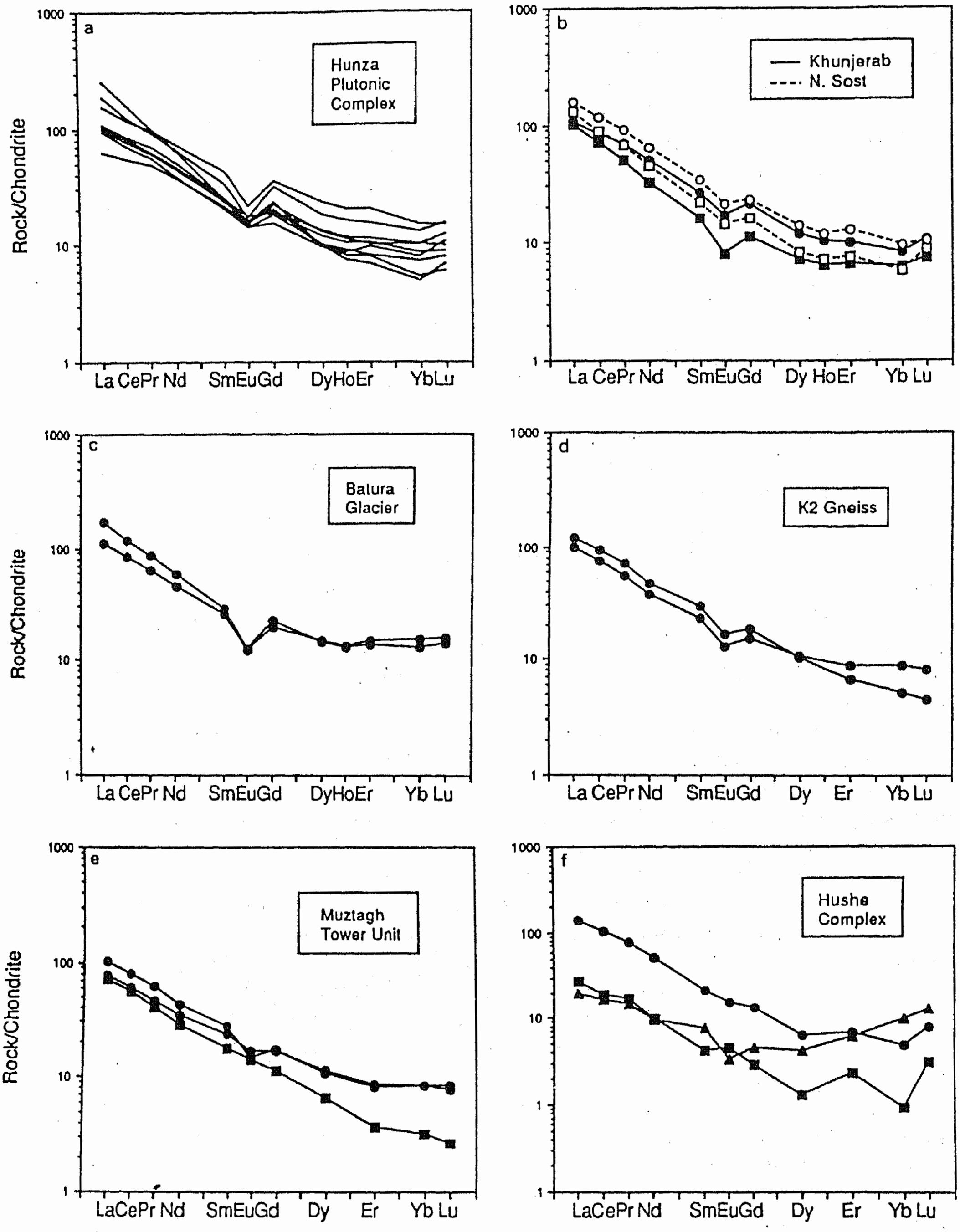


Fig. 6.34. Chondrite-normalised REE patterns for the pre-collisional intrusive units of the central Karakoram. All Hunza samples are quartz diorites or granodiorites. Circles indicate granodioritic lithologies, squares show monzogranites and triangles show leucogranites, although some samples have subsequently been gneissified. (Chondrite value from Taylor and McLennan 1981). After Crawford and Searle 1992.

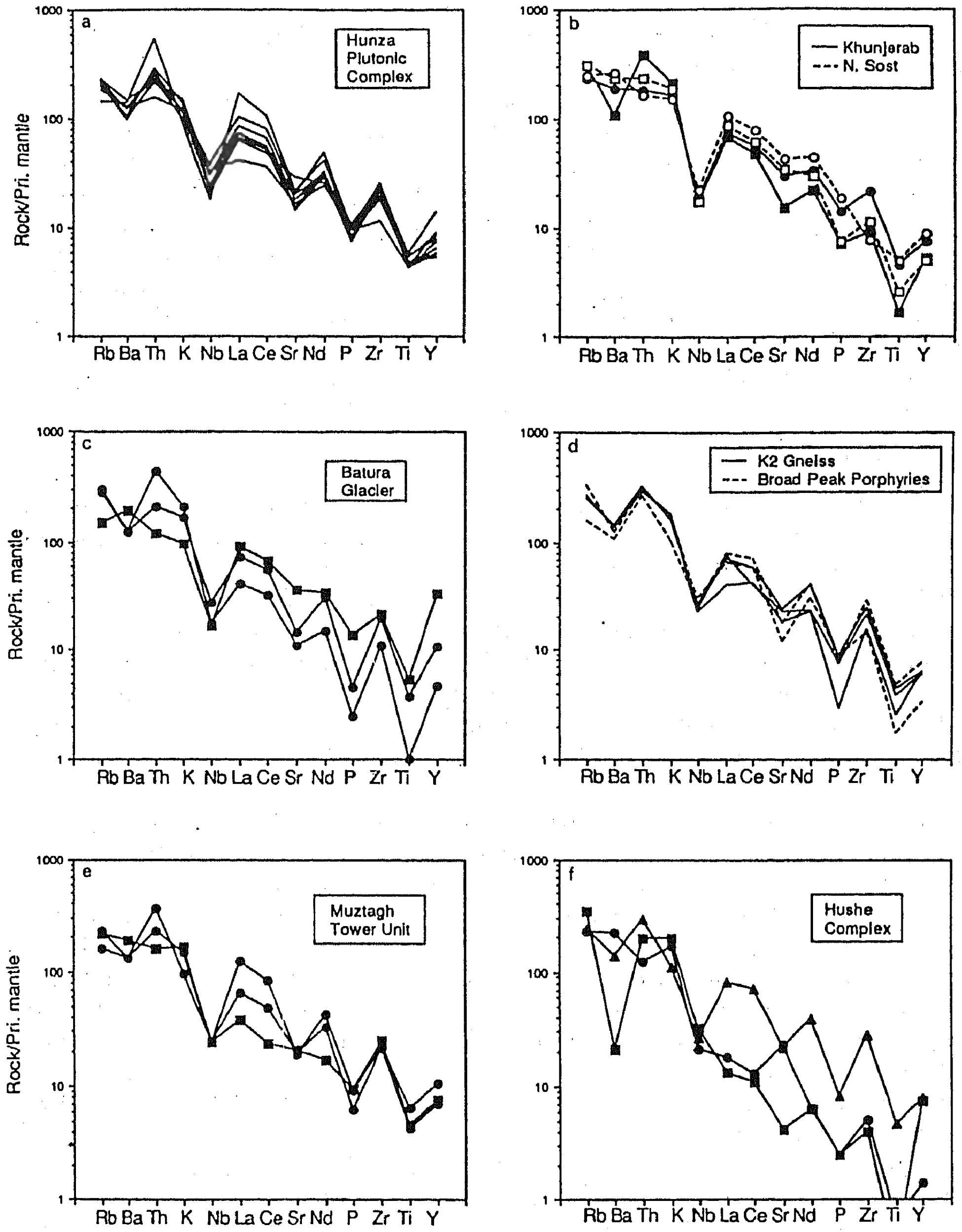


Fig. 6.35. Multi-element variation diagrams for the pre-collisional intrusive units of the central Karakoram, normalised to a primitive mantle composition of Taylor and McLennan (1985). Symbols as in Figure 6.34. (After Crawford and Searle 1992).

Karakoram plate and Kohistan occurred along the Shyok Suture during the Late Cretaceous. This resulted in metamorphism and deformation of the earlier plutons in the batholith. Magma generation continued long after the suturing, resulting in the younger, post-collision granites which are at least partly derived from crustal melts.

Debon et al. (1987) suggested that the plutons may have been emplaced along extensional fractures developed in the over-riding Karakoram plate. Intrusive mechanisms such as hydraulic fracturing along E-W trending normal faults, similar to the Andean Batholith (Pitcher et al. 1985), have been envisaged by Searle (1991). He suggests that during the Early Miocene an E-W striking and N-dipping low-angle normal fault was present in the batholith. This culmination collapse fault was probably related to ramping and uplift of the southern Karakoram along the Shyok Suture during the emplacement of the Baltoro unit. There are several similarities (e.g., ages, petrography, geochemistry) between the Karakoram and Kohistan batholiths, pointing to similar tectono-magmatic processes (Jan et al. 1981b). Further details can be found in Debon et al. (1987, 1996) and Searle (1991).

Tertiary leucogranites, NW Himalayas

Wadia (1933) recorded the occurrence of Tertiary tourmaline-bearing leucogranites in the northwest Himalayas. He noted that these and the associated pegmatites and aplites, despite compositional diversity, commonly contain microcline, oligoclase, quartz, garnet, tourmaline, muscovite, beryl (aquamarine), fluorite, actinolite and corundum. Krishnan (1956) reported that they were intruded mainly during the second and third Himalayan upheaval at the end of the Eocene and in Miocene. Since Wadia, many granitic bodies of the NW Himalayas in Pakistan were placed tentatively in the Tertiary, several of them erroneously. Detailed field and petrographic studies in recent years, supported by some radiometric dates referred to earlier, have shown that Tertiary granites are volumetrically minor in the region to the south of the Indus Suture.

Despite their small volume, considerable attention has been paid over the past 15 years to leucogranite occurrence in orogenic belts. This is because, according to Zeitler and Chamberlain (1991), the mechanism of leucogranite petrogenesis can provide important information about tectonic and metamorphic processes occurring during orogeny (Le Fort et al. 1987), and emplacement history of specific leucogranites can provide key stratigraphic information about the timing, rate, and magnitude of these processes (Hodges et al. 1989). Leucogranites occur throughout the Himalayan orogenic belt in the form of small plutons and dyke swarms into the hanging wall of the Main Central Thrust (MCT) or its structural analogues. They have been reported from Bhutan (Dietrich and Gansser 1981), Nepal (Le Fort 1973, 1981, Ferrara et al. 1983), Zanskar (Searle et al. 1986), and Pakistan (summarised in the following).

Chaudhry and Ghazanfar (1987) and Ghazanfar and Chaudhry (1986) gave detailed accounts of the granitic gneisses of the Kaghan Valley, most of which are Cambrian to Precambrian in age. Associated with these are many small bodies of leucogranites, aplites and pegmatites referred to as crustal anatexites. Some of these are pre-tectonic and, along with their host rocks, have been metamorphosed and deformed (folded) during the Himalayan Orogeny. Others have been considered as syn-, late- and post-tectonic, including garnet-tourmaline-muscovite leucogranite similar to those of the Tertiary age described from other places in the Himalayas. Chaudhry and Ghazanfar (1987) think that these belong to two groups: those associated with the regional Himalayan metamorphism (Paleogene), and those associated with the MCT (Miocene). Zeitler and Chamberlain (1991) reported an approximate 50 Ma U-Pb zircon age for one of the leucogranite dykes about 5 km NE of the MCT and just west of Naran. Occurring in calc-schists of the Indian plate sequence, it consists of plagioclase, alkali feldspar, quartz, biotite, muscovite, tourmaline, zircon and garnet. Four more samples of granite and pegmatite from this area have mean U-Pb zircon ages of 45 to 49 Ma (Smith et al. 1994).

Shams (1969) recognised that the massive to weakly porphyritic tourmaline-bearing granite (with albite, microcline and garnet) is a young (?Tertiary) associate of the Mansehra Granitic Gneisses. The largest outcrop of this granite is about 20 km^2 in area. Some of the Mansehra pegmatites and aplites studied in detail by Ashraf (1992) may also be of this period. Jan et al. (1981b) thought that the undeformed pegmatites in the Besham area may be crustal melts related to Himalayan metamorphism. The leucogranite near Karora is another possible member of the Tertiary granites (Le Fortune et al. 1992). A tourmaline granite, petrographically similar to others of its type, is closely associated with the Ilam Granitic Gneisses in Lower Swat. Together with leucogranite dykes and mm-scale layers in the Manglaur Formation, these have been considered as synmetamorphic melts derived from the rock sequence during the Paleocene (DiPietro 1990).

The Lower Swat sequence was buried under the Kohistan sequence to 35–45 km depth during the Eocene, followed immediately by exhumation (DiPietro 1991). The entire cycle of metamorphism and deformation lasted only 7 to 16 m.y., since initial contact between India and Kohistan during Early Eocene to cooling of the rocks through Ar blocking temperature in hornblende (500 ± 50°C) in Late Eocene (38 Ma) (Maluski and Matt 1984, Zeitler et al. 1989a, Treloar et al. 1989b). DiPietro and Lawrence (1991) regard that 38 Ma is the approximate minimum age of the first three phases of folding. It is worth noting that a dyke of leucogranite from upper amphibolite grade (kyanite-bearing) rocks, 15 km SSE of Saidu, has a U-Pb zircon age of 35 Ma (Zeitler and Chamberlain 1991).

The Himalayan leucogranites clearly represent near-minimum melts of crustal rocks in which there was significant volatile (mainly water) activity, as is obvious from their mineralogy, and highly radiogenic isotopic compositions (Zeitler and Chamberlain 1991). Ferrara et al. (1986) plotted a large number of analyses of leuco- and other Tertiary Himalayan granites on a normative quartz-albite-orthoclase diagram. There is a clustering of the analyses in the low-temperature area of this diagram (Tuttle and Bowen 1958). The genesis of the leucogranites has been explained in several ways. The most widely accepted model for the central Himalayan examples involves fluxing of the hot hanging wall material by volatiles escaping from the dehydrating foot-wall of major fault zones such as the MCT (Le Fort 1981). It is worth noting that the leucogranites in Naran occur in amphibolite grade rocks thrust over low-grade sediments along the MCT (Chaudhry and Ghazanfar 1987, Gerco 1991). In Swat, the leucogranites occur in the upper levels of a nappe-scale duplex zone (Lawrence et al. 1985, DiPietro and Lawrence 1991). There are Late Neogene-Pleistocene leucogranites in the Nanga Parbat Massif. These are described in the following section.

PLIOCENE – PLEISTOCENE

There are at least two areas in Pakistan where igneous rocks of this age are exposed. The Nanga Parbat–Haramosh Massif (NPHM) in western Himalayas contains very young (Plio-Pleistocene) leucogranites. The only other occurrence of such young granites (1.9 to 0.8 Ma), in our knowledge, has been recorded in Japan Alps (Harayama 1992). The Chagai arc in Balochistan contains Pleistocene volcanic (and granitic ?) rocks, including the Koh-i-Sultan group of stratovolcanoes. As will be seen, the two magmatic episodes owe their origin to entirely different processes.

Nanga Parbat–Haramosh Massif (NPHM)

Following overthrusting of the Kohistan arc and consequent Early Tertiary metamorphism, the NPHM experienced a rapid and accelerated unroofing. Ar-Ar and fission-track cooling ages suggest that in the interval between 25 to 10 Ma the massif was cooled at the same rate as neighbouring Kohistan (10–15° C/m.y.). Over the past 10 Ma, however, the massif has experienced a more rapid denudation than Kohistan at mean rates of about 5 mm/year and cooling rates locally exceeding 70° C/m.y. (Zeitler 1985, Zeitler et al. 1993). This unroofing was accomodated initially along ductile shearing and later along cataclastic faulting on its margins, with local strike-slip motion along the Shahbatot Fault. The Raikot or Liachar Fault on its western margin carried the massif back over Kohistan and Quaternary sediments (Lawrence and Ghauri 1983, Butler and Prior 1988, Zeitler and Chamberlain 1991, Smith et al. 1992).

Migmatites containing up to 5 cm thick stringers of granite (partial melting products) are abundant in the core of the massif (Zeitler et al. 1992). Leucogranite dykes and pegmatite are common in the western part of the massif. They are up to 2 m in thickness, but rarely form kilometre-sized stocks. They cross-cut the dominant metamorphic foliation in the massif and the shear fabric associated with the late faulting on the edge of the massif. Although undeformed in hand specimens, they show deformation textures in thin sections.

The pelitic rocks in the massif consist of the assemblage sillimanite-potash feldspar-cordierite formed at 650 ± 50° C, 6 ± 1 kbar. The leucogranites consist of typical S-type anatectic assemblage of two feldspars-quartz-two micas-sillimanite-cordierite-garnet-tourmaline-apatite-zircon-monazite formed at 600° C, 4.1 ± 1 kbar. Zeitler et al. (1993) suggest that the NPHM has undergone a recent metamorphic episode culminating in extensive partial melting (under water-under-saturated environments), as recently as about 1.0 Ma. The rapid denudation is considered to be at least partly responsible for initiating decompression melting and high-grade metamorphism.

Young leucogranites occur throughout the massif. Tourmaline-bearing leucogranites form abundant dykes and sheets, possibly representing stock-works above a large leucogranite intrusion in the northern part of the massif (Butler etal. 1992). There also is a coarse-grained homogeneous body (Jutial leucogranite), at least 3 km in diameter, exposed on the east side of the Phurparash Valley. Considered to have formed at about 700° C, it consists of near minimum-temperature melt composition of equal proportions of quartz, oligoclase and perthite, with two micas, zircon and apatite. Tourmaline, where present, is skeletal and quartz grains may show weak orientation. Sheets of similar leucogranite, 1–2 m thick, occur in the neighbouring host rocks, together with a muscovite-tourmaline facies that rarely contains biotite and garnet.

Detailed geochemical data are available on these northern leucogranites (Butler et al. 1992). Like leucogranites in the rest of the Himalaya, these have a distinct geochemistry: high (>70%) SiO_2, and alkalies (particularly K and Rb), and depleted Ca, Sr, Y, and Zr. These characteristics, combined with very high Sr^{87}/Sr^{86} ratios (>0.88), can be attributed to fractional melting of a sedimentary source. The peraluminous nature of the analyses and Rb/Sr ratios of 2 to 8 are consistent with vapour absent melting of metasediments involving incongruent melting of muscovite. The leucogranites have high Th content; double of the reported for other leucogranites from the Himalaya. Butler et al. (1992) regard that this reflects a similarly high Th content (and hence heat production) in the protolith. Enhanced internal heat in the protolith gneisses will enhance the rate of melt production.

Before closing this section, it may be appropriate to comment briefly on the very young granodiorite in Japan Alps. This forms a 13 × 4 km pluton that intrudes Late Pliocene (4.2 Ma) volcanics (Harayama 1992). Investigation of the thermal history of the pluton indicates high thermal input from deeper levels in early stages (Late Pliocene) and cooling through rapid uplift (2,000 m) during the Quaternary.

Chagai arc, Balochistan

Since the early Miocene emplacement of calc-alkaline plutons, the Chagai arc, apparently, did not experience any magmatism until the Late Pliocene and Pleistocene. The renewed magmatism does not coincide with the main Chagai arc; the Quaternary volcanic centres, instead, occur on its western margin. Farhoudi and Karig (1977) related this magmatism to subduction under Makran. According to Arthurton et al. (1979), the long quiescence in the magmatism is difficult to explain if it is assumed that subduction of the oceanic lithosphere of the Arabian plate continued unabated during the Neogene. Periodicity in volcanism and migration of the volcanic front, however, have been recorded in other andesitic arcs also, e.g., Central America (Carr et al. 1982) and Cascade Range (McBirney and White 1982). In central Oregon Cascades, there have been pulses of volcanic activity with a 5 Ma periodicity (McBirney et al. 1974). Interestingly, it has been suggested that the Lut and Afghan microcontinents collided during the Late Oligocene to become one block (Stocklin and Nabavi 1973, Arthurton et al. 1979). The Quaternary magmatism may, therefore, be related to a different tectonic configuration than that operating during the development of the main Chagai–Kandahar arc(s).

In Chagai the Quaternary magmatism appears to consist of three distinct groups of rocks:

(a) Small quartz diorite bodies intruding the nearly horizontal Pleistocene beds of the Kamerod Formation,

(b) Dykes of fresh olivine basalt in the Kamerod Formation. Two small, pipe-like dykes of fresh basalt occurring 3.3 km NW of Saindak Fort merit special mention. These consist of abundant phenocrysts of augite and biotite in a matrix containing abundant granules of magnetite (Ahmed et al. 1972). This description suits lamprophyre (Rock 1991).

(c) Several centres of andestic volcanoes.

Using aerial photographs and satellite images, Well (1966), and Jacob and Quitmeyer (1979) identified many volcanogenic features, including 16 eruptive centres. The flows from the young eruptive centres are locally stratified with the older alluvium. Present activity in some of these centres (e.g., Taftan, Koh-i-Sultan) is in the form of fumaroles, solfotaras, and hot springs, depositing calc tufa and aragonite (Ahmed 1953, Gansser 1971). The native sulphur deposits close to the summit of Koh-i-Sultan Volcano may represent the effluent from a recently active copper system at depth (Sillitoe 1978). The main activity in Koh-i-Sultan was directed on at least four centres from which andesite and pyroclastic debris were ejected, forming coalesced cones (HSC 1960).

A similar set of rocks characterised the Quaternary magmatism in the adjacent Iranian Balochistan (Gansser 1971). The Taftan and Bazman group of volcanic rocks comprise andesitic (to dacitic ?) flows and pyroclastic material. Some minor centres and satellite volcanoes have erupted olivine basalt. The Bazman Volcano has an underlying young granite similar to the Zahidan

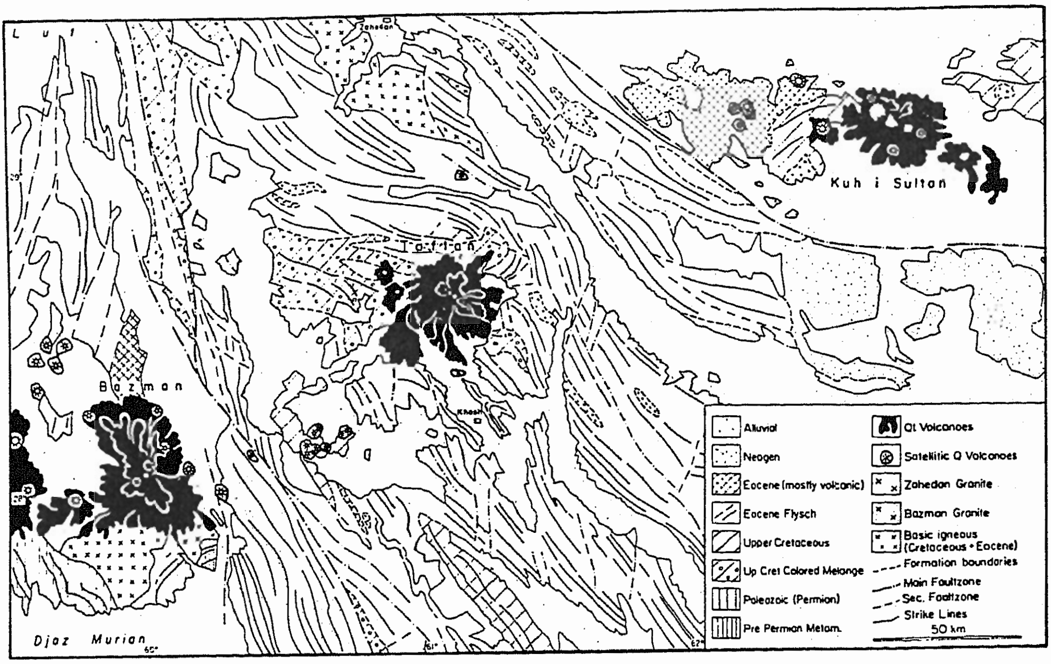


Fig. 6.36. Geological map of the Balochistan volcanoes (Late Neogene–Quaternary) in SE Pakistan. (From Gansser 1971).

Granite. The Bazman Granite is pink and contains abundant pegmatite and aplite. To the northwest it is intruded by biotite dacite containing devitrified glass. The dacitic rocks are covered by andesitic flows and tuffs of the Bazman Valcano with satellite olivine basalt on its flank (Fig. 6.36).

The Quaternary andesitic arc displays the followng characteristics: (1) it extends ENE-WSW for about 300 km and comprises over a dozen centres, (2) the arc seems to bear no relation to the major pre-volcanic structures in the region; the volcanic arc, in fact crosses several older structures (Gansser 1971), (3) the volcanic arc coincides with the southern margin of the combined Lut–Afghan block, and (4) north of the Makran Ranges there are large subsiding fore-arc basins (Jazz Murian and Mashkhel), flanked to the north by the Quaternary arc (Farhudi and Karig 1977). We support the generally held view that the Quaternary magmatism is related to subduction of the Arabian plate under the Makran region of Balochistan and is a continuation of the Chagai arc magmatism. The Middle Miocene and Pliocene pause in the magmatism of the Chagai arc, probably, is connected with a new set of conditions in the continental margin (i.e., Late Oligocene closure between the Iranian and Afghan microplates) under which the Arabian plate was subducting.

7

Metamorphism

Pakistan is a collage of several terranes each of which has had its own tectonic, magmatic, and metamorphic history. As already discussed, Pakistan is essentially made up of (1) the northern margin of the Indian plate, Tethyan sediments and Himalayan foreland basin molasse, (2) one or more Central Asian microcontinents in its north and northwest, (3) a series of island arcs and ophiolites between (1) and (2), and (4) the Makran accretionary prism. In western Pakistan, the boundary between the Indian plate to the east and the Chagai magmatic arc and Ras Koh Range to the west is demarcated by a series of large ophiolites emplaced during the Paleocene. In northern Pakistan the Indian plate is separated from the Karakoram plate by the Kohistan magmatic arc which was welded to the Karakoram plate during the Late Cretaceous (85–95 Ma) and to the Indian plate during Paleocene (65–55 Ma). Collisional processes have led to episodes of magmatism, deformation, metamorphism and exhumation. The Indus Suture exposes deep oceanic sediments and igneous rocks, blueschists and melanges with a shaley matrix. Locally, however, the suture lithologies consist of fault-bounded blocks and slices of the Kohistan arc and Indian plate.

The Himalayas are known as a prime example of continent-continent collision, i.e., the result of the closure of Tethys and indentation of the Indian plate into the collage of terranes which made up Central Asia (Searle and Treloar 1993). This region provides unique opportunities to integrate metamorphic history and pressure-temperature-time trajectories with the tectonic development of island arcs, Andean-type margins, sutures and overthrust belts, and with the cooling-uplift rates, the erosional history, sedimentation and subsidence in the foreland basins (Windley 1983). Such integrated studies have just been started in selected areas of northern Pakistan; however, large areas of the western part of the country are poorly understood due to a lack of accessibility and hostile terrane conditions.

We describe here the broader aspects of metamorphism in various units in the following arbitrary order: Chagai and Ras Koh, the ophiolitic thrust belt and Indus Suture zone, Karakoram plate, Kohistan arc, and Indian plate. The Chagai and Ras Koh areas have undergone low-grade metamorphism, but record several episodes of deformation and uplift related to interaction between different tectonic blocks. The ophiolites also show low-grade metamorphism, whereas the Indus Suture melanges contain high-P rocks. The Karakoram plate displays a prolonged and complex metamorphic history starting before the Shyok Suture (Searle and Treloar 1993, Barnicoat and Treloar 1989), and the tectonometamorphic evolution of the Kohistan arc can be related to the Shyok and Indus Sutures. The Chilas Complex may have supplied sufficient thermal energy to drive regional metamorphism in Kohistan. Metamorphism in the northwestern margin of India occurred soon after, and in response to, thrusting of the Kohistan arc over the Indian plate. But there is evidence of Neogene metamorphism related to rapid uplift in the Nanga Parbat region, as well as of one or more episodes of pre-Himalayan metamorphism.

CHAGAI AND RAS KOH

The Chagai Ranges constitute an E to W-trending belt of calc-alkaline volcanic and plutonic rocks, and sediments of Cretaceous and Tertiary ages. It is generally thought that these rocks represent an island arc developed on the southern margin of the Afghan block due to northward subduction of an oceanic lithosphere. The Ras Koh Range is a NE to SW-trending tectonic block of Late Cretaceous and Early Tertiary volcanic, plutonic, and sedimentary rocks to the south of the Chagai magmatic arc. It apparently consists of rocks of diverse tectonic regimes. The northeastern part of Ras Koh contains Cretaceous rocks similar to those of Chagai (i.e., Sinjrani Volcanics), but elsewhere there is a bimodal basalt-rhyolite association of probable intra-plate setting, more likely continental. In the north and west of Ras Koh are ophiolitic bodies of possible oceanic affinity. These may have been obducted during collision when the Chagai arc was thrust onto the Ras Koh Range.

The two terranes have been described in detail in Chapter 6. It appears that the Cretaceous volcanic and plutonic rocks in this region have been metamorphosed in greenschist facies; chlorite, epidote, albite, actinolite and mica are common, as is serpentine in ultramafic rocks. The timing (and hence mechanism) of metamorphism is not known, but it may be very Late Cretaceous or Early Paleocene because post-Paleocene volcanics do not display the pervasive "alteration" so characteristic of the Sinjrani Formation. The Chagai arc underwent a major episode of deformation (folding and faulting), uplift and erosion before the deposition of the Paleocene-Eocene sediments. It is possible that this tectono-metamorphic phase (including the obduction of the ophiolites) was associated with collision between the Chagai arc and Ras Koh block. Arthurton et al. (1979) reported other episodes of deformation, uplift and denudation (see Chapter 6). A major one of these occurred at Oligocene-Miocene boundary when continental conditions lasting to the present day were established, and many calc-alkaline plutons were emplaced. Unfortunately, no information exists on the relationships of these to metamorphism. McCormick (1989) has reported wedges of schists and hornblendite produced by metamorphism induced by the emplacement of ophiolites in the Ras Koh.

A portion of the Chagai arc is dragged to the northeast by oroclinal flexing into the Chaman transform zone near Spinatizha (Lawrence et al. 1981). The calc-alkaline volcanic and plutonic rocks of the island arc here rest upon a metamorphic complex that may be the pre-Cretaceous basement of the arc. The complex comprises meta-igneous rocks, both volcanic and plutonic. The former consist of pyroclastic material and flows with some marble. These rocks display low-grade metamorphism, shearing and tectonic foliation. Much of the deformation occurred during metamorphism along discrete, probably deep crustal, shear zones. A typical metavolcanic assemblage consists of quartz, albite, actinolite, epidote, opaques ± chlorite ± biotite ± muscovite ± prehnite ± sphene ± calcite.

The meta-intrusive rocks are predominantly granite and granodiorite orthogneisses. These bodies are elongated in the direction of foliation of the host meta-volcanics and are cataclasised, with plagioclase augen surrounded by ribbon quartz. Such textures do not occur in the intrusions post-dating the Sinjrani Volcanics. Lawrence et al. (1981) suggest that the Spinatizha rocks may be an older portion, and the roots, of the volcanic arc brought to the surface by vertical motion adjacent to this segment of the Chaman Fault.

BELA – ZHOB – WAZIRISTAN OPHIOLITES

Apart from petrography summarised in Chapter 6, little data are available on the metamorphic petrology of these complexes. Lithologies which characterise a typical ophiolite are present, but tectonic processes have commonly disrupted the stratigraphic sequence; they may better be termed as mega-melanges. The ophiolites display greenschist and low-grade amphibolite facies assemblages developed locally or extensively: chlorite, actinolite, epidote, sodic plagioclase ± carbonate ± quartz after mafic volcanic and plutonic rocks, and serpentine ± carbonate ± talc ± chlorite after ultramafic rocks. It is not clear whether these parageneses are related to sea-floor emplacement or post-emplacement metamorphic regimes. Information is also lacking on the nature of metamorphism in the residual peridotites (i.e., metamorphic tectonites of Coleman 1977).

In Bela Ophiolite, the volcanic rocks and gabbros are spilitised and there are beds of serpentinites ± ophicalcite debris overlain by unmetamorphosed sedimentary rocks (Sarwar 1992). These may, therefore, have undergone sea-floor, metamorphism. Extensive and intensive alteration in volcanic (hosting promising copper mineralisation) and some plutonic rocks in the Waziristan Ophiolite may also have a similar origin. In the Khost area of Afghanistan, adjoining Waziristan, there are glaucophane-schists representing the accretionary prism. These have been thrust onto the Indian plate during Paleocene (Beck et al. 1995).

The Muslimbagh Ophiolite also displays greenschist parageneses, but Sawada et al. (1992) have reported a more complicated picture. They regard that the sheeted dyke complex in Saplai Tor Ghar consists of dolerite, gabbro-diorite and plagiogranite emplaced in three stages. This Complex displays high-grade amphibolite to greenschist facies metamorphism and strong to weak foliation. The plagiogranite is considered as a product of wet melting of the dolerites, but no petrogenetic details have been published.

Ahmad and Abbas (1979) described slices of metamorphic rocks beneath the north end of the Jang Tor Ghar and along the west side of the Saplai Tor Ghar massif. They noted an inverted metamorphic sequence decreasing from amphibolite facies adjacent to the massifs to greenschist facies away from the contact but at greater depth. McCormick (1989) reported a well-developed sequence of metamorphic rocks 300 m thick adjacent to the ophiolite: hornblendite at the contact, followed by garnet gneiss, chlorite-garnet schist, marble, garnet (? mica) schist and, finally, two-mica schist. The chlorite schist and/or two-mica schist occur at several other places in the complex. McCormick (1989) thinks that the aluminous rocks and marble may be flyschoid sediments and the hornblendite may be a metamorphosed volcanic rock. Williams and Smyth (1973) reported such metamorphic zones in the Newfoundland ophiolite and suggested that graywackes were metamorphosed by heat of fusion associated with the emplacement of the ophiolite. Ahmad and Abbas (1979) also proposed frictional heat or residual heat associated with the Muslimbagh ophiolite for metamorphism in its sole.

INDUS SUTURE

For simplicity, the Indus Suture zone here is regarded as consisting of tectonic melanges trapped between the Kohistan arc and the Indian plate. The melanges in Mohmand–Malakand, Mingora–Shangla and Allai areas consist of plutonic, volcanic and sedimentary rocks of the Neo-Tethys, together with some blocks torn from

Kohistan and Indian plate. In the Babusar–Nanga Parbat area, the melanges consist essentially of rocks belonging to India and Kohistan with little oceanic material sensu stricto. In this section we deal with the ophiolitic material; the Kohistan and Indian plate rocks are described in their respective sections.

Malakand – Mohmand Ophiolites and Melange

The large ophiolite klippe, stretching from Dargai to Utmankhel for 30 km, is dismembered and consists predominantly of ultramafic rocks with a subordinate amount of mafic cumulates and dolerite dykes (Ahmed 1984). The neighbouring melange contains extensive outcrops of metavolcanic rocks and associated sediments (Hussain et al. 1984). Volcanic rocks and mafic-ultramafic lenses occur further west, together with manganiferous sediments, in Bajaur. Petrographic data suggest that these rocks have been affected by widespread low-grade metamorphism.

The volcanic rocks consist of chlorite, epidote, quartz, albite ± amphibole ± white mica; pelites comprise quartz, white mica ± chlorite ± sodic plagioclase ± garnet; and manganiferous sediments are converted to piemontite schist. The mafic plutonic rocks may display the development of saussurite, chlorite and fibrous amphibole (uralite), and locally host rodingite (Qaiser et al. 1970, Ahmed 1988). The ultramafic rocks are more or less serpentinised (± chlorite ± carbonate ± talc). In places, especially along shear zones, they are metamorphosed to talc-tremolite and talc-carbonate ± quartz ± fuchsite ± epidote, with local occurrences of green beryl and emerald. Chromites display alternation to ferritchromite, magnetite and chlorite. In the greenstones of Prang Ghar, Rafiq and Jan (1991) reported a 200 × 2 m lens of an unusual rock consisting of abundant chloritoid (up to 55%), Fe-Ti oxides (up to 28%), quartz, white mica, and traces of biotite, rutile, tourmaline and apatite. Some of the neighbouring greenschists are also chloritoid-bearing.

Studies on the nature and conditions of metamorphism are lacking. The mineral assemblages, however, point to greenschist to low-grade amphibolite facies temperatures and moderately high pressures. The metamorphism was probably related to the collision of Kohistan with Indian plate during Early Paleocene. The ultramafic tectonites show a steep metamorphic banding superimposed upon a gentle magmatic layering (Rossman et al. 1970).

Mingora – Shangla – Allai Melanges

A series of tectonic melanges is associated with the Indus Suture in these areas, and in Shamozai to the west of the Swat River. The Mingora–Shangla melange zone has been divided into blueschist, greenschist and ophiolitic melange units (Kazmi et al. 1984). These rocks were considered by Jan et al. (1981d) as oceanic trench assemblage. The blueschist melange consists of blocks of metavolcanics, serpentinite, metagabbro/dolerite, metgraywacke, marble, metachert and piemontite schists in a matrix of phyllitic rocks (Jan 1991, Jan and Symes 1977). The petrography of the rocks has been presented by Shams (1972, 1980), Jan et al. (1981d) and Kazmi et al. (1984). Glaucophane or crossite occurs in blueschists, metagraywackes and, rarely, pure calcite rocks. Phengite is widespread and jadeitic pyroxene occurs locally (Guiraud, personal communication). Lawsonite has not been found, but in Malam–Jaba a rare metabasite contains rhombic porphyroblasts of calcite possibly pseudomorphing lawsonite.

There is little information on the other melange units but the 7 × 1 km lensoid ultramafic body near Shangla has been studied in considerable detail (see Chapter 6). This body is strongly serpentinised and on its eastern margin altered to dolomite, talc and/or quartz, with local development of chromian tourmaline and emerald (Jan et al. 1972, Kazmi et al. 1986). Chromites are marginally altered to ferritchromite ± chlorite ± magnetite. The olivine in some peridotites has been metamorphosed to nearly pure forsterite containing blades of iron oxide (Arif and Jan 1993). Some lenses of ultramafic rocks to the west of Shangla consist of serpentine, talc-carbonate or talc-tremolite.

The melange zone in Allai, 30 km NE of Shangla, is composed of ultramafic rocks, metagabbros, greenstones, greenschists, metagraywacke, phyllite, limestone, metachert and blueschist. There also are blocks of metasediments and amphibolites derived from the Indian plate and Kohistan arc, respectively. This range of lithologies is similar to that of Mingora–Shangla and, like the latter, it has also been divided into ophiolite, blueschist and greenschist melange units. Microprobe data on a sample of graywacke show that the phengite contains 8–10% FeO + MgO whereas the alkali amphibole consists of glaucophane with crossite margins (Majid and Shah 1985). Baig (1989) noted that glaucophane in his samples was, instead, surrounded by actinolite. More recently, Hamidullah et al. (1991) reported a much greater range of amphibole compositions in the greenschists and blueschists.

The melange rocks in the two areas contain the following diagnostic parageneses:

A) Mafic to felsic rocks
  1) glaucophane-chlorite-albite-quartz ± epidote ± calcite ± garnet ± phengite (blueschists)
  2) chlorite-epidote-albite-quartz ± phengite ± calcic amphibole ± biotite ± sphene (greenschists and metagabbros)
  3) albite-quartz-actinolite-hornblende-epidote (plagiogranites)
  4) albite-quartz-blue-green amphibole (albitite)

B) Metasediments
  5) quartz-phengite ± chlorite ± plagioclase ± amphibole ± calcite (metapelite)
  6) quartz-glaucophane ± phengite (metachert)
  7) quartz-phengite-chlorite-glaucophane-plagioclase-epidote-carbonate (metagraywacke)
  8) quartz-albite-white mica-piemontite ± margarite ± chlorite ± spessartine ± rutile ± magnetite (manganiferous)

C) Ultramafic rocks
  9) serpentine (chrysotile, lizardite, antigorite) ± magnetite ± carbonate ± talc
  10) antigorite-forsterite-diopside-magnetite
  11) talc-tremolite ± chlorite
  12) talc-dolomite ± quartz ± fuchsite ± chlorite ± epidote
  13) dolomite-quartz

Detailed petrological studies are lacking, but Jan et al. (1981d) suggested that the blueschists were metamorphosed at about 410° C and 7 kbar. Similar estimates (380–450° C, 7–8 kbar) have been proposed for those of Allai (Baig 1989). A greenschist facies overprinting at lower pressure (and elevated temperature ?) affected the blueschists. In addition to local growth of garnet and calcic amphibole, some rocks show extensive passage to greenschists. This is entirely consistent with the structural analysis of the blueschists: glaucophane is associated with an earlier and actinolite rims on glaucophane with a later fabric (Baig 1989).

Ar-Ar and K-Ar geochronology has yielded ages of 100 ± 20 and 67 ± 12 Ma for amphibole (Maluski and Shaeffer 1982), and 83.5 ± 2 (Maluski and Matte 1984) and 84 ± 1.7 Ma (Shams 1980) for white mica of Shangla blueschists. These ages are similar to those reported for the blueschists in the Indus Suture zone in Ladakh (Honegger et al. 1989). The high-P metamorphism in the suture has been regarded by Jan (1991) as related to Cretaceous subduction. It is interesting to note that the radiometric dates in Shangla and Ladakh coincide with the rapid northward movement (15 cm/year) of India following its breakup from Gondwana in Mid - Cretaceous (Windley 1984). Exhumation must have quickly followed, as suggested by the Early Paleocene collision of Kohistan with India (Beck et al. 1995). During uplift, lower pressure assemblages were overprinted on the blueschists of Pakistan at 4–5 kbar, a situation quite similar to that noted in eastern Ladakh (Virdi 1981, Jan 1987). It is important to note that uplift-cooling ages of about 85 Ma in the southern part of Kohistan (Treloar et al. 1989) coincide with the radiometric ages on Shangla blueschists. Therefore, there is a possibility that the high-P metamorphism may be related to thrusting of the Kohistan arc over the Neo-Tethys before collision with India.

It is likely that much of the Indus Suture melange formed during the Paleocene collision and obduction. This opinion is supported by the presence of blocks belonging not only to the Neo-Tethyan regime but also to India and Kohistan. Differences in grade (and style ?) of metamorphism in different units and blocks are in agreement with this view. The slaty to phyllitic matrix in Shangla has not been studied, but probably would reveal the lowest grade of metamorphism in the entire melange. The Shangla serpentinised peridotites contain talc, antigorite, and forsterite + iron oxide formed from a less forsteritic olivine. These rocks have suffered low- to (?) medium-grade metamorphism, possibly during obduction (Arif and Jan 1993). Barbieri et al. (1994) have suggested that the assemblages and textures in the gabbroic rocks may be suggestive of greenschist-epidote amphibolite facies sub-seafloor metamorphism (350–>500° C, 2 kbar).

KARAKORAM BLOCK

The Karakoram block here is defined as the assembly of rocks located between the Rushan–Pshart Suture in the north and the Shyok Suture in the south. It consists of the north Karakoram region or Northern Karakoram terrane of Gaetani et al. (1990), south-east Pamir, and eastern Hindu Kush. As mentioned earlier, the Karakoram plate in northern Pakistan is divided into four units: (1) northern Chitral, (2) Karakoram 'sedimentary' zone north of the Karakoram Batholith, (3) Karakoram Batholith, and (4) Karakoram metamorphic complex south of the batholith. Several stratigraphic units of Chitral and the area to the east have tentatively been correlated by Pudsey et al. (1985) and Gaetani et al. (1993), however, a lack of detailed studies in this structurally complex region does not permit us to draw analogies between metamorphism of the two areas.

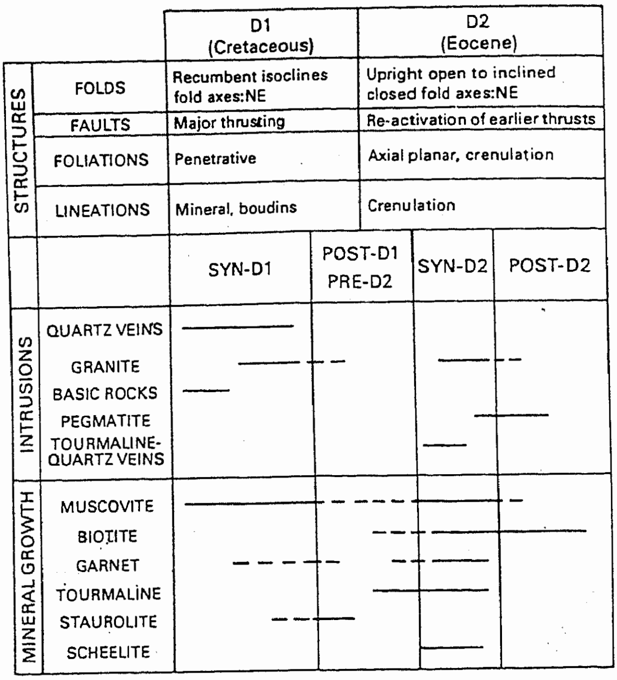
Northern Chitral

The Chitral area to the north of Shyok Suture can be divided into two geological domains, one to the south and the other to the north of the Reshun Fault (Desio 1975, Calkins et al. 1981, Pudsey et al. 1985). The southern domain consists of the Chitral Slate, Koghozi Greenschist, Krinj and Gahiret limestones, Reshun Formation, and Karakoram Batholith. The Chitral Slate (? Permian to Cretaceous) contains illite, chlorite, quartz and some albite, and is structurally complex. The Koghozi Greenschist (Jurassic ?) consists of chlorite, epidote, quartz and albite, possibly derived from tuffs. The Krinj and Gahiret limestones are Cretaceous, the latter being recrystallised and containing graphite and traces of phlogopite. The Reshun Formation (? Late Cretaceous) comprises conglomerates, red shales and micritic limestone showing little deformation.

The northern domain consists of granitic plutons and Devonian to Permian limestones, quartzites and shales designated by different names (see Pudsey et al. 1985). These rocks are traversed by several thrusts and show at least two phases of deformation. Leake et al. (1989) have divided these rocks in the Tirich Mir area into three units separated by faults. Adjacent to Reshun Fault occurs a narrow wedge of greenschists, micritic limestones, ferruginous dolomitic carbonate, cherts, phyllites and breccias of Sewakht Formation. It may correlate with the Devonian Shogram Formation (limestones and shales) of the Mastuj Valley to the east (Desio 1966). This is followed to the northwest by the monotonous phyllites of Lutkho Formation. This unit shows an increase in metamorphic grade towards the Tirich Fault on its northwest. This is followed by Arkari Formation (Lun Shales of Desio 1975, Pudsey et al. 1985), with subordinate quartzite and a prominent marble member. In the southwest of the area, the Arkari Formation has been metamorphosed to amphibolite facies.

Metamorphic assemblages in the Arkari Formation have been studied in some detail by Leake et al. (1989). Phyllites consist of quartz, muscovite, biotite or chlorite ± garnet ± staurolite. Pelitic schists contain quartz, feldspar, muscovite, biotite, together with variable porphyroblastic assemblage of garnet, staurolite, kyanite, cordierite, andalusite and tourmaline. Marbles may contain accessory amounts of micaceous minerals and quartz, and calc-silicate rocks consist of calcite, quartz, muscovite, chlorite, biotite/phlogopite, clinozoisite, tourmaline, titanite and zircon. Locally developed skarns are made up of quartz, calcite, actinolite, clinozoisite, muscovite, titanite, idocrase and anorthite. Quartzites, frequently containing scheelite, consist of a variety of minerals, depending upon bulk composition of the beds: quartz, muscovite, clinozoisite, chlorite, amphibole, garnet, anorthite, titanite, and several ore minerals. There also are feldspar-quartz-muscovite-clinozoisite-sphene-biotite gneisses and tourmaline-rich schists and gneisses.

Structural analysis of the area to the southwest of Tirich Mir (Leake et al. 1989) reveals two phases of deformation. A greenschist facies metamorphic event was associated with isoclinal D_1 folding. The metamorphic grade increased towards the end of D_1 and reached its maximum after D_1 when staurolite crystallised. During the D_2 event, two micas and garnet continued to crystallise together with clinozoisite. Post-D_2 crystallisation is restricted to biotite and amphibole. Figure 7.1 summarises the tectono-metamorphic history of the area. Metamorphism is apparently Barrovian-type (Leake et al. 1989), and shows an increase towards the southwest. The occurrence in pelitic rocks of kyanite on the one hand and cordierite and andalusite on the other has not been elucidated. It is possible that the low-pressure minerals were overprinted on Barrovian-type parageneses during a thermal event related to granitic plutons. In fact the granite plutons may have provided the thermal energy during regional metamorphism. Leake et al. (1989), for example, noted that the highest grade rocks in the southwest of the area were intruded by numerous leucogranite stocks and sills.



		D1 (Cretaceous)	D2 (Eocene)
STRUCTURES	FOLDS	Recumbent isoclines fold axes:NE	Upright open to inclined closed fold axes:NE
	FAULTS	Major thrusting	Re-activation of earlier thrusts
	FOLIATIONS	Penetrative	Axial planar, crenulation
	LINEATIONS	Mineral, boudins	Crenulation

Fig. 7.1. Summary of tectonic, intrusive, metamorphic and mineralisation history of the Miniki Gol area, Chitral (from Leake et al. 1989).

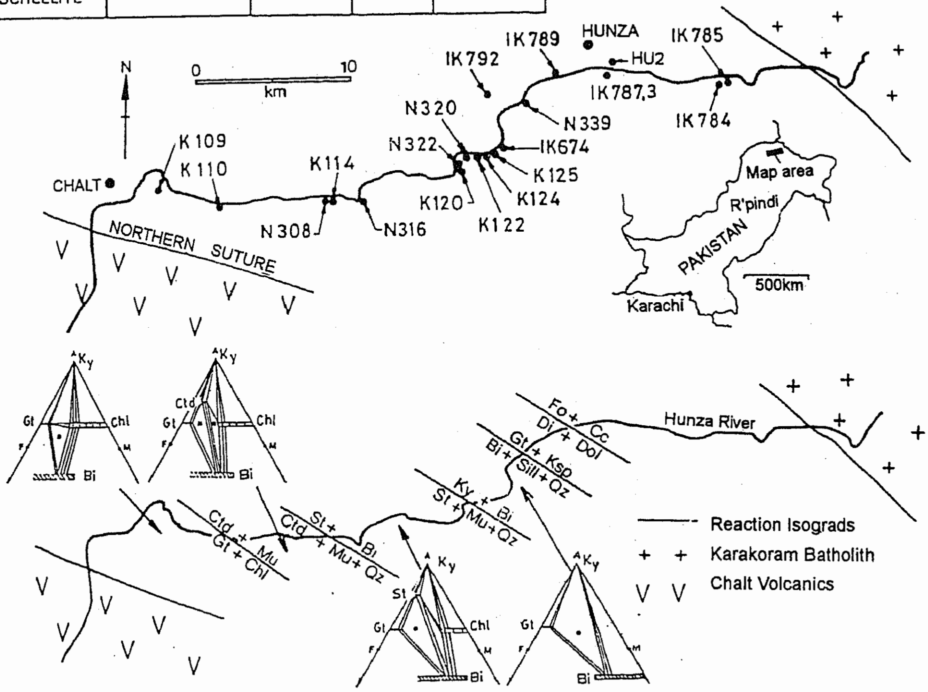


Fig. 7.2. Inverted reaction isograds that dip northwards on the southern margin of the Karakoram plate between the Shyok Suture at Chalt and the Karakoram Batholith at Hunza. Fo=forsterite, Cc=calcite, Di=diopside, Dol-dolomite, Gt=garnet, Ksp=K-feldspar, Bi=biotite, Sill=sillimanite, Qz=Quartz, Ky=Kyanite, St=staurolite,Mu=Muscovite, Ctd=chloritoid, Chl=chlorite. From Broughton et al. (1985).

Sedimentary zone north of the Karakoram Batholith

In the north-central Karakoram of Pakistan, three main structural belts have been recognised by Gaetani et al. (1990a,b, 1993). South to north they consist of Guhjal, Sost and Misgar units (Fig. 5.4). The Guhjal unit, immediately north of the Karakoram Batholith, consists of dolomite to calcareous marbles interbedded with dark slates and phyllites (see Chapters 4 and 5). A SW-dipping reverse fault places the unit above the Sost belt (Zanchi 1993). The Sost unit is composed of an antiformal stack of Permo-Cretaceous sediments. The central part of this unit is cut through by large E-W-trending sinistral strike-slip faults. Structurally overlying the Sost unit is Misgar unit (the Misgar Slates of Desio and Martina 1972). This is a monotonous sequence of dark slates deformed at low metamorphic grade. The magmatic rocks of the northern part of the sedimentary zone have been discussed in Chapter 6.

The sedimentary rocks appear to have passed through a very complex and polyphase structural evolution (Table 4.8). Several phases of folding, thrust imbrication, and late-stage extensive wrench-tectonics are the most distinctive characteristics of the structural framework (Zanchi 1993). At least two major episodes of deformation can be identified in the Mesozoic sedimentary succession of the area (Gaetani et al. 1993). The older event took place between Lias and early Mid-Jurassic. Some of the Jurassic rocks (Ashtigar Formation) indicate deposition in a collisional basin close to a newly-formed suture belt. It is possible that small, shallow seas existed in the area before Barremian. The second deformation occurred during the Cretaceous and is interpreted as related to the final welding of the Kohistan arc and Karakoram microplate to Asia, but it could also be related to the (? southward: Jan and Asif 1983) subduction responsible for magmatism in the Khunjerab–Tirich Mir granitic belt.

The rocks generally display very low-grade metamorphism. Pelitic lithologies in the three units show slaty cleavage, re-orientation of fine-grained sericite and recrystallisation of quartz. Locally, the rocks have suffered a higher grade of metamorphism. The meta-sediments between the Shimshal Pass and Shuijerab, for example, are intensively transpressed, with axial plane foliations defined by biotite and muscovite, and local occurrence of post-kinematic garnet. Folding of previous foliation and syn- to post-kinematic andalusite poikiloblasts are apparent in the contact aureole of the Shimshal Pass granodiorite (Zanchi 1993). Some of the rocks here also contain epidote and chloritoid. Along the northern bank of the Baltoro Glacier, the Passu Slates (Carboniferous) became progressively more metamorphosed and contain chlorite, biotite, local garnet and andalusite. The folds are associated with strong regional cleavage or schistosity and pre-date the Baltoro Granite and lamprophyre dykes (Desio and Zanettin 1970, Searle 1991). The metamorphic grade decreases northward. The shales, sandstones and conglomerates of Khalkhal Formation (? Cretaceous) form the highest stratigraphic level of the Gasherbrum sedimentary sequence. On Broad Peak, the rocks are highly deformed and metamorphosed to greenschist facies.

Karakoram Batholith

In the summary on the petrological aspects of the batholith in Chapter 6, it was mentioned that the Karakoram Batholith can be divided into two major groups of rocks. The first group consists of plutons which are metamorphosed and deformed, and the second group consists of unmetamorphosed, undeformed plutons. Mineral parageneses, together with radiometric dates, have been presented in Chapter 6 and are, therefore, not described any further. It appears that maximum metamorphic grade in these rocks reached upper amphibolite facies and was accompanied locally by partial melting. The division of the deformed and undeformed plutons coincides with the Late Cretaceous (about 85 Ma) collision of the Kohistan arc with the Karakoram plate.

Karakoram Metamorphic Complex

A narrow belt of metamorphic rocks lies to the south of the Karakoram Batholith and north of the Shyok Suture. These rocks, forming the Karakoram Metamorphic Complex of Searle (1991), were previously divided by Desio (1979) into the Dumordu, Ganchen and Askore Groups. The rocks have been studied in several places. Collectively referred to as the Darkot Group (Ivanac et al. 1954, Pudsey et al. 1985), they display only low grade metamorphism in the western Karakoram. In Hunza and further east, however, they show up to sillimanite grade metamorphism. Searle (1991) has provided a detailed integrated account of the rocks. The key points of the published data from Hunza and Braldu–Hushe areas are given in the following.

Hunza Valley

North of the Shyok Suture at Chalt, there is a progressive northward increase in metamorphism via a series of reaction isograds (Fig. 7.2) that separate pelitic zones in which stable mineral pairs are: garnet-chlorite, chloritoid-biotite, staurolite, kyanite-biotite, and sillimanite-biotite (Broughton et al. 1985). Metamorphic conditions reached up to 630–670° C at 5.5 kbar, with an XCO_2 of fluid = 0.66 in minor calc-silicate rocks. The assemblage at the ruby deposit in Hunza contains garnet-bearing gneisses and schists formed at 600-620° C and water vapour pressure of 6–7 kbar, with 20 mole % CO_2 for ruby-corundum marbles northeast of Aliabad (Okrusch et al. 1976). The isograds dip to the northeast underneath the Karakoram Batholith and are thus inverted (Windly 1983).

Further data on metamorphic mineral assemblages have been provided by Prior (1987) and Crawford (1988), and summarised by Searle (1991). The Dumordu Unit is dominantly marble with minor calc-schists, mica schists and plagioclase–biotite ± hornblende ± almandine gneisses. Impure marbles contain one or more of the phases phlogopite, fuchsite, pyrite, diopside, tremolite, forsterite, ruby and spinel. Leucocratic gneisses are frequently interbedded on a small scale with the marbles. The Ganchen pelites, metamorphosed up to sillimanite grade and containing rotated porphyroblasts (Powell and Vernon 1979, Prior 1987), have intercalations of amphibolites with garnet and biotite. The Askore unit amphibolites contain hornblende, plagioclase, quartz and garnet with minor zoisite, sphene, clinopyroxene and apatite. Deformed granodiorite sheets within the metamorphic complex look like augen gneisses, and east of Hunza there are some migmatites.

Metamorphism in the area may have been associated with the southward thrusting on the margin of the Karakoram plate after collision with the Kohistan arc during the Late Cretaceous. Searle (1991) thinks that it is post-collision but prior to the intrusion of both Sumayar and Baltoro plutons of 20–25 Ma. K-Ar and Ar-Ar ages (40–12 Ma) indicate a prolonged cooling history for the metamorphic complex. The inverted metamorphism, if true, can be explained in two ways: (1) thrust emplacement of hot granites over cold sediments, or (2) thrusting of deeper level, higher grade rocks over shallower colder rocks. Searle et al. (1986, 1989) and Searle (1991) favour the latter mechanism for both the Hunza and Braldu–Hushe areas.

Braldu–Hushe area

Rocks in this area show a complex and prolonged history of deformation and metamorphism. The earlier part of that history is overprinted by the strong tectono-metamorphic effects of continental collision. The area has been investigated by several workers summarised in Searle (1991). The description of the various units is taken from him, together with Figure 7.3 showing the metamorphic grade and Figure 7.4 showing P-T grid of the assemblages in the area.

Hushe Complex: This Complex is closely associated with the older components of the Kande plutonic unit. It consists of foliated hornblendite, diorite/amphibolite, granodiorite, monzonite and K-feldspar megacrystic granite. Thin bands of marble and psammitic metasediments are associated with the orthogneisses. Radiometric ages suggest that these subduction-related plutonic rocks were generated along the southern margin of the Karakoram in an Andean-type margin as far back as the Early Jurassic. The complex underwent a Jurassic low-P metamorphism, the common assemblage being andalusite-staurolite-garnet-two micas-quartz-plagioclase formed at 525° C, 3 kbar. In places, sillimanite replaces andalusite.

Dumordu Unit: East of Askole and south of the Braldu River, this unit consists of pure marble with minor amounts of graphitic schists, garnet-biotite amphibolite, and metapelites ranging from chlorite-chloritoid or biotite schists (Chingang Valley) to kyanite and sillimanite-bearing gneisses (upper Braldu River). High-grade marbles and calc-silicates around the snout of Biafo Glacier contain diopside, forsterite, phlogopite, hornblende, garnet, calcite, sphene, epidote and plagioclase in various proportions. There also are metamorphosed graywackes and orthoquartzites. The Dumordu Unit has been intruded by the Jurassic plutons.

Ganchen Unit: This consists of pelites with subordinate amphibolites and gneisses, and minor marble. The mineral assemblages in the pelites vary with metamorphic intensity from chlorite- and chloritoid-bearing to those containing kyanite and, possibly, sillimanite. Kyanite-bearing rocks are common and the assemblage kyanite-staurolite-garnet-two micas-plagioclase-quartz at Braldu–Panmah junction indicates > 5.5 kbar pressure. The unit has been intruded by the undeformed Mango Gusar Granite, which has a U-Pb zircon age of 37 ± 0.8 Ma.

Askole Amphibolite Unit: This unit comprises amphibolites interbedded within marbles, pelites, and gneisses. The amphibolites may contain almandine-rich garnet, diopside, and epidote, and are interpreted to be meta-igneous. The marbles and pelitic gneisses are high-grade, the latter containing kyanite or sillimanite, garnet and biotite.

Panmah Ultramafic Unit: A prominent belt of tectonic melange immediately south of the Karakoram Batholith consists of blocks of gabbro, basalt, chert and ultramafic rocks in a shaly matrix. The rocks are altered/metamorphosed, with talc, antigorite, magnesite and chlorite commonly grown in the ultramafic lithologies. The origin of these rocks is not known.

Dassu Gneiss: There is a domal structure containing orthogneisses (granodiorite, quartz

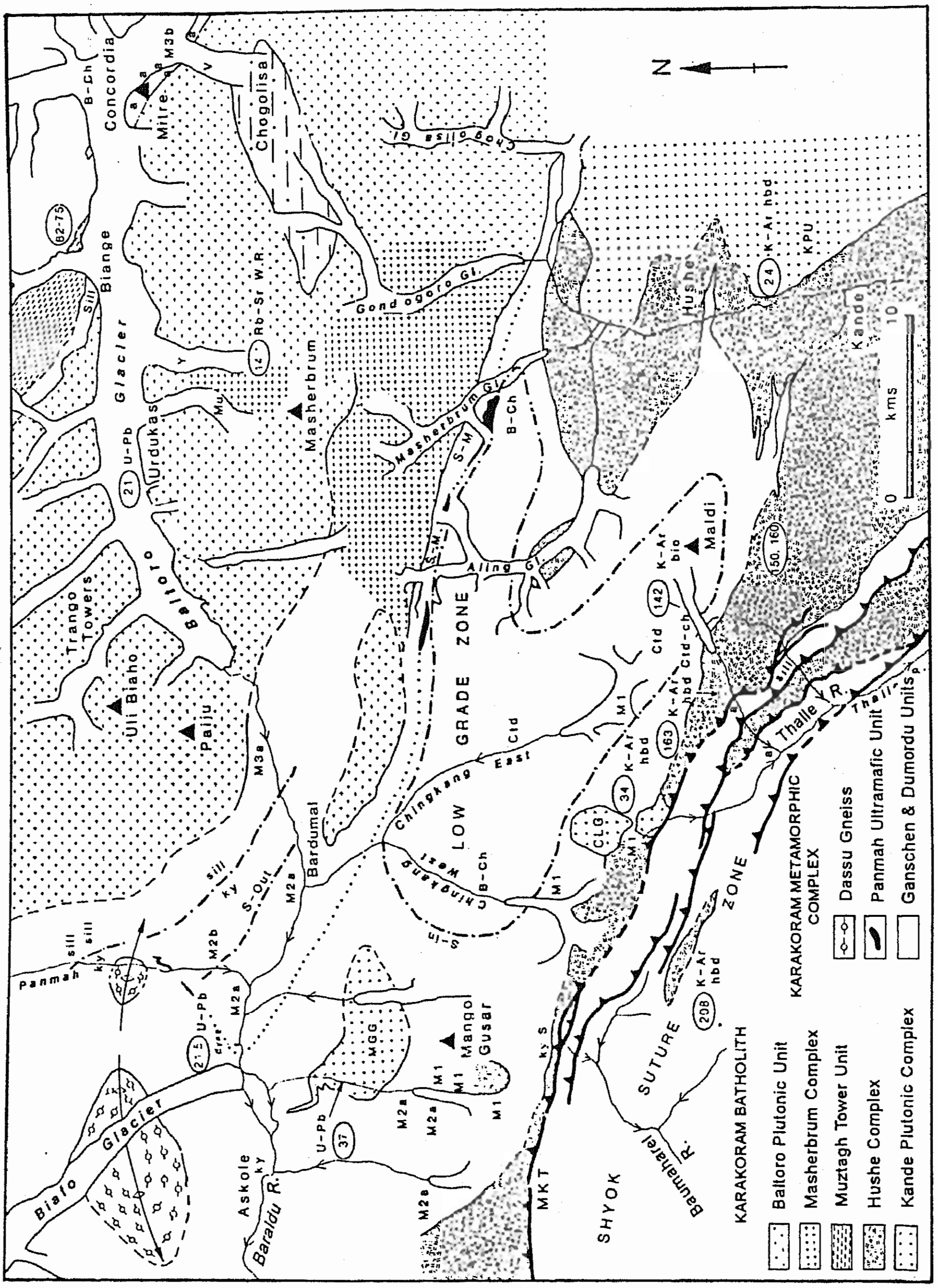


Fig. 7.3. Map showing lithologic units and metamorphic grade in the Karakoram Metamorphic Complex south and west of Baltoro Glacier. CLG, Chingkang-1a granite; MGG, Mango Gusar granite; KPU, Kande Plutonic unit. M_1, M_2, etc. refer to metamorphic episodes; W.R., whole rock. Encircled figures indicate ages.(From Searle 1991).

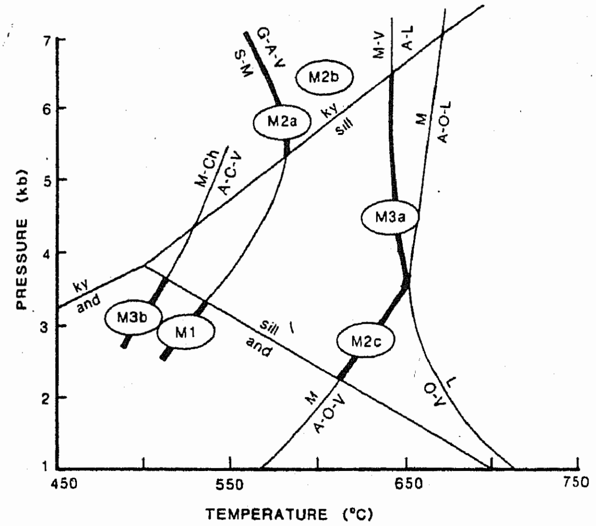


Fig. 7.4. Pressure-temperature grid showing key metamorphic assemblages and reactions in the region shown in Fig. 7.3. All assemblages contain biotite, plagioclase and quartz. A, andalusite, kyanite or sillimanite; C, cordierite; Ch, chlorite; G, garnet; L, granitic liquid; M, muscovite; O, orthoclase; S, staurolite; V, vapour. M1, M2, M3 refer to metamorphic phases and locations of assemblages are shown in Fig. 7.3. Granite wet melting curve is after Fyfe (1970). (From Searle 1991).

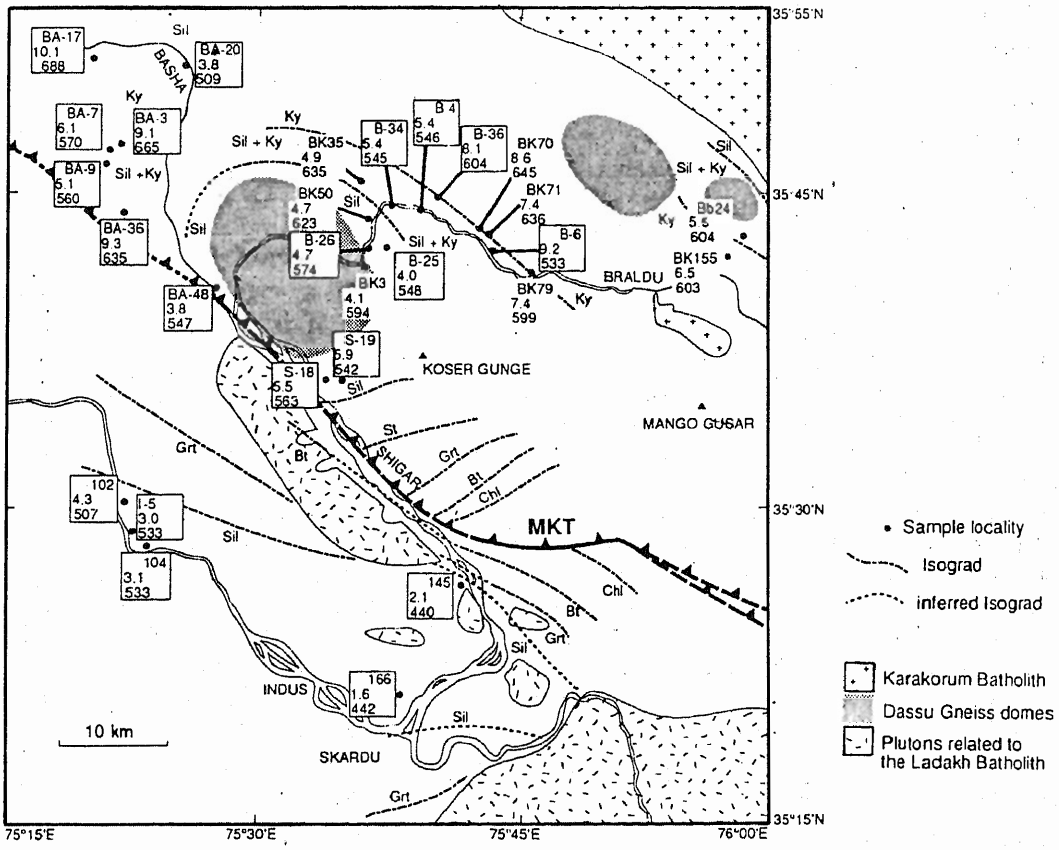


Fig. 7.5. Metamorphic map of Baltistan region, northern Pakistan, showing isograds, sample localities, and pressure (kbar) and temperature (° C) determinations. (From Allan and Chamberlain 1991).

diorite, granite) and migmatite at Dassu. Such gneiss domes also occur in several other places (Allen and Chamberlain 1991, Searle 1991, Le Fort et al. 1995). The Dassu felsic gneiss is composed of biotite, two-feldspar, quartz, sillimanite ± garnet ± muscovite. There are pods, veins and dykes of leucogranite containing garnet, two micas ± tourmaline ± beryl. Foliation in the surrounding high-grade pelitic and calc-silicate gneisses wraps around the domes. Metamorphic isograds are also folded around the Dassu dome, so these structures may be post-metamorphic (Searle 1991).

In Mangol Bluk dome, isoclinally folded bands of ortho- and paragneisses are refolded in a great antiformal structure of N 140 direction (Le Fort et al. 1995). These authors, Allen and Chamberlain (1991) and Searle (1991) have discussed briefly the thermal and structural relations of the rocks. Of the four mechanisms, i.e., extension, diapirism, compression or wrenching, at least extension does not seem to be applicable to the formation of the domes in the area.

Tectono-metamorphic evolution of the Karakoram Metamorphic Complex

Data on pressure and temperature condition in the complex, together with metamorphic isograds, are shown in Figures 7.4–7.6. The Cretaceous-Tertiary magmatic and tectono-metamorphic history of the area is shown in Figure 6.31. Searle et al. (1989) and Searle (1991) suggest that the complex passed through four temporally distinct phases of metamorphism: (M_1) phase of low-P regional metamorphism synchronous with pre-collision (Jurassic) plutonism; (M_2) medium-high-P, high-T Barrovian metamorphism predating the 37 Ma undeformed, unmetamorphosed, Mango Gusar pluton. The upper age estimate (50 Ma) is uncertain, but the metamorphism and three phases of deformation (D_1–D_3) have been related to collision tectonics. This event includes a continuous evolution from kyanite to sillimanite; (M_3) phase refers to high-T, low-P contact metamorphism around the large Baltoro plutonic unit (25–21 Ma), and (M_4) to retrograde metamorphism related to post-Miocene thrusting along the Shyok Suture.

Bertrand et al. (1988) contemplate two phases of regional metamorphism for the rocks to the north of the Shyok Suture. An early (pre-Mango Gusar pluton) medium-P metamorphism with a garnet-kyanite-sillimanite paragenesis was associated with D_1 or D_2 isoclinal folding, and a younger low-P thermal event (D_2–D_3, producing sillimanite) corresponded to the doming phase (open folds) between 20 and 8 Ma (LeFort et al. 1995).

Allen and Chamberlain (1991) provided further data on the metamorphic rocks to the north and south of the MKT in Baltistan. According to them the Karakoram Metamorphic Complex suffered a medium- to high-P metamorphism syntectonic with D_1 deformation. The prograde isograds were subsequently deformed by south-verging folding and emplacement of gneiss domes (D_2–D_3). Decompression metamorphic reactions occurred during nappe formation (D_2) so that high-P rocks are associated with the higher level nappes, creating an inverted pressure metamorphic sequence (8–9 kbar rocks over 5–6 kbar rocks) with little variation in temperature with the structural level (550–620° C) (Fig. 7.6).

The exact timing of metamorphism is poorly constrained, except that it is older than 37 Ma. Like Hunza, it is also considered to be related to the collisional tectonics. Petterson and Windley (1985) reported that collision between the Karakoram plate and Kohistan arc took place some 85 Ma ago. Recently, it has been suggested that Kohistan-India collision may have occurred about the Cretaceous-Paleocene boundary (Klootwijk et al. 1992, Beck et al. 1995). Apparently the various phases of deformation and metamorphism are related to these two collisional events.

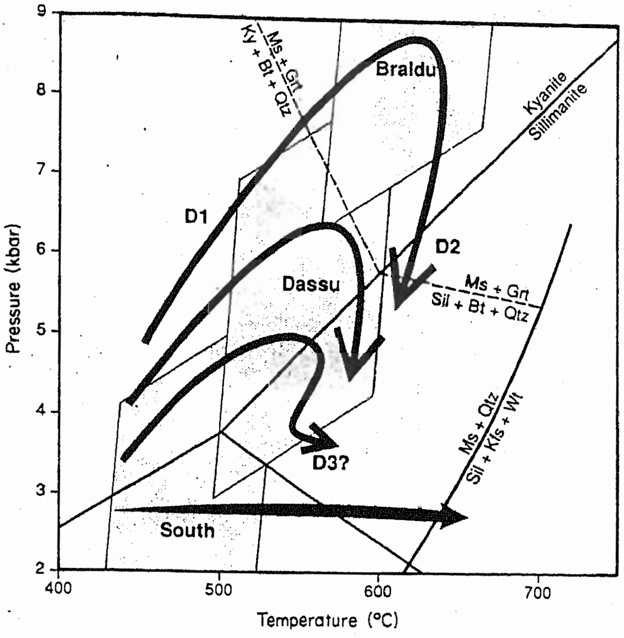


Fig. 7.6. Pressure-temperature plot with reactions. Aluminosilicate phase diagram from Holdaway(1971). Muscovite breakdown curve from Chatterjee and Johannes (1974). Dashed line gives slope calculated from entropy and volume data for the reactions muscovite + garnet = 2 sillimanite/kyanite + quartz + biotite. Arrows indicate the P–T history inferred for rocks from south and north of the MKT. Paths from north of the MKT are also labelled D_1 , D_2 (nappe stage), and D_3 (dome stage). Shaded regions represent error boxes of ± 2 kbar and ± 50° C for averages of the P–T determinations for rocks from south of the MKT, the Dassu Nappe and the Braldu Nappe. (From Allan and Chamberlain 1991).

Main Karakoram Thrust

Petrographic details of the Shyok Suture melange are given in Chapter 6. The Main Karakoram Thrust (MKT) is commonly considered synonymous with the Shyok Suture. The MKT has been interpreted as a late (<5 Ma ?) south-directed break-back thrust with several hanging-wall splays that offset the regional pre-37 Ma isograds (Rex et al. 1988, Searle et al. 1989, Coward et al. 1987). It may also have been an important structure earlier in the evolution of the Karakoram (Searle 1991). The uplift and exposure of the igneous and metamorphic rocks of the southern Karakoram are connected to southward thrusting on the MKT. In Baltistan, the MKT juxtaposes the medium- to high-P rocks of the Karakoram against the low-P rocks of the Kohistan (Allen and Chamberlain 1991). This has, therefore, obscured the position of the Shyok Suture. Bertrand et al. (1988) contend that the actual Shyok Suture may be represented by the bands of ultramafic rocks within the Karakoram Metamorphic Complex. If so, the suture has been complexly folded and metamorphosed by the D_2 nappe and D_3 dome stages, leading to the speculation that the D_2–D_3 tectono-metamorphic phases may be related to the closure along the Indus Suture (Allen and Chamberlain 1991).

KOHISTAN ARC

In Chapter 6 we gave a summary of the petrography and geochemistry of the magmatic rocks which make up the so-called Kohistan island arc. Here we present a brief account of the metamorphic petrology of Kohistan, roughly along a north to south traverse between the Shyok and Indus Sutures. A fair amount of data have been published on the mineral chemistry and thermobarometry of Kohistan. But much of these concern selected parts of the southern half of the terrane. Integrated accounts of metamorphism are lacking except those of Bard (1982a,b) and, for the amphibolite belt, Treloar et al. (1990).

General petrological aspects

Northern volcanics and sediments: The Cretaceous-Tertiary rocks of the Yasin and Dir Groups and Shamran and Utror Volcanics have suffered a low-grade metamorphism. Chlorite (± epidote) is common in appropriate lithologies. Metamorphism seems selective, possibly aided by fluid migration along pore spaces and shear zones. In Utror area, for example, volcanic rocks containing abundant chlorite and epidote may occur adjacent to those containing abundant glass. The Chalt Volcanics are strongly deformed and show tight isoclinal folding. These have also undergone a low-temperature (greenschist facies) pervasive metamorphism and contain chlorite, epidote, sodic plagioclase and, locally, hornblende. Higher grade (amphibolite facies) conditions prevailed locally. On the northern side of the Haramosh massif and to the north of Shandur, in the vicinity of the Shyok Suture, the volcanics are converted to amphibolites and amphibole gneisses, with local biotite schists. The pressure of metamorphism is not known but the Karakoram metasediments in the immediate hanging wall of the suture in Hunza Valley show Barrovian-type facies series. In Haramosh area, rhombohedral calcite and Fe-poor epidote have been suspected of pseudomorphing lawsonite (Le Fort et al. 1995), raising the possibility of high-P metamorphism related to suturing.

Kohistan Batholith: The batholith comprises deformed and undeformed plutons. The former, consisting of a bimodal trondhjemite and gabbro-diorite suite, are characterised by E-W folds and strong foliation. Detailed studies by Petterson and co-workers (1985, 1990, 1993) have shown that the early stage plutons owe their deformation and metamorphism to collision along the Shyok suture. The unaltered deformed rocks contain stable biotite and amphibole, suggesting amphibolite facies metamorphism. Many rocks also display chlorite and epidote assemblages, but these might have grown subsequently. Because the batholith has been the site of magmatism for some 70 million years, regional metamorphic processes have probably combined with magmatic process. It would not be surprising at all if some of the earliest plutons have undergone partial melting.

Jaglot Schist Group: The sedimentary and volcanic rocks of this group show strong deformation, isoclinal folding, and medium- to high-grade metamorphism. The pelitic rocks range from biotite schists, through garnet and staurolite schists, to sillimanite ± K-feldspar schists and gneisses. Calc-silicates contain calcite, diopside, tremolite ± garnet, and volcanics are amphibolitised. There is evidence for partial melting and it has been suggested that some of the plutonic rocks in the Kohistan Batholith may be related to the Jaglot Group. Chamberlain et al. (1989a,b) studied some samples of the schists and noted that garnet cores formed at higher pressure (9–9.5 kbar) than rims (7–7.5 kbar, 580° C). They thought that the rims of the garnet grew when the Kohistan terrane was thrust onto India along the Indus Suture.

The metamorphosed volcanic and sedimentary rocks of the Ladakh arc in Deosai–Skardu area were briefly described in Chapter 6. South of Skardu the rocks display a low-grade (greenschist facies) metamorphism. On the eastern flank of the Nanga Parbat–Haramosh massif, however, the volcanic rocks have been converted to garnet amphibolites (Askore amphibolite). In Skardu area, the Katzara Formation shows complex deformation (two phases) and has been metamorphosed from biotite to sillimanite-K-feldspar zone. The metamorphism here has been differently interpreted. Hanson (1989) showed regional metamorphic isograds cut by later plutons. Allen and Chamberlain (1991), on the other hand, report that the rocks underwent a static low-P (2–4 kbar, 450–650° C) thermal metamorphism related to intrusion of igneous plutons.

Chilas Complex: This huge complex consists principally of mafic to intermediate "gabbronorites" containing plagioclase, two pyroxenes, ilmenite, magnetite, apatite, ± quartz ± hornblende ± biotite ± K-feldspar. The plagioclase is labradorite to andesine (An_{64} to An_{40}) and antiperthitic in many rocks, whereas the K-feldspar is invariably perthitic. Orthopyroxene ranges from En_{76} to En_{48}, clinopyroxene is diopsidic (Mg_{75-55}) and hornblende is pargasitic. Apparently intrusive into these rocks are small bodies of ultramafic-mafic-anorthosite (UMA) association containing very calcic plagioclase (An >84) and more magnesian pyroxenes than the principal norites. Both these groups have been intruded by dykes of hornblende-plagioclase pegmatites and amphibolites, which have caused local hydration of the host rocks to amphibolites. The field, mineralogical and petrological aspects of the complex have been described in many papers referred to in Khan and Jan (1992) and Khan et al. (1993).

After emplacement, the complex underwent an essentially static, low-P granulite facies re-equilibration at about 800° C and 5–6.5 kbar (Misch 1949, Jan and Howie 1980, 1982, Bard 1982b), During this "episode", the primary textures in at least the principal gabbronorite were recrystallised to a polygonal granoblastic metamorphic fabric. Although the UMA association still retains several igneous textures and structures, including excellent layering, the calcic plagioclase and olivine have invariably reacted to produce orthopyroxene, clinopyroxene and/or hornblende-spinel coronas. The recrystallisation was at least locally associated with, or shortly followed by, high-T deformation: some rocks are gneissose, or granulated with porphyroclastic feldspar and pyroxene showing straining (Jan 1979d). This, apparently, was followed by a phase of static hydration when pyroxenes were partially replaced by pargasitic amphibole under high-T amphibolite facies conditions (Treloar et al. 1990).

The principal deformation within the complex is confined to a series of folds with northward vergence (e.g., the asymmetric hanging wall antiform: Coward et al. 1987), and narrow shear zones of limited displacement. Some of these shears are ductile and hydration along these resulted in the formation of amphibolites (± garnet ± epidote ± scapolite) and hornblendites. Others are low-T shear zones with greenschist facies assemblages (chlorite, actinolite, epidote) or cataclastic gouge infills. These have grown during and after uplift. The amphibolite facies deformation and metamorphism are synchronous with the main deformation in the southern amphibolite belt described in the following. The extensive zone of amphibolites with relics of gabbronorites forming the southern part of the complex is a product of the Kamila shear. Such shears are not common in the northern part of the complex.

There is some disagreement over the emplacement age of the complex despite a concordia U-Pb zircon age of 84 Ma from a gabbronorite in Upper Swat (Zeitler et al. 1980). Ar-Ar and K-Ar ages from the Chilas Complex indicate that cooling through 500° C (hornblende blocking temperature) occurred around 80 Ma (Treloar et al. 1989), similar to the amphibolites in the Kamila shear. These ages have important implications: 1) the Chilas Complex was emplaced soon after collision between the Karakoram plate and Kohistan arc, 2) the granulite facies re-equilibration occurred quickly after emplacement, 3) the amphibolite facies shears, which have been related to imbrication and southward thrusting of the complex, followed this re-equilibration quickly, and 4) the Chilas Complex may have provided thermal energy for the Late Cretaceous regional metamorphism in the Kohistan arc.

Southern amphibolite belt and the Kamila shear zone: In Chapter 6 we noted that the rocks of the southern amphibolite belt may have formed from tholeiitic and calc-alkaline magmas generated in island arc and oceanic settings. This complex magmatic scenario is well-matched by the complicated tectono-metamorphic history of the belt. It is principally occupied by amphibolites, but there is a wide range of plutonic rocks and minor metasediments. The amphibolites are fine- to coarse-grained and homogeneous to banded; much of the banding is the result of shearing. Relics of metamorphosed pillow lavas and gabbros suggest that the amphibolites are derived from both volcanic and plutonic precursors. Apparently, the amphibolite belt may initially have been dominated by gabbroic plutons. There are abundant relics of plutonic rocks in Upper Swat, Khwaza Khela, Kayal-Pattan and Sapat areas, thus raising the possibility that such rocks might have constituted the entire southern edge of the belt. Our geochemical studies in the Kayal–Pattan area show that the coarse-grained amphibolites are identical to, and hence derived from, the gabbronorite relics.

The amphibolite belt is structurally complex with at least two phases of deformation (isoclinal folding, shearing) and metamorphism (Windley 1983). It has been noted that in the Indus and Swat Valleys the amphibolite belt and the southern part of the Chilas Complex show a deep to mid-crustal structure predating the Himalayan collision between the Indian plate and Kohistan. This structure was named the Kamila shear zone by Treloar et al. (1990). Extending along the length of Kohistan, it narrows from 35 km in the Indus Valley to only 3.5 km to the south of Jal, Thak Valley. The metagabbros of the Chilas Complex were transported southwards across the shear zone onto a stack of high-P rocks that have been assembled in the hanging wall of the Tethyan subduction zone. The shear zone is constituted by an array of mylonitic zones anastomosing around areas that show little evidence of strain. These unstrained areas contain relics of the precursors of the amphibolites, which in the Indus Valley are metagabbroic. The intensity of the shear zones decreases rapidly some 12 km north of Dassu. The shear fabric is ductile, with the development of a compositional segregation of leucocratic material, and the alignment of minerals and elongated lenses of deformed mineral aggregates. The majority of shear criteria has a south to southwest verging thrust sense (Khan 1988, Treloar et al. 1990). Hydration along the shears has produced amphibolite facies assemblages. Post-dating these ductile shears is a series of low-temperature shear zones developed during the exhumation of the rocks. Some of these contain greenschist facies assemblages (chlorite, actinolite, epidote), and others are represented by cataclasites and gouge fillings. Although a few of these show south-verging thrust sense of movement, most are extensional north-side down faults. Some of these low-T shear zones developed prior to collision between Kohistan and India; others may be break-back structures developed during the collision which breached the MMT (Indus Suture), locally thickening the Kohistan (Searle et al. 1987).

Variation in the bulk composition and degree of metamorphism has resulted in a range of parageneses in the 300 km long amphibolite belt. Pargasitic to tschermakitic hornblende + plagioclase and/or epidote make up most of the amphibolites, but a wide range of minerals has been reported in different areas: quartz, garnet, clinopyroxene, cummingtonite, actinolite, biotite, muscovite, fuchsite, paragonite, margarite, chlorite, green spinel, magnetite, corundum, and K-feldspar. Corundum ± staurolite amphibolite containing Al-tschermakite and secondary margarite occurs sparingly in Dir and Swat (Jan et al. 1971). Metagabbros in Upper Swat locally contain veins bearing high-pressure assemblages: garnet-clinopyroxene-hornblende, and zoisite-kyanite.

Metamorphism and deformation within the amphibolite belt seem to have occurred under amphibolite facies conditions. Jan (1988) proposed conditions of 550 to 680° C, 4.5 to 6.5 kbar, with metamorphic grade showing an increase from the centre towards the Chilas Complex to the north and the Indus Suture to the south. Treloar et al. (1990) suggested that maximum operating pressure may have been 9–10 kbar. Whether the entire belt has undergone a high-P metamorphism is not clear, but there are suggestions that the southern part of the belt may have been metamorphosed at considerable depth. Jan (1988) reported relics of the high-pressure assemblage garnet-clinopyroxene ± plagioclase ± hornblende in amphibolites north of Alpurai. Bard (1982b) estimated 10 ± 1 kbar pressure for kyanite-zoisite-staurolite assemblages in veins within amphibolites just north of Pattan. Chamberlain et al. (1991) noted that the volcanic components of the southern amphibolite belt caught up in the Indus Suture melange at Babusar were metamorphosed at 585° C, 9 kbar. There are several occurrences of high-P veins; those of the Khwaza Khela area contain kyanite-zoisite-garnet-diopside-corundum-quartz-paragonite-calcite-scapolite formed at 500–600° C, 10–12 kbar (Jan and Karim 1995). These and the nearby occurring melange assemblage, probably, post-date the amphibolite facies metamorphism.

Metamorphic conditions in the amphibolite were appropriate for partial melting and there are occurrences of migmatites. It has been suggested that some of the granitic rocks in the amphibolite belt and the Kohistan Batholith may be derived from partial melts of the amphibolites. The granitic, trondhjemitic and pegmatitic sheets in the amphibolites range from deformed garnet-epidote-bearing varieties to those cutting the fabric within the host rocks. Treloar et al. (1990) think of them as largely syntectonic but they may span a longer period. Between the Jal shear zone and Indus Suture, there is a broad zone of relatively lower grade basaltic and andesitic volcanic, and sedimentary rocks intruded by diorite (Khan 1988). These may be the unsheared equivalent of the Kamila zone.

^{40}Ar-^{39}Ar cooling ages derived from hornblende within the shear zone suggest that by 83 Ma the amphibolites had cooled through 500° C. This also gives minimum age for the shearing and implies that deformation occurred long before the India-Kohistan collision. K-Ar muscovite age of 66 ± 2 Ma from a pegmatite cross-cutting the main shear fabric indicates that substantial post-shearing uplift and cooling had occurred within Kohistan by 60 Ma (Treloar et al. 1989).

Jijal Complex: This 150 km^2 tectonic wedge occurs in the hanging wall of the Indus Suture to the south of the amphibolites. The northern half of this Complex consists of garnet granulites and the southern half of ultramafic rocks, both of which display good layering in some parts. The ultramafic rocks are devoid of garnet and plagioclase, and consist of diopsidites (± olivine ± orthopyroxene), websterites, peridotites and dunite containing accessory chromite and segregated chromite deposits. The mineral chemistry of these has been presented in detail by Jan and Windley (1990). Olivine ranges from Fo_{94} to Fo_{84}, orthopyroxene from En_{91} to En_{82} and Ca-rich diopside from Mg_{44} to Mg_{50}; chromite has characteristically high Cr/(Cr + Al) ratio. These rocks apparently equilibrated at about 800° C and 8–12 kbar.

The garnet granulites range from intermediate to ultramafic in composition and are represented by a variety of parageneses. Basic rocks are the most abundant and, together with intermediate members, consist typically of plagioclase-garnet-clinopyroxene-quartz-rutile ± ilmenite ± magnetite ± hornblende ± epidote ± scapolite. The retrograde versus prograde status of the last three minerals is not clear. Near Pattan these rocks contain enclaves of two-pyroxenes-plagioclase-pargasite-ilmenite granulites considered to be the precursors of the garnet granulites. There are layers and several metres large masses of "garnetites" containing clinopyroxene and hornblende. In the southern part of the granulites are small bodies of garnet websterite and garnet diopsidite ± hornblende ± green spinel. The petrography and mineral chemistry of the complex has been investigated in detail. Jan and Howie (1981) and Bard (1982b) suggested that the high-P metamorphism occurred at 700–800° C and 11–15 kbar. Miller et al. (1992) thought that the mineral assemblages are a direct product of magmatic crystallisation at high pressure. Yamamoto (1993) reported an anti-clockwise P-T-time path for the granulites, with the highest point at 950° C and 17 kbar.

Retrograde assemblages formed during uplift-cooling. Plagioclase-hornblende-epidote-paragonite-quartz paragenesis records conditions of 500–550° C, 8–9 kbar (Jan et al. 1982). Anorthositic lithologies consist of Fe-poor epidote-kyanite-paragonite-quartz ± hornblende. Ductile shears and thin veins contain garnet, hornblende and epidote. Lower grade (greenschist facies) mineral assemblages (albite, epidote, actinolite, chlorite) occur as alteration products in microfractures and veins, and some granulites along the southern contact near Jijal are mylonised with abundant growth of chlorite and green-blue amphibole.

Hornblendites make an important component of the Jijal Complex. Occurring in bodies up to several tens of metres across, these commonly contain some garnet, rutile and, in rare cases, clinopyroxene. Locally, however, the rocks consist entirely of hornblende with secondary chlorite and epidote. Feldspathic rocks are sparingly associated with these rocks and it is tempting to think that the hornblendites are a product of hydration of the granulites during uplift. But these are traversed by veins of garnet-clinopyroxene assemblage, casting doubt on such an interpretation.

Little is known about the age of metamorphism and deformation of the Jijal Complex. Sm-Nd ages range from 91 ± 6 to 114 ± 39 Ma (Coward et al. 1986, Sano et al. 1996, Yamamoto and Nakamura 1996), but it is not clear whether these record magmatism or metamorphism. For a long time the Jijal rocks looked out of context in their surroundings because of their high pressure of recrystallisation. But the discovery of high-P rocks in the neighbouring amphibolite belt lends support to the idea that these rocks may represent the metamorphic base of the thickened Kohistan arc.

Tectono–metamorphic evolution of Kohistan

Bard (1982a,b) presented a comprehensive model for the thermo-tectonic evolution of the Kohistan arc and showed several broad zones of metamorphism from greenschist to amphibolite facies. He contended that Kohistan underwent two major phases of tectono-metamorphic events, D_1 and D_2. The D_1 event occurred during the Late Cretaceous (85 Ma), synchronous with blueschist metamorphism in the Indus Suture. It was related to the obduction of the arc over India. The D_2 event was presumed to be related to the Upper

Eocene-Lower Oligocene collision of India and Asia and synchronous with the first Alpine phase. According to him (Fig. 7.7), D_1 metamorphism increased in grade, downwards in

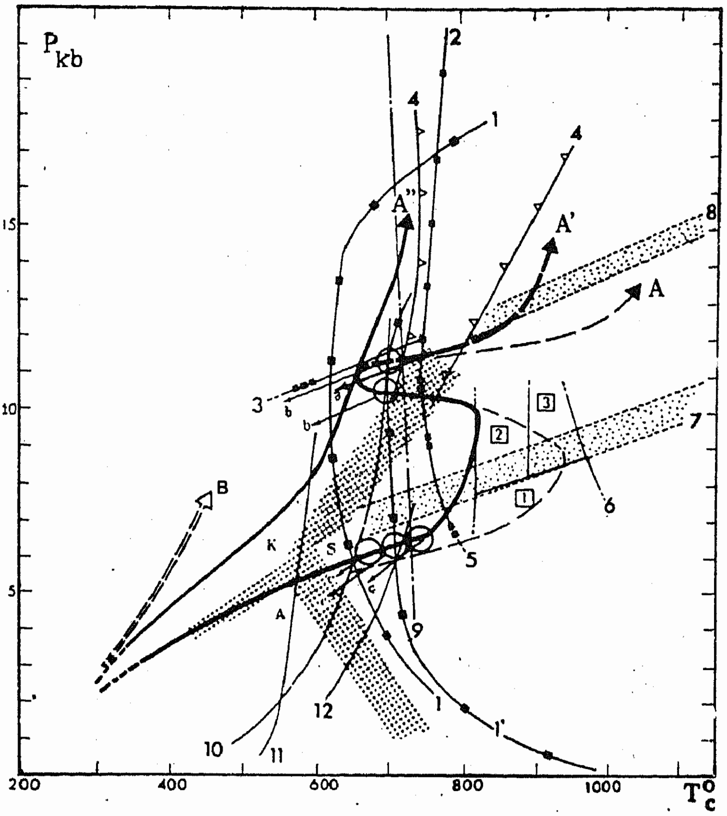


Fig. 7.7. Tentative $P(P_L)$, T paths during the metamorphic evolution of the Kohistan island arc; geotherms A, A' respectively for earlier and late-D_1 event crosscutting the granulitic belt (metanoritic layered intrusive) and the Jijal Complex; geotherm A" for syn- to late-D_1 event for the latter complexes; geotherm B for syn- to late-D_2 event (K-S-A: average stability fields of the Al_2SiO_5 polymorphs); 1 : beginning of partial melting of water-saturated gabbros (Wyllie 1977); 1' : approximate melting curve of amphibolitic rocks (Binns 1969); 2: zo + ky + quartz → H_2O + L (Boettcher 1970); 3 : an + H_2O (low P) → zo + ky + quartz (high P) (Boettcher 1970); 4 and 4' : pure Mg-staurolite high-P stability field (Schreyer and Seifert 1969); 5: an + zo + quartz + gt + H_2O → L (Boettcher 1970); 6 : various stability fields in the system CaO-MgO-Al_2O_3- SiO_2-H_2O : [1] an + forst + H_2O, [2] : an + hb + cpx + sp + H_2O, [3] : an + cpx + opx + H_2O; 7 : approximate reaction field of ol + an → opx + cpx + sp in tholeiitic basalts (see literature in Ringwood 1975; and Griffin 1971); 8 : approximate reaction field of opx + an → cpx + gt in tholeiitic basalts (Ringwood 1975; Griffin 1971); 9 : Al-ep → zoisite (Holdaway 1972); 10 : epidote + quartz → gt + an^{*}+ mgt + H_2O (Liou 1973; NNO buffer); 11 : chlorite "out" in metabasites (Liou et al. 1974) as isograd separating the greenschist facies from the so-called epidote-amphibolite facies; 12 : epidote "out" from reaction ep + An_{0-20} → An_{30-40} + quartz + H_2O under HM buffer conditions (high fO_2 conditions) (Liou et al. 1974). Circles with arrows : possible sub-in situ melting of amphibolitic rocks giving late D_1 uprising ky + zo + st (a) and/or gt + zo trondhjemitic melts (b) and dioritic sub-autochthonous intrusives (c). (From Bard 1983b).

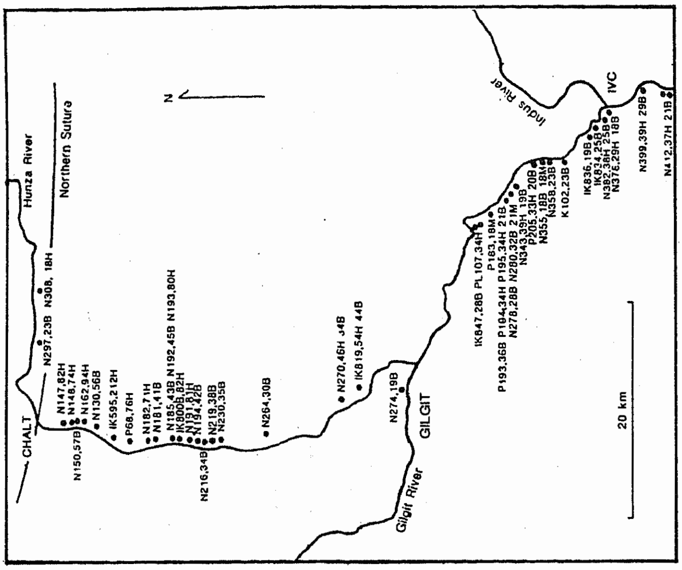


Fig. 7.9. K-Ar and Ar-Ar geochronological data along the Hunza and Gilgit Rivers. (From Treloar et al. 1989).

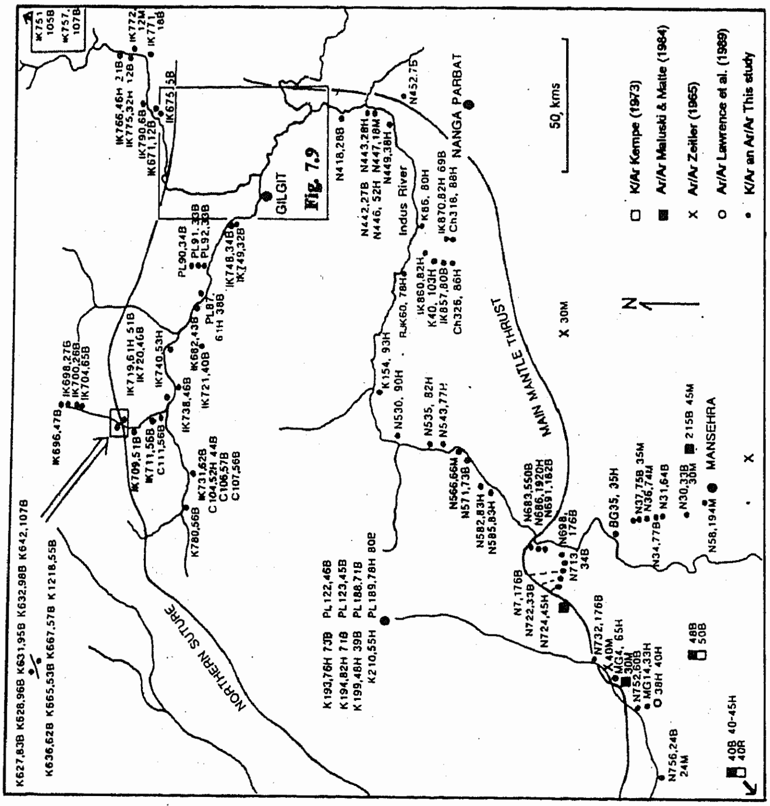


Fig. 7.8. K-Ar and Ar-Ar geochronological data from the Kohistan island arc and adjoining parts of the Indian and Asian plates. B, biotite; H, hornblende; M, muscovite; R, riebeckite. N147, 88H represents sample N147, with an 88 Ma hornblende age, and so on (From Treloar et al. 1989).

the Kohistan pile (i.e., southward in the present geographic context) but, more importantly, towards the "ensandwiched" Chilas Complex emplaced some 15–20 Ma before the obduction of the arc over the Indian plate. This Late Cretaceous thermal phase was, thus, strongly controlled by remnant magmatic heat in the Chilas Complex. The "distorted" or "sinuous" aspect of the PT trajectory in crustal segments has been argued to persist some 20 Ma after the emplacement of large plutons (Wells 1980). The distribution of the isotherms during the D_2 event, on the other hand, was regarded to be simple (Bard 1982a).

Over the last ten years, considerable data have been published on the radiometric ages (Figs. 7.8, 7.9), structure and petrology of the Kohistan arc, necessitating a modification of this model. It is now thought that Kohistan collided with the Karakoram plate some 85 Ma ago (Treloar et al. 1989) and with the northwestern margin of India during Early Paleocene (Klootwijk et al. 1992). Structural (Coward et al. 1987) and geochronological (Treloar et al. 1989) studies show that much of the deformation in Kohistan predated collision of Kohistan with India and is a combination of shearing and folding probably related to both suturing and subduction processes. Crustal shortening accompanied closure along the Shyok suture. In the northern part of the arc, dips of bedding and cleavage are clearly related to a number of large scale folds, such as the Jaglot syncline and Gilgit anticline (Coward et al. 1982, T. Khan et al. 1993) which predate the second stage Eocene plutons. Deformation in southern Kohistan is expressed in the 35 km wide Kamila shear zone involving the southern amphibolites and the southern part of the Chilas Complex.

The Kamila shear zone shows an earlier ductile (high temperature), and a later brittle (low temperature) deformation. Hornblende cooling-uplift ages suggest that shearing occurred before 80 Ma, well before Kohistan-India collision. Essentially, the shear zone reflects the southward thrusting of the Chilas Complex (Fig. 6.24). Many of the brittle shears, however, may have formed during the thrusting of the arc over the Indian plate.

To sum it up, there is a strong likelihood that the pre-80 Ma deformation and metamorphism in the Kohistan is related to overthrusting of the Karakoram plate onto Kohistan arc. It resulted in thickening of the crust, leading to high-P assemblages in the amphibolite belt, formed at 550 to 650° C and 9–10 kbar. The emplacement age of the Chilas Complex is disputed. It probably passed through a phase of granulite facies recrystallisation/re-equilibration and weak deformation soon after emplacement but before its involvement in the Kamila shear. This event may be related to the closure along the Shyok Suture, to subduction processes and/or thickening of the arc crust due to magmatism, with energy having been provided by the remnant heat in the complex or by younger plutons. But the Chilas Complex does not show much of the deformation that is assumed to be associated with the Shyok Suture and which has affected the stage I plutons in the Kohistan Batholith as well as the older volcanic and sedimentary rocks. Thus, there is a possibility that the complex post-dates the suture. In any case, there would be sufficient heat associated with this huge body to drive widespread regional metamorphism and partial melting of the Kamila amphibolites. There is an interesting concordance in the U-Pb zircon age of the complex (84 Ma), initiation of the stage II plutonism in the Kohistan Batholith (85 Ma), and the regional cooling through 500° C (about 83 Ma) (Zeitler et al. 1980, Petterson and Windley 1985, Treloar et al. 1989). It is likely that there would be insufficient heat production in a basic arc terrane to generate melting and high-T metamorphism solely through post-thickening thermal conduction, and hence direct heat advection from the mantle would have been a prime requirement (Treloar et al. 1996).

The Jijal stack may have assembled synchronously with shearing in the Kamila zone, essentially below the shear zone and above the subduction zone. These lower crustal rocks would have recrystallised at higher than 10 kbar pressures realised in the lower part of the Kamila zone, as previously discussed. Alternatively, the Jijal rocks represent a discrete, already assembled block metamorphosed 104 Ma ago.

The temperature-time plot showing the cooling history of the Kohistan rocks using geochronological data from various sources is shown in Figure 7.10 (Treloar et al. 1989). The cooling history suggests a decrease in cooling ages toward east (Nanga Parbat) and, therefore, an increase in cooling rates and exhumation over the last 20–30 Ma. The plots of ages against the distance from the western margin (Raikot–Sassi) of the Nanga Parbat massif are shown in Figure 7.11. There is a rapid decrease in the ages of hornblende, biotite, zircon and apatite towards the Raikot Fault. The uplift and erosion rates, which cause exhumation, are greater near the Nanga Parbat Syntaxis. The exhumation rate in Nanga Parbat has increased over the last 20 Ma to a present figure of 5.5 mm/year (Searle 1991, Zeitler et al. 1993, Winslow et al. 1994) in response to a very rapid uplift, estimated at 7 mm per year during the Recent (Butler and Prior 1988).

Uplift and cooling started soon after amphibolite facies metamorphism and much of southern Kohistan had cooled to <500° C by 80 Ma and <300°C before it started colliding with India (Treloar et al. 1989). This uplift/cooling history is documented in the microstructures and greenschist facies assemblages in superimposed shear zones. Subduction of a young buoyant feature, such as a ridge or young oceanic crust, beneath the Kohistan margin may have been responsible for this unroofing (Sullivan et al. 1993). But the principal uplift was thrust-related, and probably started when Kohistan collided with the Indian plate. Fission-track and Ar-Ar cooling ages show that uplift rates in southern Kohistan increased from 0.07 to 0.2 mm/year during the period 55 to 12 Ma, and to more than 0.3 mm/year since then (Zeitler 1985, Zeitler et al. 1982). The uplift, thus, has a substantial tectonic component and is not merely an isostatic rebound to unroofing of the thickened crust. Treloar et al. (1991) suggest a rapid uplift and exhumation through extensional faulting within the upper part of the Indian plate and the suture zone, and erosion of topographic high. They regard that during this later period the Indus Suture ceased to act as a south-verging thrust fault but began to act as a north-side-down normal fault.

The uplift and denudation episodes are partly recorded in the molasse sediments deposited in the Himalayan foreland basin. The sedimentation of the Murree Formation commenced during the Late Eocene (Bossart and Ottiger 1989), presumably in response to Kohistan-India collision. Pronounced deposition of coarse silisiclastic sediments began in the Siwalik Group at 18.3 Ma (Johnson et al. 1987). The heavy mineral population of the Siwalik sediments contains strongly pleochroic orthopyroxene, blue-green hornblende (Johnson et al. 1985), and Mg-Fe-Ca garnet (Bajwa et al. 1987). These minerals are typical of the Chilas gabbronorites, the southern amphibolite belt, and the Jijal Complex, respectively, of the Kohistan arc. The blue-green hornblende shows a rapid increase in the Siwalik sediments by 11 Ma (Cerveny et al. 1989). This suggests that a pronounced uplift was underway in the Siwalik source area, and Kohistan had been connected to the Indus River drainage and exhumed to deep levels, probably by 19 Ma but certainly by 11 Ma ago (Jan and Karim 1995).

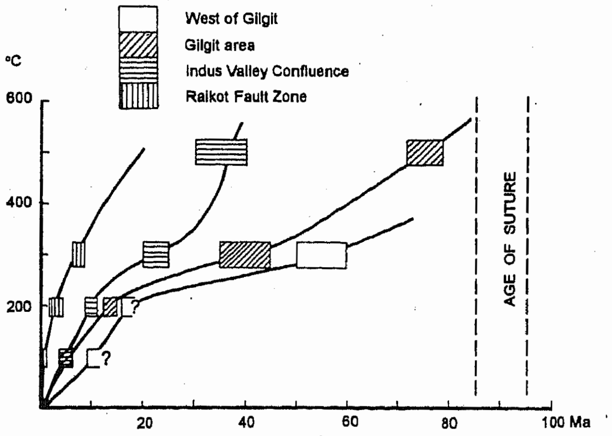


Fig. 7.10. Temperature-time plot showing cooling curves for rocks from western Kohistan, Gilgit area, Indus Valley confluence and rocks adjacent to the Raikot Fault zone on the western margin of the Nanga Parbat syntaxis. Late and increased cooling in eastern Kohistan than in western Kohistan in last 20 Ma can be related to exhumation along the Nanga Parbat syntaxis. (From Treloar et al. 1989).

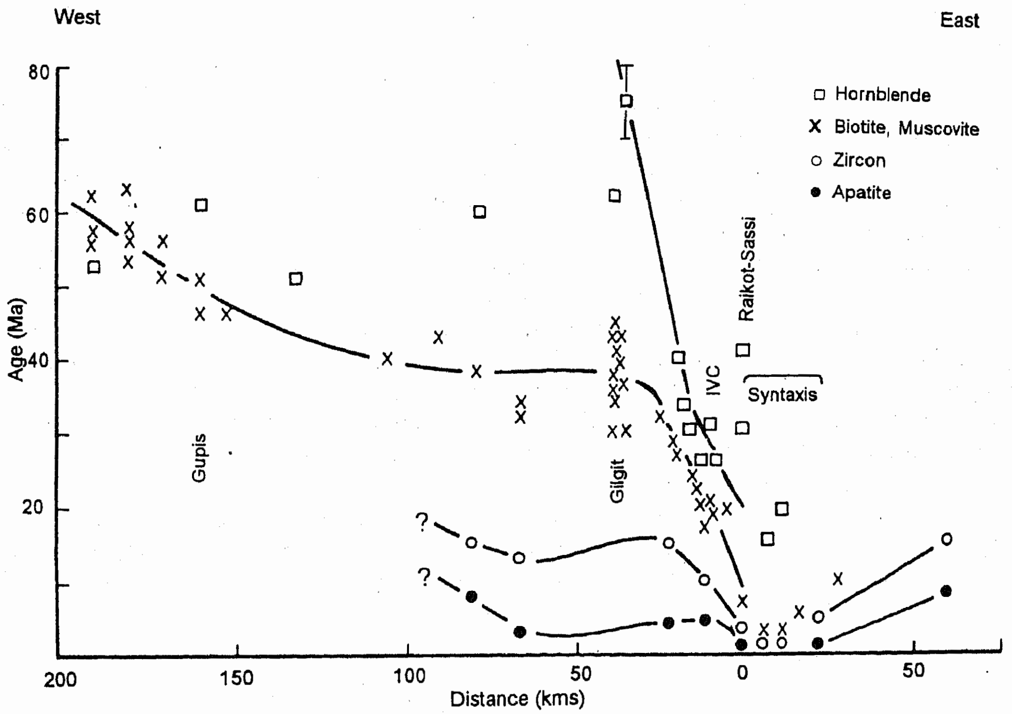


Fig. 7.11. Plot of age against distance west of the Raikot Fault zone. Eastward decrease in cooling ages towards Nanga Parbat syntaxis may be noted. (From Treloar et al. 1989).

INDIAN PLATE

The Indian plate within Pakistan has a long and complicated thermo-tectonic history dating back to at least the Early Proterozoic and, possibly, Late Archean. A wide range of metamorphic assemblages from subgreenschist facies to granulite facies, formed mostly under medium-pressure conditions, have been reported, especially in the northwest Himalayas and these have been the focus of many investigations. Field and petrographic studies, supported by some radiometric dates, suggest that the region has been affected by several episodes of deformation, and more than one cycle of regional metamorphism. For want of detailed information, especially geochronological, the metamorphism of the Indian plate is here described under pre-Himalayan and Himalayan categories. But before going into metamorphism, we would like to briefly recapitulate the stratigraphic set-up of the northwestern margin of the Indian plate in northern Pakistan (earlier discussed in Chapters 4, 5 and 6).

The Indian plate rocks to the south of the Indus Suture have been divided into hinterland and foreland (Pogue et al. 1996) or internal and external zone sequences by Treloar et al. (1991). The internal zone rocks crop out to the north, and in the hanging wall of the Panjal-Khairabad Thrust and are metamorphosed. The external zone rocks occur to the south of the Panjal-Khairabad Thrust and are unmetamorphosed. The latter range from Precambrian to (dominantly calcareous) Mesozoic-Eocene Tethyan shelf sediments (Calkins et al. 1975, Yeats and Hussain 1987). These were thrust southward along the Main Boundary Thrust onto Miocene and younger Siwalik molasse deposited in the foreland basins (Khan et al. 1985, 1987, 1988, Johnson et al. 1985, 1986, Burbank et al. 1986).

Within the crystalline internal zone, the Indian plate can be divided into three lithostratigraphic packages (Treloar et al. 1991). At the base is a Late Archean(?) to Proterozoic basement complex of gneisses, calc-silicates and amphibolites, with Early Proterozoic (up to 2.2 Ba) ages. In Besham area, this group of rocks was intruded by a suite of 1,500 Ma sodic granite gneisses (Le Fortune et al. 1992). These basement rocks may have passed through more than one episode of deformation and high-grade metamorphism before the deposition of Late Precambrian and Early Paleozoic sediments (DiPietro et al., in prep.). These include the argillatic-turbiditic sediments of the Tanawal, Hazara, Manglaur and Jobra Formations, the arenaceous to carbonaceous sediments of the Abbottabad Formation, and the Cambro-Ordovician granitoids (Mansehra, Swat). Baig et al. (1989) suggested that these rocks underwent a Paleozoic low-grade metamorphism. Unconformably overlying these is a sequence of pelites, calc-pelites, marbles and amphibolites of Carboniferous to Early Mesozoic: the Panjal sequence of Kaghan (Greco and Spencer, 1993) and Kashala Formation of Swat (DiPietro et al., in prep.). By analogy with the neighbouring regions, these units must have been overlain, before collision, by a thick sequence of Late Mesozoic to Early Eocene limestones and sandstones deposited on the northern margin of the Indian plate; however, none have been preserved within the internal zone of the Pakistan Himalayas (Treloar et al. 1991). Analysis of the sedimentary thicknesses to both east and west of Nanga Parbat suggests that there was a 10 km thick sequence of Cambrian to Eocene sediments overlying some 6 km thick Precambrian to Cambrian sequence. Of this, about 4 km are Permo-Trias to Eocene in age (Brookfield 1993, Treloar 1995a).

Pre-Himalayan Metamorphism

Kirana Hills and Nagar Parkar

Little attention has been paid to the metamorphism of the Indian plate in Pakistan outside the Himalayan orogen. From the description given elsewhere in this volume, it is obvious that the rocks of the Salt Range do not display mineral parageneses which can be attributed to regional metamorphism. Further south in the Kirana Hills, however, pelitic and arenaceous lithologies are represented by low-grade phyllites, slates and quartzites (Alam et al. 1992). Some of the mafic rocks associated with these metasedimentary rocks contain abundant amphibole, epidote, and chlorite (Khan and Chaudhry 1991). But it is not clear whether this paragenesis is a product of deuteric alteration or the low-grade regional metamorphism which affected the sedimentary rocks. The 870 ± 40 Ma Rb-Sr isochron age reported for the igneous suite (Davies and Crawford 1971) has been interpreted by Baig et al. (1988) as suggesting a metamorphic/orogenic episode.

The Nagar Parkar area of southeastern Sind contains Late Proterozoic granites intruding a basement consisting of mafic rocks. The granites are variably weathered, but are undeformed and unmetamorphosed. The mafic basement consists essentially of volcanic and plutonic rocks. It shows folding, shear deformation, and a greenschist facies metamorphic paragenesis of amphibole (? actinolite), epidote and chlorite. Geochronological data are lacking but the granites have been tentatively correlated with the 740 ± 20 Ma plutons in the neighboring Rajasthan (Butt et al. 1994). It is likely that the low-grade metamorphism in the basement here and in the Kirana Hills is related either to the Delhi (1450 Ma) Orogeny or to the post-Delhi (850–750 Ma) event of Roy (1988).

Himalayan region

Field and petrographic studies provide sufficient evidence to suggest that the main metamorphic recrystallisation in the Precambrian rocks of the Himalayas is associated with the Tertiary Himalayan Orogeny (Greco and Spencer 1993). For a long time, no clear evidence of pre-Himalayan metamorphism was reported in the area, mainly because of a lack of sufficiently detailed studies. Kazmi and Rana (1982) were the first to indicate a Precambrian magmatism, metamorphism and orogeny in southern Hazara –Attock region. Baig et al. (1987, 1988) and Williams et al. (1988) provided field evidence of a Precambrian metamorphism in northern Pakistan. Baig et al. noted that the rocks below the Cambrian unconformity in Abbottabad area show low-grade metamorphism and deformation that is truncated at the unconformity. Baig et al. (1988, 1989) used this evidence to propose that the rocks were deformed during a Late Precambrian (900–600 Ma) "Hazaran" Orogeny that they equated with the Pan-African Orogeny. Since then, further evidences of pre-Himalayan metamorphism and deformation from northern Pakistan have been documented (Chaudhry et al. 1989, Treloar et al 1989c). DiPietro et al. (in prep.) suggest that each of the five phases of magmatism (Archean?, Early Proterozoic, Early Paleozoic, Late Paleozoic, Paleogene) in the Indian Plate metasedimentary rock sequence of northern Pakistan was associated with deformation of varying degrees and style, but only the Archean and Paleogene intrusions appear to be clearly associated with regional high-grade metamorphism. Baig et al. (1991) noted that the garnet to sillimanite grade metamorphism in the Besham basement occurred before the deposition of the unconformably overlying Karora Group. The conglomerate at the base of the Karora Group (?Early Proterozoic according to Dipietro et al. in prep.) contains clasts of Besham gneisses in a phyllitic matrix. More recently, Treloar (1995a) has expressed the opinion that the Precambrian basement in northern Pakistan may not have been affected at all by the Early Tertiary Himalayan metamorphism.

Based on published geochronological data and new K-Ar and Ar-Ar ages, Treloar et al. (1989c) and Treloar and Rex (1990a, b) suggested that the Indian plate in northern Pakistan has had an extended thermal history dating back to the Early Proterozoic. They reported several pre-Himalayan dates, the oldest of which is a 1,920 ± 24 Ma Ar-Ar hornblende age for an amphibolite pod in a granite north of Besham. DiPietro et al. (in prep.) regard this to be the age of intrusion rather than metamorphism because, according to them, there is no evidence for amphibolite facies metamorphism in the Early Proterozoic rock sequence outside the zone of strong Paleogene metamorphism. We disagree with this suggestion because (1) pre-Himalayan high-grade metamorphism has been reported from some other places in NW Himalaya (e.g., Nanga Parbat), and (2) the abundance of amphibole suggests that the paragenesis is metamorphic. In fact, this age is similar to several others from the region and may point to one or more thermo-tectonic episodes. Zeitler et al. (1989), for instance, reported a U-Pb zircon age of 1,850 Ma for the Iskere Gneiss of the Nanga Parbat Massif, and Shah et al. (1992) suggested from Pb isotope ratios that the Besham Pb-Zn deposits formed 2,120 to 2,200 Ma ago and were re-equilibrated during metamorphism at 1,950 Ma.

On the basis of extensive field, petrographic, and Ar-Ar age data, Baig (1991) reported several pre-Himalayan thermo-tectonic episodes for the Besham area which preserves amphibolite facies metamorphism. Ar-Ar dates of the amphiboles can be grouped into 2,031 ± 6 to 1,997 ± 8 Ma, 1,950 ± 3 Ma, and 1,887 ± 5 to 1,865 ± 3 Ma. These were followed by sodic granite intrusions at 1,517 ± 3 Ma, and the Hazaran (Pan-African) Orogeny at 900 to 600 Ma. More recently, DiPietro et al. (in prep.) have argued on stratigraphic grounds that the particular metamorphism which Baig et al. (1988) and Baig (1991) related to the Hazaran Orogeny may in fact be Early Proterozoic (pre-Tanawal). They believe that the whole-rock Rb-Sr ages of 951 to 728 Ma for some Hazara rocks (Crawford and Davies 1975) may not suggest the true age of this earlier metamorphism, but instead record a Late Proterozoic low-grade metamorphic overprint.

Supported by detailed field data, DiPietro et al. (in prep.) argue that (1) there are major unconformities above the Archean(?) and Early Proterozoic(?) rock sequences of the Indian plate in Pakistan, (2) the Archean(?) rocks are characterised by plutonism, deformation, and possibly high-grade metamorphism, and (3) Early Proterozoic(?) rocks are characterised by plutonism, deformation, and possibly low-grade metamorphism.

Chaudhry et al. (1989), on the other hand, argue that pre-Himalayan metamorphic conditions in the Mansehra area may have reached at least amphibolite facies. They think that the Tanawal Formation underwent chlorite- to sillimanite grade of regional metamorphism before it was intruded by the Cambrian Mansehra Granite. The pluton is in no way confined to the thermal axis of the area and isograds of regional metamorphism are abruptly truncated by it in many places. A thermal aureole has been superimposed at an angle on the regional metamorphism. On the basis of ages of higher Himalayan rocks in India (Mehta 1980), Chaudhry et al. (1989) regard this metamorphism as 1,800–1,900 Ma old and related to Karelian Orogeny. It should be noted, however, that most authors consider this metamorphism as Himalayan in age, and DiPietro et al. (in prep.) regard the Tanawal Formation as Late Proterozoic.

The Nanga Parbat-Haramosh Massif is a structural half window. Within its core, Indian plate gneisses are updomed from beneath a cover of overthrust volcanics of the Kohistan-Ladakh magmatic arc. A range of metamorphic parageneses of low- to high-pressure amphibolite to granulite facies has been reported from the Nanga Parbat Massif. The calcareous assemblages include the unusual parageneses anorthite + wollastonite and anorthite + calcite (Misch 1964). The high-grade metamorphism and associated melting has been described by most workers to be of Tertiary age. Treloar et al. (1995) and Wheeler et al. (1995) have, however, argued that the Early Proterozoic basement (Iskere Gneisses) and Late Proterozoic cover (Shengus Gneisses) in Nanga Parbat have been affected by polyphase (Tertiary and pre-Tertiary) deformation, metamorphism, and magmatism.

The gneisses, both para and ortho with many textural and lithological variants, are commonly migmatised and show strong mineral shape fabrics parallel to banding. A typical metamorphic paragenesis in the Iskere Gneiss consists of orthoclase-biotite-muscovite-plagioclase-quartz; kyanite and garnet occur locally but the association kyanite + orthoclase has not been reported. The main paragenesis in the Shengus Gneiss comprises quartz-plagioclase-orthoclase-garnet-biotite-kyanite and/or sillimanite-rutile; garnet and orthoclase form porphyroblasts with irregular outlines (Wheeler et al. 1995). Both types of gneisses contain mafic sheets, which truncate the migmatitic banding in the gneisses locally. Moreover, the mafic sheets (now amphibolites) do not display any evidence of migmatisation or granulite facies metamorphism. So Treloar et al. (1995) and Wheeler et al. (1995) suggested that the peak metamorphism and migmatisation in the host gneisses predates the intrusion of the discordant sheets. This relationship is commonly obscured by subsequent (i.e., Himalayan) deformation at amphibolite facies which formed LS fabric (tectonites with schistosity and lineation) in both gneisses and mafic sheets and transposed earlier structures. Since (1) the generation and emplacement of basic magmas in tectonic settings involving compression and crustal thickening is very difficult, and (2) there is no record of Tertiary basic magmatism in the NW Himalaya, Treloar et al. (1995) and Wheeler et al. (1995) assumed that the mafic sheets were of Permo-Triassic (Panjal) or Precambrian age, and the main Nanga Parbat metamorphism as pre-Triassic. Thus, like the Scourian dykes of the Lewisian Complex, the mafic sheets separate two major thermo-tectonic events (Fig. 7.12).

Younger episodes of orogeny have been proposed. According to Baig et al. (1989), one of these with associated metamorphism, deformation, and magmatism is marked by the Cambro-Ordovician unconformity. To this phase belong the 550–450 Ma granitic plutons of Swat, Mansehra, Kaghan, Kashmir, Nanga Parbat, and Besham. No evidence in favor of strong deformation or regional high-grade metamorphism has so far been presented for this event, but it was responsible for the uplift and partial or complete removal of the Ambar Formation prior to the deposition of the Ordovician Misri Banda quartzite.

An episode of deformation and metamorphism has been ascribed to the Late Paleozoic when extensive plutonism and volcanism took place during the Carboniferous to Permian (e.g., Ambela granite, Panjal volcanics). Chaudhry et al. (1989) noted that in the Kaghan Valley the Panjal Volcanics and older rocks are metamorphosed, whereas the Triassic Malkandi limestone and younger sequence is unmetamorphosed. The grade of metamorphism in the Panjal Volcanics does not appear to surpass greenschist facies, except where overprinted by a higher-grade Himalayan one. Chaudhry et al. (1989) relate these events to a Permo-Triassic ('Panjal') orogeny, whereas Baig (1991) relates them to early rifting of the Cimmerian microcontinent from Gondwana.

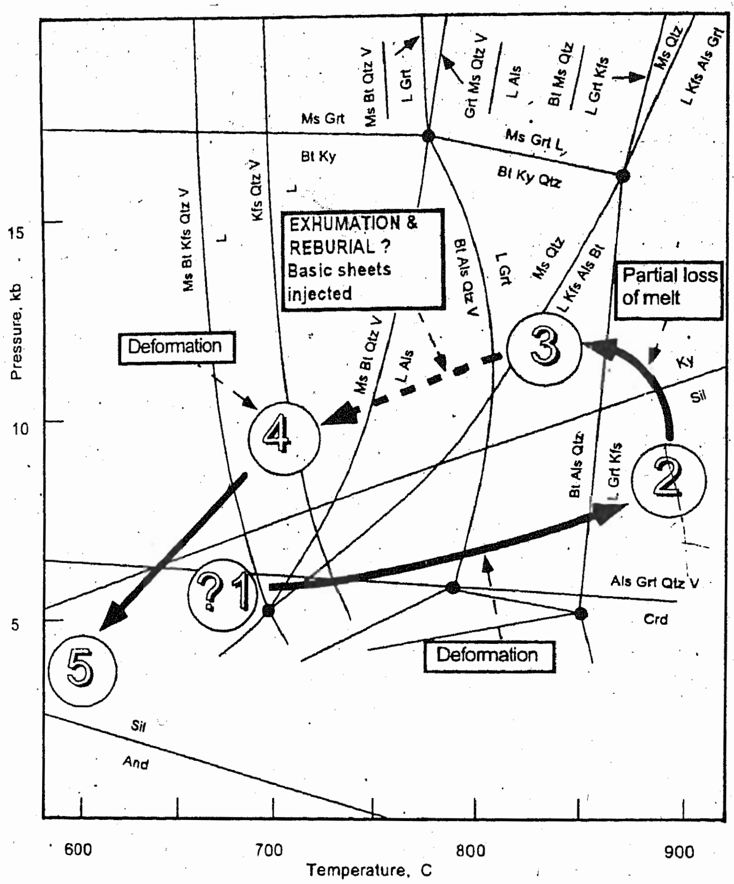


Fig. 7.12. P-T grid for the Nanga Parbat, showing important reactions in pelites at high temperatures (after Vielzeuf and Holloway 1988), and the approximate P-T 'Path'deduced from mineralogy and textures in the Shengus gneisses. The univariant reaction lines which involve ferromagnesian phases are simplified representations of divariant fields, assuming a bulk rock Mg/(Mg + Fe) ration of 0.5. In addition, some of the reaction lines were determined experimentally, not in a pure KFMASH system, but in a system containing albitic plagioclase, although its involvement is not indicated explicitly. The gap between stages 3 and 4 of the P-T path is to emphassize that these are parts of separate metamorphic cycles, separated in time by the episode of basic sheet emplacement (modified from Wheeler et al. 1995).

Himalayan Metamorphism

Related to the Himalayan Orogeny and tectonics, widespread regional metamorphism and deformation occurred during the Tertiary. Substantial information exists on deformation and metamorphism of different sectors of the Indian plate margin in the Himalayas of northern Pakistan. A synthesis of the data shows that the metamorphism ranges from very low-grade (anchi zone) in the Himalayan foothills to high-grade (granulite facies) in the Central Himalayan crystallines. The style and structural history of the main phase Early Tertiary metamorphism throughout northern Pakistan (i.e., Swat, Hazara, Kaghan, Azad Kashmir) are the same. This metamorphism has three particular characteristics:

1) Peak metamorphism was short-lived and occurred soon after the Paleocene-Early Eocene collision; i.e., within 5 to 15 Ma, depending upon how the timing of the collision is constrained;

2) Metamorphism was generally of high- to medium-high pressure Barrovian (kyanite-sillimanite)-type. It shows a positive dP/dT, suggesting growth under increasing pressure, and spiral garnet indicates that its growth was synchronous with deformation;

3) It was generally inverted, i.e., higher grade rocks, formed at greater depth, structurally overlie lower grade rocks. This probably resulted mainly from stacking of thrust slices in much of the area (i.e., Kaghan, Hazara, Besham, Swat).

Metamorphism in northern Pakistan differs from the rest of the Himalayas in one significant way. There is no evidence of the Late Oligocene to Early Miocene metamorphism so extensively developed in the footwall of the Main Central Thrust in the rest of the Himalayas (Le Fort 1986, Hodges and Silverberg 1988, Hubbard and Harrison 1989, Verma 1989). On the other hand, there is evidence of a young, low-pressure and high-temperature metamorphism associated with the rapid (up to 5–7 mm per year) exhumation of the Nanga Parbat over the past 15 Ma.

Metamorphism in Hazara-Besham-Lower Swat

Since the studies of Martin et al. (1962) and King (1964) in Lower Swat, Kazmi et al. (1984) in Swat and Calkins et al. (1975) in Hazara, a large amount of data has been published on the stratigraphy, structure, deformation, and metamorphism of this region. Treloar et al. (1989a, b) and Baig (1990) showed that the Besham basement complex was separated from the adjacent Hazara and Swat areas by post-metamorphic faults (see Fig. 6.2). The Besham basement, according to them, displays a regional Early Proterozoic amphibolite facies metamorphism and only a low-grade Paleogene (Himalayan) metamorphism, whereas the Swat and Hazara rocks show a regional Paleogene metamorphism. The field studies of DiPietro et al. (in prep.) reveal that the stratigraphy across the faults in the Besham area can be correlated without a break in metamorphic fabric. They think that metamorphism across the area is of the same age and, because it affects Mesozoic rocks, is Paleogene. As mentioned earlier, Proterozoic rocks of the NW Himalaya have experienced only low-grade pre-Himalayan metamorphism and it is only the Archean that have been affected by high-grade metamorphism (DiPietro et al. in prep.).

Dominant structures in this area are a series of N-trending folds that die out or swing east or west along the southern margin of the area in conformity with the Khairabad Thrust. These folds deform foliation with which are associated an earlier set of folds (DiPietro et al., in prep.). The Himalayan metamorphism increases from south to north both in Hazara and Swat. Rocks in the northern margin of Peshawar Basin are of low-grade, with recognisable sedimentary structures and fossils. Garnet occurs sporadically between the latitude of Jowar in Buner and Darband in Hazara. The transition from greenschist to amphibolite facies within the Permian metavolcanic rocks occurs in this zone. Further north, the rocks are in amphibolite facies and garnet is widespread in rocks of appropriate bulk composition. Metamorphic intensity reaches kyanite and sillimanite grade in the Lower Swat region. In Hazara area also, a similar situation prevails; from chlorite grade near Abbottabad to kyanite and sillimanite grade in the north (for map showing metamorphic zones, see Calkins et al. 1975). The Hazara area is occupied by the Tanawal Formation with large intrusions of the Cambrian Mansehra Granite that have locally exacerbated the metamorphic picture. In the previous pages, we have described the structure and metamorphism of the Besham and Hazara areas in detail. In the following, we summarise the main aspects of metamorphism in the Swat area.

The Lower Swat-Buner rock sequence is exposed in a dome (Loe Sar Dome) in which EW-trending folds are superimposed generally on N-trending earlier folds (DiPietro 1990). The area consists of a Precambrian to Cambrian basement and Late Paleozoic-Mesozoic cover (DiPietro et al. 1993, and in prep.). The basement comprises amphibolite facies quartz-muscovite-potash feldspar-kyanite gneisses and tremolite marble. Some rocks show a late (?D_3) fibrolitic overprint. Metamorphosed at 600–700° C and 11 kbar, these display evidence of partial melting. The overlying rocks contain a variety of garnetiferous assemblages that include quartz-muscovite-biotite-garnet, quartz-biotite-garnet-potash feldspar, and horblende-quartz-plagioclase-garnet. Overlying these are calcite-quartz-muscovite-ilmenite-clinozoisite-garnet (± chloritoid ± chlorite) marbles and black phyllite. The highest grade reached in the cover sequence during the last stages of F_3 (>600° C, 9–11 kbar) was not much different than that in the basement (DiPietro 1990, 1991).

The sequence of deformation described from the Swat area by DiPietro and Lawrence (1991) and shown in Figure 7.13 is different than those reported for Hazara by Coward et al. (1988) and Treloar et al. (1989). The chronology of the latter-named authors is one of repeated syn-metamorphic, S-verging shears and folds (D_1, D_2), followed by the major phase of post-metamorphic thrust stacking (D_2 a), and WNW-verging folds and back thrusts (D_3). W-directed structures of an early phase have not been reported from Hazara. By contrast, the Lower Swat area records syn-metamorphic, W-verging folds followed by late-metamorphic S-verging folds and the development of a structural dome in which the deepest seated rocks are at the highest grade. There is no evidence for back thrusting in Lower Swat according to DiPietro and Lawrence (1991). They suspected that following F_4 (=D_2 a) deformation, large S-directed thrusts, initiated in the foreland, carried the Mansehra and Lower Swat rocks and emplaced them above the Besham block. Later uplifting along the

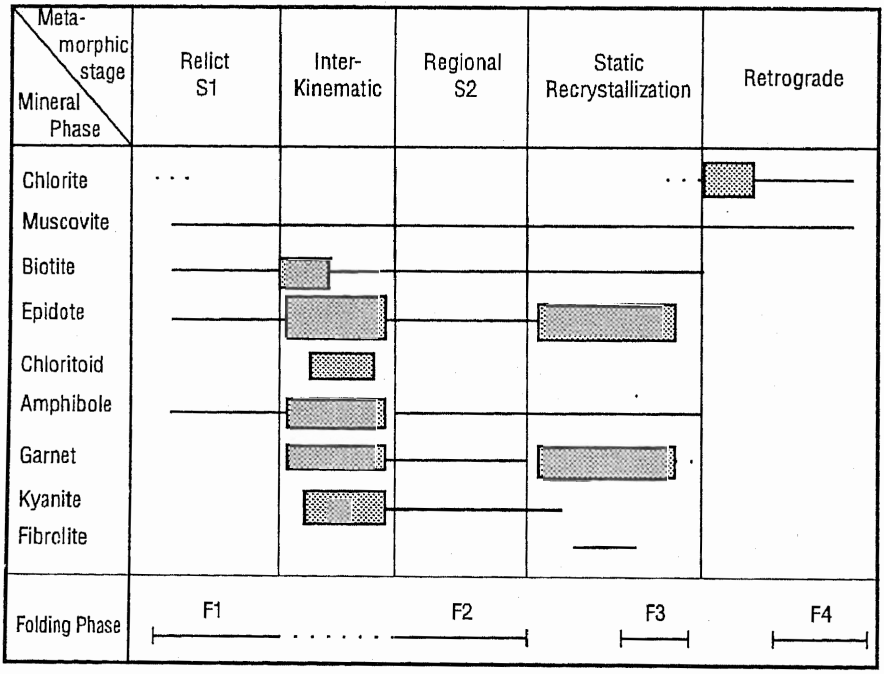


Fig. 7.13. Inferred mineral stability in relation to the deformational phases for different bulk compositions in Lower Swat. Lines represent mineral stability and boxes represent significant porphyroblastic growth. For timing, see Figure 7.18 (from DiPietro and Lawrence 1991).

border faults exhumed the Besham block. But the latest studies of DiPietro et al. (in prep.) cast doubt on the validity of this idea. The deformation history recorded in the Swat area (DiPietro 1990) correlates well with that of Hazara (Bossart et al. 1988). Several others have also suggested similar structural-metamorphic history for the two areas. The trend of the Murree and Panjal Thrusts east of the Hazara Syntaxis approximates major F_2 and F_3 folds in the Lower Swat sequence. Thus, the metamorphism and early structures in Lower Swat may represent an inner zone expression of the early deformation which affected the northwestern part of the Indian plate in Pakistan (DiPietro and Lawrence 1991).

It is likely that the Lower Swat sequence extends westward into Malakand Agency and beyond. The Cambrian granitic gneisses and the overlying Late Paleozoic-Mesozoic cover exposed around Chakdara have been studied in detail. These rocks have been affected by three phases of deformation (D_1, D_2, D_3) during which S_1 (relict muscovite fold hinges), S_2 (dominant foliation) and S_3 (crenulation cleavage) fabrics developed with folds F_1, F_2, F_3. The S_1 fabric, developed in greenschist facies, was transformed during D_2, and S_2 was formed in amphibolite facies conditions. During D_3, S_2 was crenulated, S_1 and garnet were nearly destroyed and S_3 produced. This lower grade event was followed by retrograde metamorphism (Ahmad and Lawrence 1992). Common assemblages in the area are: quartz-muscovite-biotite-garnet-kyanite-plagioclase-epidote ± staurolite, hornblende-oligoclase-garnet-epidote-biotite-quartz, calcite-quartz-muscovite-zoisite-(± graphite ± chlorite ± garnet), and quartz-muscovite-biotite-zoisite-chlorite-graphite. Khan (1992) suggests metamorphic conditions of 535–660° C, 8–11 kbar in aluminous rocks and 590–705° C, 11–12 kbar in amphibolites, but lower estimates have been reported by Ahmad and Lawrence (1992). North of the Swat River in this area, the grade of metamorphism in the Indian plate rocks is greenschist facies. This has been interpreted to be caused by faulting along the river, but retrograde metamorphism related to extensional tectonics also merits consideration.

Metamorphism in Kaghan-Hazara Syntaxis

The Kaghan Valley and Hazara Syntaxis have been investigated by many workers, especially during the past two decades. This area exposes a virtually complete section of the Indian plate crust. As in Hazara and Swat, it comprises a Precambrian to Cambrian basement and a Paleozoic to Early Tertiary cover. The basement was presumably deformed and metamorphosed during pre-Himalayan thermo-tectonic events the record of which was obliterated by Himalayan metamorphism which affected the pre-Tertiary cover also. Extensive field data and petrographic details were provided on the area by Chaudhry et al. (1986, 1987). They reported a large number of metamorphic mineral assemblages developed from a wide range of rocks of igneous and sedimentary parentages. One highly unusual assemblage in a schist from the Babusar area consists of garnet-hornblende-chloritoid-chlorite-zoisite-phengite-calcite-quartz-plagioclase (Chamberlain et al. 1991). The intensity of metamorphism increases up valley (i.e., towards north) from very low grade in the core of the syntaxis (Bossart 1986) to high amphibolite- and eclogite facies beyond Naran (Pognante and Spencer 1991). Chaudhry et al. (1986) and Greco (1991) noted that the metamorphic isograds follow thrust contacts, which means that they are a consequence of tectonic evolution rather than of inverted thermal gradient or reaction isograds.

In accordance with the classical subdivisions of the Himalayas, the syntaxial area can also be divided into three elements: Higher Himalayan crystalline (HHC) nappe, Lesser Himalaya, and Sub-Himalaya (Greco 1991, Ghazanfar 1993, Greco and Spencer 1993, Chaudhry et al. 1994). Each one of these is composed of one or more tectono-stratigraphic units. In general, thrust units become older towards the uppermost tectonic elements. Figure 7.14 shows the distribution of the metamorphic facies and large-scale interpretation of the main structures in the area. In the following, the three elements are briefly described.

The deepest crustal level is represented by the HHC basement, with a predominantly carbonatic Phanerozoic cover. The HHC is placed over the Lesser Himalayan detrital formations along the Main Central Thrust. The latter (MCT) has been described in detail by Chaudhry and Ghazanfar (1987) and Spencer (1995). Polyphase folding, penetrative schistosity, and ductile shearing affect the nappe structure of the basement and cover rocks. The deformation style is very ductile at the base of the HHC. The Lesser Himalayan and Sub-Himalayan packages are separated from each other by the Main Boundary Thrust (locally named as Murree Fault). Both these packages are folded, internally imbricated and, in the case of the Lesser Himalaya, the cover may be detached from the basement.

The Sub-Himalayan element , which consists of the Murree Formation, is exposed in the core of the syntaxis. These Early Tertiary detrital rocks have been little affected by metamorphism, with development of stilpnomelane in some arenaceous rocks. Bossart (1986) suggested that the very low grade (prehnite-pumpellyite facies) metamorphism developed in the stratigraphic base of the detrital sequence due to sedimentary thickening. This means that the style of metamorphism here is entirely different than the one in the other two packages.

The Lesser Himalaya consists of Proterozoic to Cambrian Salkhala unit (metapelites, quartzites, marbles) and augen gneisses, and a cover of Carboniferous to Triassic Panjal unit (pelites, conglomerates, volcanics and calcareous rocks). Metamorphism in these rocks ranges from very low-grade to upper greenschist facies/epidote amphibolite facies (garnet-zone). There is a metamorphic transition from the amphibolite facies ductile deformation in the hanging wall (HHC) of the MCT to greenschist facies in the footwall (Lesser Himalaya). In these mylonitised rocks, hornblende remained stable and only garnet grew after deformation. Greco (1991) estimates conditions of 550–650° C and 6–9 kbar for this MCT-related deformation which, along with internal imbrication in the footwall sequence, produced some inverted metamorphism.

The HHC occupies Upper Kaghan and consists of a basement of granitic gneisses intruded into older paragneisses (The Sharda Group of Chaudhry et al. 1987) which have a complex deformation history. These may be overlain by (?) Cambro-Ordovician units. The overlying cover sequence consists of interlayered metasediments and amphibolites in the base, and calcareous and pelitic top with interlayered mafic flows. Recumbent folding and ductile thrusting resulted in several nappes that were subsequently refolded into late-stage domes (e.g., Besal) and basins (e.g., Burawai and Bela). Both the basement and cover rocks have experienced Barrovian-type syn- to post-deformational metamorphism with peak conditions suitable for partial melting (Figs.7.14 and 7.15). Locally, the mafic rocks of the cover were converted to eclogites at 650 ± 50° C and 13–18 kbar (Pognante and Spencer 1991, Spencer et al. 1990). These consist of omphacite, garnet (grossular 15–33 mole%, pyrope 12–22 mole%), quartz, rutile ± phengite. With a Sm-Nd age of 49 Ma (Tonarini et al. 1993), it can thus be concluded that by this time the Indian plate cover in Kaghan was buried to a depth of more than 50 km beneath the Kohistan arc.

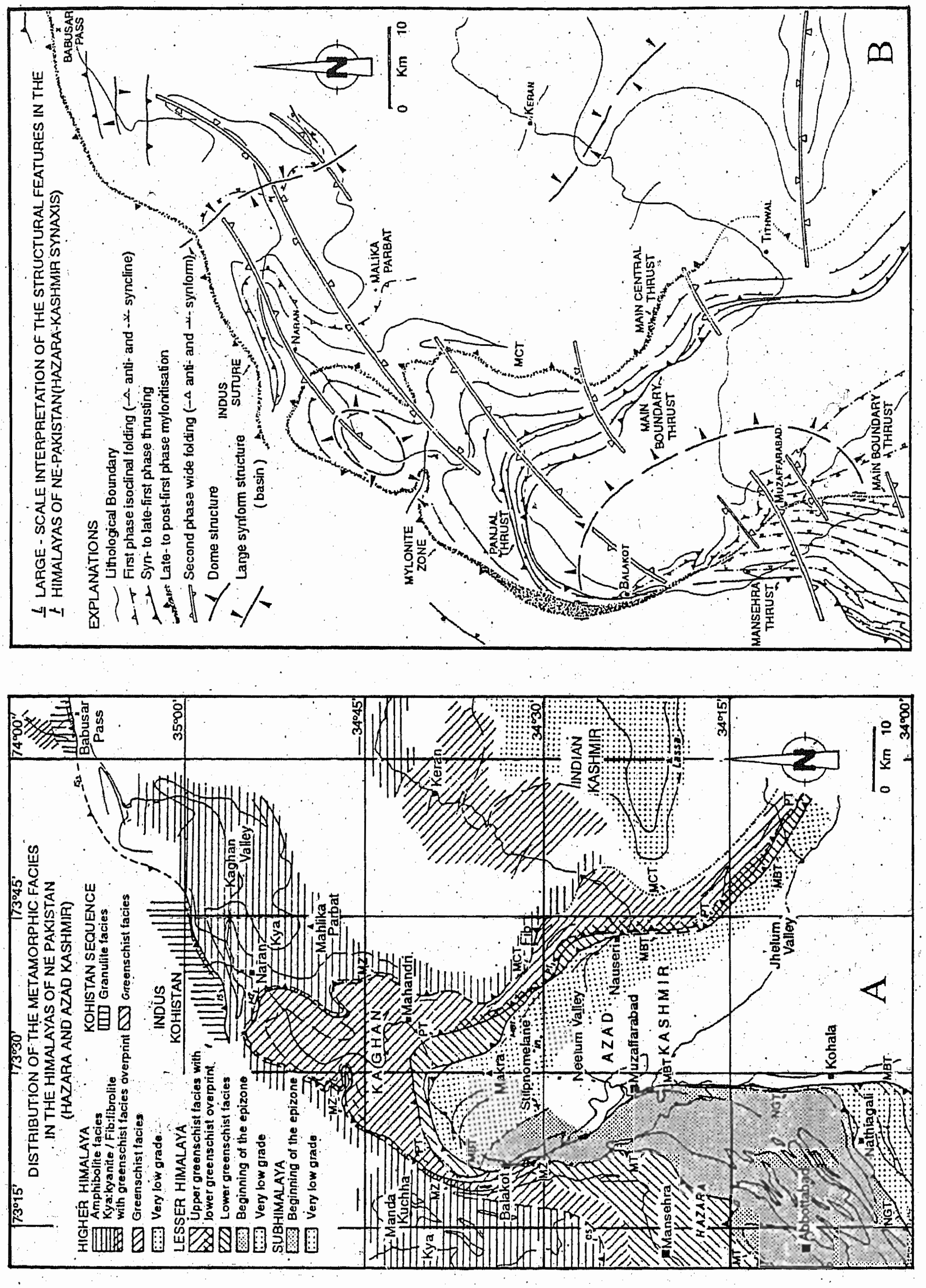


Fig. 7.14. A) Distribution of metmorphic facies, and **B)** large scale interpretation of the main structures, in the Hazara Syntaxis area (modified from Greco 1991, and Greco and Spencer 1993, respectively).

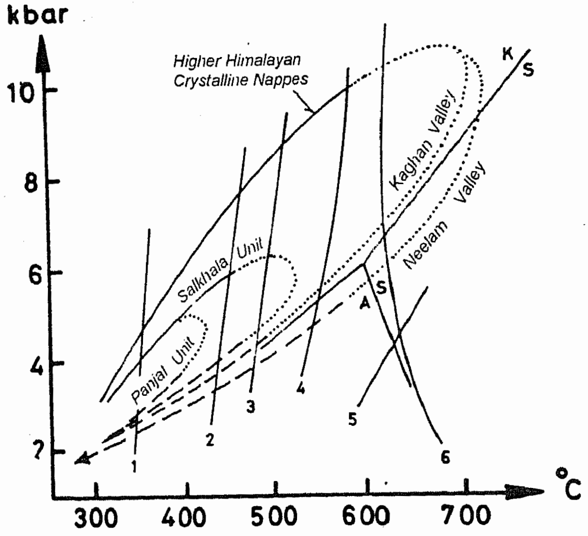


Fig. 7.15. Pressure-temperature trajectories for metamorphic rocks east of the Hazara Syntaxis. Metamorphism may have started during the Late Cretaceous subducion of the Indian plate, but the peak metamorphic conditions prevailed soon after 55–58 Ma collision between the Kohiatan arc and Indian plate. This was followed by uplift together with a series of thrusts (MCT, Panjal, MBT) in Miocene. Phase boundaries shown are: 1: pumpellyite-actinolite, 2: stilpnomelane-staurolite, 3: chlorite-hornblende, 4: chloritoid-staurolite, 5: staurolite-biotite, 6: partial melting in gneisses. A = andalusite, K = kyanite, S = sillimanite (modified from Greco 1991).

The geology of the Neelum Valley is an eastward extension of that of Kaghan. Here too an old basement was metamorphosed and migmatised during Early Proterozoic before the deposition of Salkhala Formation. This Precambrian sequence was intruded by Cambro-Ordovician granites described in Chapter 6. These were followed by deposition of Late Paleozoic and Early Mesozoic rocks, including volcanics and dykes of Panjal affinity. The entire area was affected by deformation and chlorite to sillimanite grade Barrovian-type regional metamorphism during the Himalayan Orogeny (Fontan and Schauppe 1994). Like upper Kaghan, there are local eclogitic rocks (Gumot Nala) yet to be studied.

Nanga Parbat-Haramosh Massif

Since the early studies of Misch (1935, 1949, 1964), the Nanga Parbat-Haramosh Massif has been studied by many workers. For references and review, the reader may consult Smith et al. (1992, 1994), Zeitler et al. (1993), Treloar et al. (1995), and Wheeler et al. (1995). It is characterised by a very rapid rate of uplift, accelerating to over 5 mm a year currently, and an equally rapid rate of denudation. It was uplifted about 30 km during the Himalayan Orogeny, of which at least 6 km occurred in the past one million years (Zeitler 1985). Tectonic controls on the uplift are poorly understood (Butler and Prior 1988, Zeitler and Chamberlain 1991). The massif has passed through a prolonged thermo-tectonic history dating back to Early Proterozoic. Hence, there is a difference of opinion on timing of the development of the high-grade metamorphic mineral parageneses present in the massif. Some regard that the parageneses preserve Precambrian thermo-tectonic episodes with local overprinting of the Himalayan metamorphism. Others have argued in favour of widespread recrystallisation and re-metamorphism during the Paleogene and/or Neogene. It has been known since long that on the one hand the massif preserves medium-P (kyanite-sillimanite) metamorphic assemblages and on the other there are low-P (sillimanite-cordierite) assemblages.

It was shown in the previous sections that the northwestern edge of the Indian plate in Pakistan underwent strong deformation and up to kyanite/sillimanite-grade metamorphism during the Himalayan Orogeny. It is not surprising at all, then, if the metamorphic pargeneses in the Nanga Parbat are also presumed to be related to the Himalayan episodes. Wheeler et al. (1995), however, suggested that the massif passed through at least three events of deformation-metamorphism and two of melting (Fig. 7.12). They argued that much of the Indus River section of the massif is occupied by migmatised gneisses metamorphosed in granulite facies before the intrusion of Permo-Triassic or older mafic dykes. This package was strongly deformed, with intense recrystallisation at amphibolite facies, during the Early Tertiary collision, producing N-S-trending tectonites with schistosity and lineations which are locally mylonitic. This strong fabric was folded into a tight antiform trending NNE. No new fabric was produced during this folding that accompanied syntaxis growth (Treloar et al. 1991), but shearing on the western margin produced steeply plunging lineations, indicating relative uplift of the massif.

A distinction between the Himalayan and older metamorphic episodes is not so simple. Apart from prolonged deformational and thermal history which is still in progress, the metamorphic conditions following the subduction of India under Kohistan were similar to those prevailing on the retrograde path of the early metamorphism (Fig. 7.12). According to Wheeler et al. (1995) and Treloar et al. (1995), kyanite was stable during the early Himalayan deformation and sillimanite grew after deformation ceased. It is worth noting that the parageneses that these authors consider pre-Himalayan contain zoned garnet porphyroblasts suggesting growth during crustal thickening, i.e., increasing PT as in Kaghan and Swat (Khattak and Stakes 1992). Because of these complexities and the possibility of a Neogene metamorphic overprint, Wheeler et al. (1995) lay stress on combining detailed field studies with laboratory data.

The metamorphic mineral assemblages of the Indus-Astor part of the massif have been studied by many worker (e.g., Chamberlain et al. 1989, Madin et al. 1989, Khattak and Stakes 1992, Wheeler et al. 1995). As mentioned in a previous section, many parageneses have developed from pelitic, granitic, mafic, and calcareous precursors. There are kyanite/sillimanite (± plagioclase ± K-feldspar ± biotite ± garnet) gneisses and schists, a wide range of calc-silicates (± biotite ± rutile ± wollastonite), and amphibolites (± garnet ± clinopyroxene ± quartz ± epidote ± rutile). The highest grade metamorphic conditions in the area exceeded 10 kbar and 700° C (Khattak and Stakes 1992). The eastern part of the massif to the south of Astor is also occupied by similar rocks, i.e., kyanite/sillimanite schists and gneisses, garnet-staurolite schists, calc-silicates and marbles, and garnetiferous amphibolites (± clinopyroxene). These rocks apparently pass to the south into those of upper Neelum Valley and Kaghan. But in the Babusar area, Chamberlain et al. (1991) have reported lower grade rocks in the vicinity of the Main Mantle Thrust. Misch (1964) reported the stable association of wollastonite and anorthite from the higher reaches of the Nanga Parbat, possibly resulting from the mineral reaction garnet + quartz = 2 wollastonite + anorthite.

Migmatitic gneisses (containing quartz, plagioclase, micas, kyanite, garnet ± rutile ± K-feldspar), marbles, and metabasites (clinopyroxene-plagioclase-garnet-quartz ± rutile ± hornblende) have also been described from the northeastern edge of the massif in the Stak Valley (Pognante et al. 1993, Le Fort et al. 1995). Zanettin (1964) had previously reported the stable assemblage of zoisite and kyanite from the higher reaches of the Stak

Valley, probably as a result of 4 margarite + 3 quartz = 2 zoisite + 5 kyanite + 3 H_2O or margarite + 3 anorthite = 2 zoisite + 2 kyanite. Peak metamorphism (650–700° C, 8–13 kbar) here was followed by rapid-exhumation and cooling, with little medium- to low-pressure re-equilibration (Pognante et al. 1993). The high-pressure metamorphism has been related to rapid subduction and exhumation.

An entirely different association of rocks, apparently affected by Late Miocene-Pliocene metamorphism, has been documented in the northwestern part (Tato region) of the massif (Zeitler et al. 1993). Three metamorphic zones have been recognised. Between Liachar Thrust and Tato, pelitic rocks contain kyanite + garnet + muscovite + quartz. To the south, the metamorphic gradient changes sharply within 10 km to sillimanite + muscovite and, finally, sillimanite + potassium feldspar + cordierite in the core of the massif. Here the rocks are migmatised and contain abundant stringers, veins (1–5 cm thick) and dykes of leucogranite (former melts). These range from granite to granodiorite and tonalite, and consist of typical S-type assemblage: quartz, plagioclase, potash feldspar, biotite, sillimanite, cordierite, garnet, tourmaline. Veins and dykes of similar composition and age (less than 12 Ma), and tourmaline pegmatites have also been reported from other parts of the massif. A stock-size body has been described from the northwestern terminus of the massif (Butler et al. 1992). Zeitler et al. (1993) and Winslow et al. (1993) reported that the highest grade rocks formed at about 650° C, 6 kbar, and the melt was injected at about 600° C, 4 kbar.

Detailed geochronological studies of this area by Smith et al. (1992) and Zeitler et al. (1993; and references there in) clearly show a close relationship between rapid uplift, exhumation, metamorphism, partial melting, and fluid flow over the past dozen million years. Neither a traditional crustal-thickening model for metamorphism, nor the fluid-fluxing model for granite genesis are adequately applicable to these rocks. The Plio-Pleistocene episode of high-grade metamorphism and anatexis occurred during an interval in which the Nanga Parbat Massif was undergoing rapid denudation at a rate of about 7 mm per annum. In response to this rapid denudation, partially dehydrated lower crustal rocks could have melted by decompression at 30 km (Zeitler and Chamberlain 1991, Smith et al. 1992). Rising melts would have transferred heat to shallow levels, resulting in Late Neogene metamorphism. Treloar et al. (1995) question the regional extent of this metamorphism and argue that the low-pressure metamorphism may be restricted to zones of channelised fluid flow.

The trace element signatures of the intrusives are consistent with generation by vapour-absent melting of pelitic metasediments characterised by unusually high heat production (George et al. 1993). The composition of the Nanga Parbat protolith is appropriate for the derivation of the Neogene granite melts (Butler et al. 1992, George et al. 1993). This is also supported by U-Pb dating of the granites. The margins of the zircon grains have Neogene age whereas the cores, like the Iskere Gneisses, are 1.8 Ba old (Zeitler et al. 1993). However, Treloar et al. (1995) think that the massif rocks (granulites) had already been melted once during the Precambrian and a more plausible source for the young granites would be Indian shelf sediments thrust under the massif.

Timing of Collision

Recent geological studies have established that the Himalayan deformation and metamorphism are related to Tertiary tectonic processes involving continental collision and thrusting. The timing of the Kohistan-India collision has been constrained differently. From limited geochronological data coupled with structural-metamorphic studies in Swat, DiPietro and Lawrence (1991) and DiPietro et al. (in prep.) concluded that the Kohistan arc was emplaced onto India approximately 34 to 26 Ma ago. This estimate may be unrealistically late because there are evidences of the closure of the Neo-Tethys by the Eocene. Bossart and Ottiger (1989) noted that the early molasse in the Hazara syntaxial area is of latest Paleocene age and contains high-Cr chromite probably derived from the Kohistan rocks. Structural, stratigraphic, and sedimentological studies in northern Pakistan support the view that initial collision may have occurred about the Cretaceous-Paleocene boundary (Yeats and Hussain 1987, Beck et al. 1995, 1996, Pivnick and Wells 1996). Deformation of this age has been reported from Swat area by DiPietro et al. (in prep.). It is, therefore, probable that some kind of collisional activity on the northwestern edge of the Indian plate occurred at about 65 Ma (see Chapter 4).

Treloar (1995a, and in press) has argued that this may not necessarily be the timing of continent-continent (i.e., Kohistan-India) collision. Collision at 65 Ma would create a real space problem underneath Tibet. At a convergence rate of 15-20 cm per year, some 1500-2,000 km of the Indian plate continental crust will have been subducted under Tibet by 55 Ma. None of the proposed models for post-collision behaviour of the Asian plate (e.g., Dewey et al. 1987, Le Pichon et al. 1992) can provide a satisfactory explanation for this extra material. Thus, instead of truly marking initial collision between India and Asia (or Kohistan), the 65 Ma event may be one of temporary slowing of the rate of subduction of the Tethyan oceanic crust beneath Asia. It may have been associated with ophiolite emplacement, or consequent upon the attempted subduction of an active ridge, oceanic island, or a slice of continental crust located to the north of continental India. A candidate for the last possibility is the Kabul block (Treloar and Izatt 1993).

The 65 Ma age coincides with the well-documented age of the ophiolite emplacement onto the leading edge of the Indian plate. Brookfield and Reynolds (1981), Searle (1986) and Beck et al. (1995, 1996) have presented evidence for the Late Cretaceous to Early Paleocene emplacement of the Dras, Spongtang, and Waziristan ophiolites in Ladakh and northwest Pakistan, respectively. Ar-Ar dating of the Bela and Muslimbagh ophiolites and their metamorphic soles also suggests an emplacement age of 65 to 70 Ma (Mahmood et al. 1995, Gnos et al. 1996). Thus, we also subscribe to the view that the Early Eocene (50–55 Ma) slowing down of the motion of the Indian plate may actually be timing the collision of its northwestern edge with the Kohistan arc. Yoshida (1996) presents paleomagnetic evidence from northern Pakistan which constrains this collision at 58 Ma, but Beck et al. (1996) argue it may have happened a little earlier. In the following, we show that regional metamorphism occurred soon after collision (Treloar 1989, Zeitler 1989).

Timing of Metamorphism

The timing of metamorphism in the Himalayas of Pakistan has been dated more precisely than the timing of collision. Geochronological studies in Kaghan and Swat areas show that the principal phase of the Himalayan metamorphism preserved in the internal zone of the Indian plate in northern Pakistan is of Early Tertiary age. Several phases of deformation have also been reported from the region (Chaudhry and Ghazanfar 1987, Greco et al. 1989, Williams 1989, Baig 1990, DiPietro 1991, DiPietro and Lawrence 1991, Treloar et al. 1991). A summary of the deformation has been given in Chapter 4 and in a following section.

Sm-Nd dating of garnet-clinopyroxene eclogite from the upper Kaghan Valley determines the age of regional metamorphism at 49 ±7 Ma (Tonarini et al. 1993). Granite sheets in this area have U-Pb zircon ages of about 50 Ma (Zeitler and Chamberlain 1991). These granites cut the main phase regional fabric and are, apparently, a product of partial melting associated with the Early Tertiary regional metamorphism. Zircon rims within pre-Tertiary granites in the Kaghan and Swat areas were also dated by U-Pb method. The 47 ±3 Ma age of these has been inferred to represent a metamorphic overprint on older zircon cores (Smith et al. 1994). Ar-Ar dating of hornblendes from Swat and Kaghan shows that post-metamorphic cooling through the hornblende blocking temperatures of about 550–500° C occurred 35–40 Ma ago (Treloar and Rex 1990, Chamberlain et al. 1991). This is entirely consistent with the temperature-time paths based on K-feldspar thermochronology from Swat Kohistan. Krol et al. (1994) reported a rapid cooling (and hence unroofing) of the Kohistan arc at about 44 Ma. Treloar (in press) has provided further description of the mid Eocene uplift and erosive exhumation of the rocks of the Kohistan sequence.

A synthesis of geochronological and field data suggests that the area was affected by regional deformation and synchronous main phase metamorphism no later than 50 Ma, followed by a cooling of at least 10–20° C per Ma. Baig (1991) regarded that early Himalayan metamorphism and deformation in northern Pakistan was associated with melange emplacement and occurred between 70 to 64 Ma ago. DiPietro et al. (in prep.) also subscribe to this view and consider that peak metamorphic temperatures were obtained during the Paleocene-Eocene. For Swat area, however, DiPietro (1991) suggested that metamorphism was Eocene to Oligocene in age and recorded the collision of the Indian plate with the Kohistan arc. Chamberlain et al. (1991), Treloar (1995a) and Treloar et al. (1991) assigned an Eocene age (50 Ma) to the main phase of the Himalayan metamorphism. Such an age constraint implies that deformation, heating and metamorphism (to more than 650° C) took place soon after collision, that is within 15 Ma if collision occurred at 65 Ma and less than 10 Ma if it is constrained at 55–58 Ma (Treloar 1995a, and in press). This is a rather short interval of time for regional metamorphism which reached upper amphibolite facies and was at places accompanied by partial melting.

Post-metamorphic Cooling History

Recently performed radiometric studies in northern Pakistan decipher the cooling rates of the Indian plate rocks following the principal phase of the Himalayan regional metamorphism. Treloar et al. (1989c), Treloar and Rex (1989a, b) and Chamberlain et al. (1991) showed that cooling through 550–500° C hornblende Ar-Ar blocking temperature was achieved by 35–40 Ma in Swat and Hazara region. Similar ages were reported by Lawrence et al. (1989) and Baig (1990) from Besham and Swat. Other cooling ages include: 35 Ma muscovite Rb-Sr (blocking T = 500° C), 30 to 23 Ma K-Ar muscovite (T ≐ 350–400° C) and biotite (275–325° C) by Treloar and others (op.cit.); and fission-track ages of 23–20 Ma for zircon (T = 200 ± 25° C) and 16–19 Ma for apatite (100–110° C) by Zeitler (1985). Figure 7.16 shows the relationship of cooling with time. The resultant time-temperature curve exhibits that from 30 to 18 Ma (especially, 25 to 20 Ma) the rocks cooled at much faster rate than during the preceding and following periods. This stage of rapid cooling (300° C in 10 Ma or an average of 30° C per Ma) must equate with rapid unroofing and exhumation , at the end of which the rocks were at a temperature of

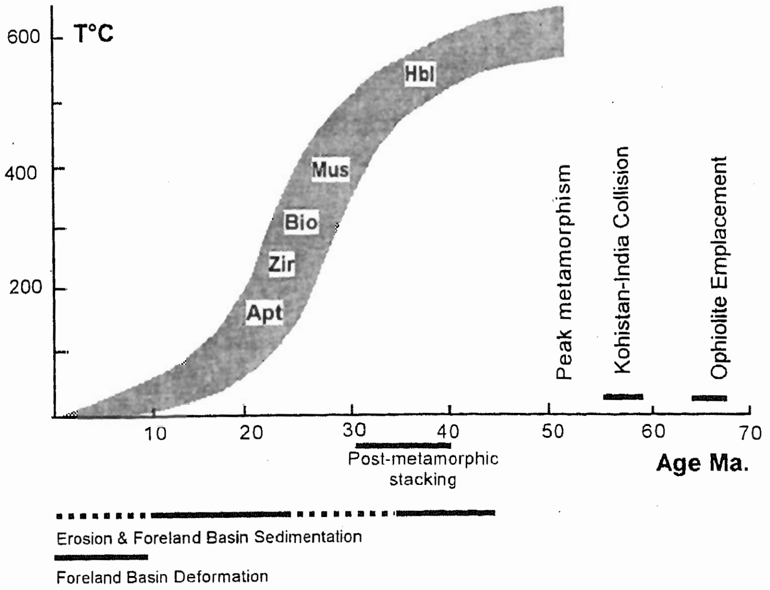


Fig. 7.16. Temperature-time relation showing the cooling history of the internal zone of the NW Himalaya in Pakistan. Apatite and zircon ages are based on fission track (Zeitler 1985), muscovite and biotite based on K-Ar, and hornblende based on Ar-Ar data (modified from Treloar et al. 1991c). Bars at the base show the relationship between cooling (which was at its maximum between muscovite and apatite blocking temperatures), erosion, sedimentation and deformation.

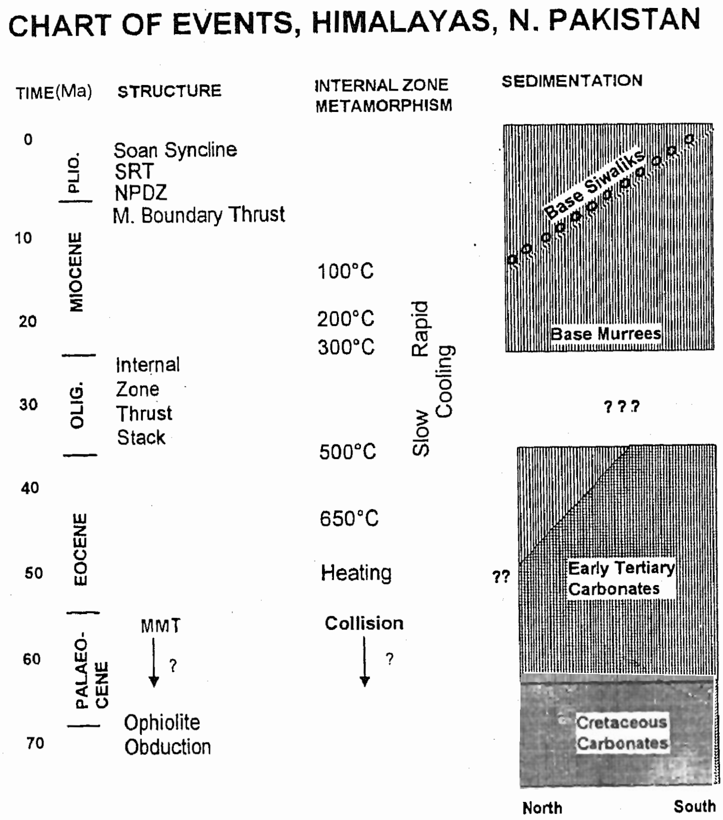


Fig. 7.17. Time chart of events within the Himalaya of Pakistan. MMT: Main Mantle Thrust or Indus suture; NPDZ: North Potwar Deformation Zone; SRT: Salt Range Thrust. (Modified from Treloar et al. 1991c).

only 100° C and just a few kilometres below the topographic surface (Treloar 1995a, Treloar et al. 1991).

The rapid exhumation was associated with extension within the upper part of the Indian plate, the suture zone and the Kohistan arc, and was due to unconstrained tectonic denudation and erosion (Treloar et al. 1991, Chamberlain et al. 1991). It was synchronous with, and probably initiated by, south-verging crustal scale thrust stacks the growth of which post-dated Himalayan ductile deformation and Barrovian-type metamorphism (Coward et al. 1988). From petrographic data, Treloar et al. (1991) concluded that the shearing could have started early into the post-metamorphic history, about or soon after cooling through the hornblende blocking temperature (Chamberlain et al., 1991, regard that imbrication started before metamorphism). That stacking was most probably completed by 30 Ma is supported by: (1) 20 Ma zircon fission-track ages across much of the high-grade internal zone rocks, and (2) indistinguishable muscovite ages (23-30 Ma) from the shear zone and those in the unsheared rocks. But uplift/exhumation may also have been triggered by a change from subduction to strike slip motion in the MMT to allow the rise of a relatively buoyant crust (DiPietro 1991), or caused by isostatic response to crustal thickening. Treloar et al. (1991) have suggested that the entire rock package in the internal zone was uplifted bodily along the Panjal–Khairabad thrust during the Miocene. Some uplift also occurred subsequently along the shallow dipping thrusts occurring further south, e.g., Main Boundary Thrust.

The phenomenon of Miocene extension has been recently investigated; it comprises both normal faulting and collapse folding (Burg et al. 1996, Vince and Treloar 1996). It was most likely effected by crustal or lithospheric delamination beneath thickened crust rather than thin-skinned tectonics. Estimates of the absolute vertical exhumation are not clearly known, but it may have exceeded 20 km during Oligocene-Early Miocene (Treloar et al. 1991). The stripping of the overburden through denudation is recorded in the Eocene, Miocene and Pliocene molasse deposits in the foreland basins (Fig.7.17).

Deformation and Metamorphism

The Indian plate rocks in northern Pakistan and Azad Kashmir have passed through at least three phases of deformation (Kazmi and Rana 1982, Chaudhry and Ghazanfar 1987, 1992, Williams 1989, Treloar 1989, Baig 1990, DiPietro 1991, Ghazanfar 1993). The first of these, with possible synchronous metamorphism, may have accompanied the ophiolite emplacement about the Cretaceous-Paleocene boundary. But there is a difference of opinion regarding the relationship between deformation, metamorphism and timing of the Kohistan-India collision. DiPietro and co-workers and Chamberlain et al. (1991) think that peak metamorphism occurred before this collision. The views of DiPietro et al. (in prep.) are summarised in Fig.7.18. Treloar and co-workers think that peak metamorphism was consequent upon this collision. In the following we present a summary of the latter model (i.e., Treloar et al. 1991, Treloar 1995a).

The deformation accompanying the southward thrusting of Kohistan onto India was characterised by ductile simple shear (S_1) and incorporated the D_1 and D_2 events of Coward et al. (1982, 1988). Textural studies, including spectacularly rotated garnet porphyroblasts, show that metamorphism was synchronous with this shear. Treloar et al. (1989a, b), Treloar (in press) and Khattak and Stakes (1994) showed that garnet cores formed at lower pressure and temperature than garnet margins. Such a path of increasing pressure

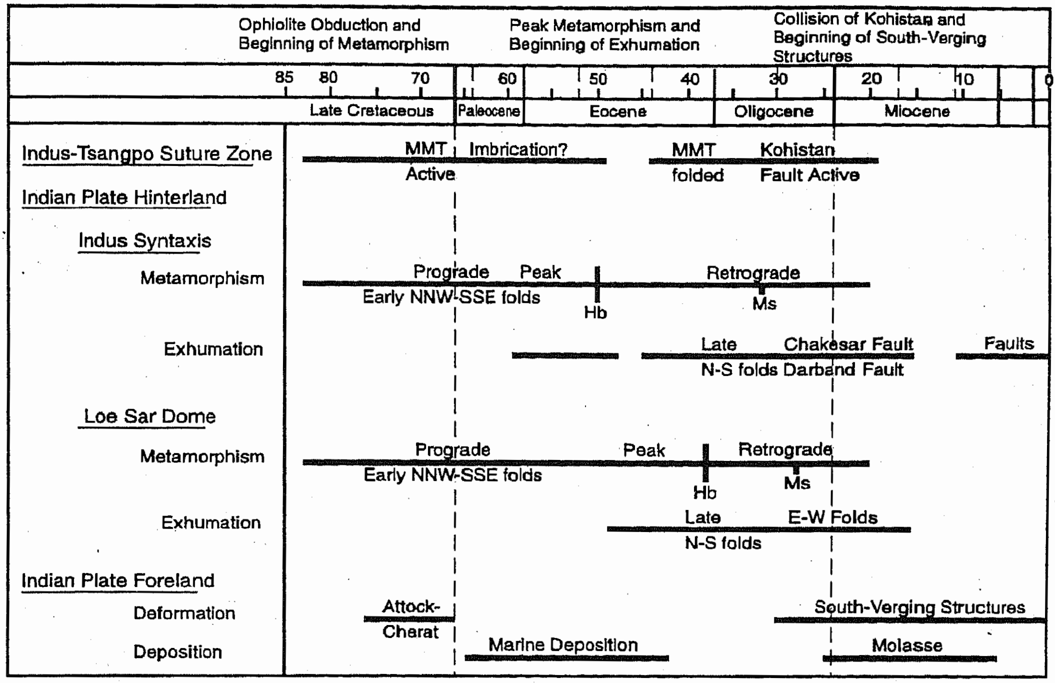


Fig. 7.18. Sequence of Himalayan events based on Ar cooling ages of hornblende and muscovite, and on thge timing of metamorphism relative to development of folds and faults. F$_3$ folds are associated with S$_3$ foliation. (After DiPietro et al. in preparation).

would be consistent with metamorphism occurring during the active subduction of the leading edge of the Indian plate. Chamberlain et al. (1989b, 1991) suggested that this was in response to overthrusting of either the Kohistan arc or earlier thrust imbricates onto India.

Metamorphic conditions in the inner zone sequence have been studied by many workers (for summary, see Treloar, in press). Estimates of pressure and temperature in the kyanite zone in different blocks are displayed in Fig. 7.19. The highest grade rocks in Swat, Hazara, Kaghan and Nanga Parbat areas crystallised at 600–750° C and 7–12 kbar (compare Broughton and Windley, 1988). In upper Kaghan, even higher pressures (15 kbar, 650° C) prevailed locally to produce eclogites from mafic precursors. The kyanite-sillimanite zone conditions were suitable for partial melting of the crustal rocks to generate granitic melts. We have described in the previous chapter the occurrence of Early Tertiary granites in the area.

The second major phase of deformation within the internal zone can be related to the re-imbrication of the metamorphic pile within a south-verging, crustal-scale thrust stacks (the D$_2$ event of Treloar et al. 1989a; see also Williams 1989, Chamberlain et al. 1991). The stack consists of nappes involving a partial re-imbrication of the basement and cover sequences decoupled earlier in the deformation (Fig. 7.21). The nappes are internally imbricated by ductile to brittle shear zones, within which the main phase metamorphic fabrics are extremely reworked (Treloar 1995). Data are lacking on metamorphic overprinting associated with this stacking. Treloar et al. (1991) have noted that in the Hazara nappe, Ca-rich rims on garnet of the staurolite zone formed at about 8 kbar. The stacking

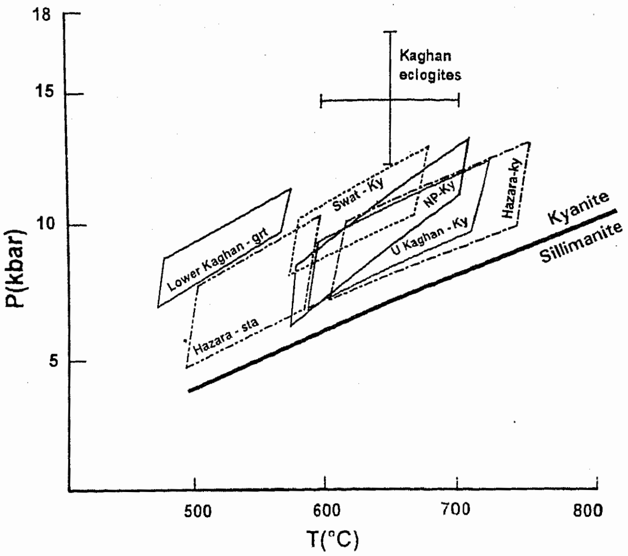


Fig. 7.19. Pressure-temperature conditions of metamorphism for selected rocks within various nappes within the Indian plate of NW Pakistan. From Treloar et al. (1989b), Treloar (1995a), with Kaghan eclogite estimates from Pognante and Spencer (1991) and Nanga Parbat (NP) data from Pognante et al. (1993).

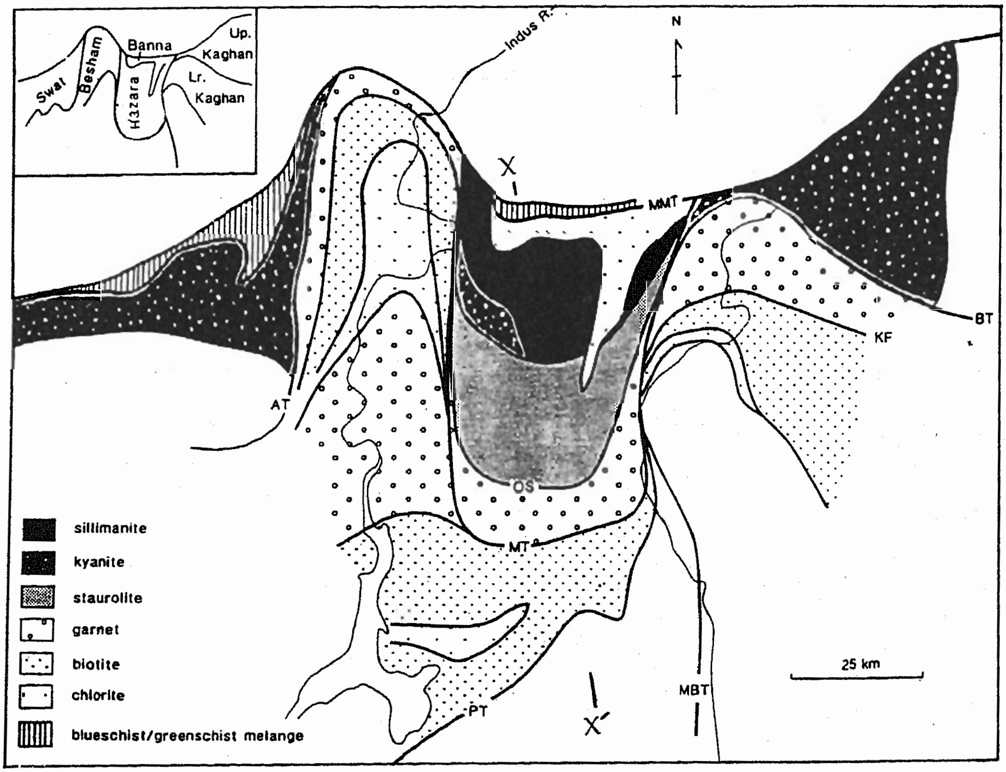


Fig. 7.20. Metamorphic map of the Swat to Kaghan section of the Indian Plate within north Pakistan. BT: Batal Thrust; KF: Khanian Fault; AT: Alpurai Thrust; MBT: Main Boundary Thrust; OS: Oghi Shear; MT: Mansehra Thrust; TP: Panjal Thrust; X-X': line of section in Figure 7.21 (from Treloar 1989).

of the metamorphic sequence during this phase was such that, within each nappe, metamorphic grade increases northwards and structurally upwards with sharp metamorphic breaks across the shears separating individual nappes (Fig. 7.20). This metamorphic inversion is not related entirely to an originally inverted thermal gradient associated with emplacement of a hot thrust slab over a colder one, magmatic intrusion at higher structural level, recumbent folding, and so on (Windley 1983, Verma 1989, DiPietro and Lawrence 1991). It resulted from stacking of previously metamorphosed and partially cooled rocks in their hanging walls than in their footwalls (Fig. 7.21). Despite the apparent upward increase in metamorphic grade, rocks in the immediate footwall of the MMT are of a lower grade (frequently greenschist facies) than those which underlie them structurally. This decrease in metamorphic grade within the uppermost part of the Indian plate and the suture zone melanges is not a primary metamorphic feature but a consequence of retrograde metamorphism (Khan 1992) related to Early Miocene extension concentrated about the MMT (Vince and Treloar 1996). The curved outcrop traces of the stack (Fig. 7.20) were produced during later folding that generated large-scale W-trending antiformal and synformal features.

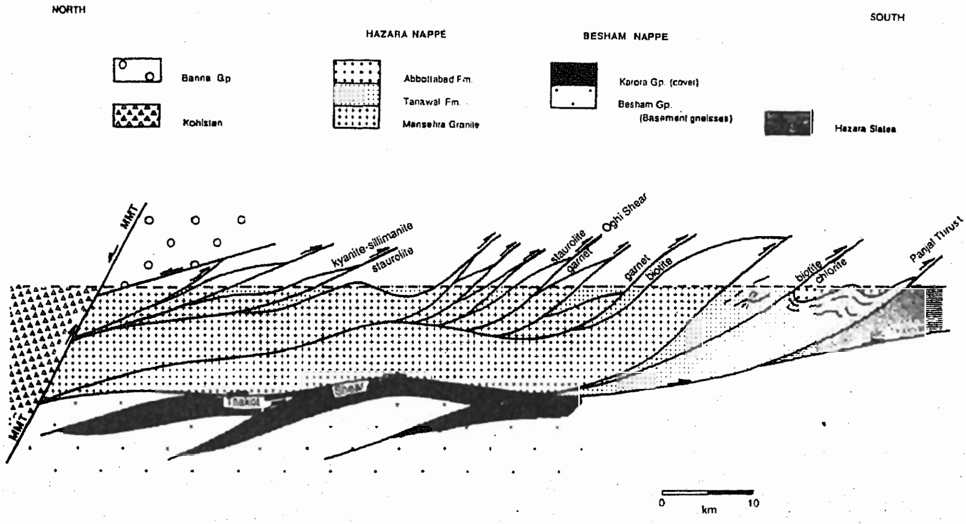


Fig. 7.21. Structural section along line X-X' in Figure 7.20, showing the tectonic inversion of metamorphism within the Hazara nappe by thrusts which post-date the metamorphic peak (from Treloar 1989, 1995a).

Mechanism of Metamorphism

The thermal evolution of orogenically thickened crust in the context of conductive heat transfer was outlined by England and Thompson (1984). This work suggests that in the absence of magmatic heat input, prograde regional metamorphism would require tens of millions of years. In the previous pages, we reported that deformation, heating, metamorphism, and subsequent initiation of cooling within the Indian plate occurred within 5–15 Ma of its collision with, and subduction beneath, the Kohistan arc. Thus, metamorphic time scale is short, and long time conductive models for the thermal evolution of the Indian plate in northwest Himalayas may be inappropriate (Treloar et al. 1989a, 1991, Treloar 1995b).

The time interval, according to Treloar and Rex (1990a) and Chamberlain et al. (1991), is insufficient to permit the attainment of middle to high amphibolite facies temperatures by the purely conductive heating of cold rocks. This could have been possible if the rocks had extremely high internal heat generating properties, for which there is no evidence. Treloar et al. (1989b) have argued that the Indian plate cover sequence was decoupled from the basement during the obduction of the Kohistan arc, with the potentially high heat producing upper cover sediments tectonically eroded and bulldozed southward in front of the advancing arc. Thus, an extremely attenuated and thinned cover sequence (on top of the Proterozoic basement) was subducted beneath Kohistan. It is worth noting that (1) the basement rocks do not overlie younger rocks in the thrust stacks, and (2) with the exception of the Nanga Parbat Massif, basement rocks do not yield post-Proterozoic Ar-Ar ages. This implies that either Early Tertiary metamorphism was restricted to the cover sequence or the metamorphosed cover sequence was transported a great distance during the post-metamorphic thrust stacking. In any case, there was insufficient quantity of high-heat producing upper crustal rocks.

To overcome these problems, Chamberlain et al. (1991) explored the possibility that Early Tertiary (ca. 50 Ma) metamorphism may have been related to imbrication of sedimentary sequences within an accretionary wedge, and emplacement of the thrust stacks onto India prior to collision with Kohistan. This possibility was discounted by Treloar (in press) on the grounds that most accretionary wedges are built on the subduction zone hanging walls from syntectonic sediments. Another possibility worth evaluation is that the main phase metamorphism predated the Tertiary collision of India with Asia. Although earlier on suggested by Baig (1990) and recently by DiPietro et al. (in prep.), this idea is hard to substantiate because (1) it would require considerably greater shortening of the Indian plate than has so far been recognised, and (2) there is no record of a strong thermo-tectonic event between the Cambrian and Tertiary.

In the absence of magmatic heat or very rapidly generated internal heat, the driving force for the heating and metamorphism soon after crustal thickening must have involved more than simple conductive heating. Allowance should be made for frictional heating along the MMT and internal heating within the upper part of the Indian plate. Details of this model, describing rapid regional metamorphism and equally rapid post-metamorphic cooling, have been given by Treloar (1995 and in press). According to him, dissipative shear heating along MMT contributed to early stages of heating of the footwall rocks. But the temperature estimates for the highest grade rocks are too high to be solely produced by this mechanism, so there probably were additional factors also. Dissipative shear heating along MMT generated an early inverted thermal profile in the upper units of the Indian plate. The Kohistan rocks in the hanging wall of the MMT are mafic and, thus, have low thermal conductivity. They would, therefore, have acted as a thermal reflector. As the temperatures in the footwall increased through the brittle-ductile transformation, the role of dissipative shear heating decreased and continued heating became a function of internal heat generation within the footwall rocks, together with the hanging wall thermal reflectivity. The metamorphic inversion was re-inforced by imbrication of the metamorphic stack as it accreted onto the footwall during the early stages of uplift exhumation.

8

Neotectonics

Pakistan is characterised by extensive zones of high seismicity and contains several seismotectonic features generated by an integrated network of active faults. The earliest indication of an active fault associated with earthquakes came in 1892, when the town of Chaman was destroyed and the great Chaman Fault was noted for the first time (Griesbach 1893). The Makran earthquake of 1945 created a number of small off shore islands, arranged along a line parallel to the coast and indicated a submarine active fault (Sondhi 1947). Photogeological reconnaissance of Balochistan (HSC 1960) revealed eight more active faults, including the Ornach-Nal Transform Fault. Later studies, amongst others by Wellman (1966), Abdel-Gawad (1971), Nowroozi (1972), Menke and Jacob (1976), Kazmi (1974, 1977, 1979a, 1979b), Seeber et al. (1979, 1980), Quittmeyer et al. (1979), Armbruster et al. (1980), Verma et al. (1980), Lawrence and Ghouri (1983), Nakata and Tsutsumi (1989), Yeats and Hussain (1989), McDougall and Khan (1990) and Treloar et al. (1991c,d) have brought into sharp focus the seismicity, seismotectonics and the active fault system in Pakistan. These features are briefly reviewed in this Chapter.

Seismicity

The documented record of seismicity consists of mainly earthquake epicenters located on the basis of modern instrumental recordings (Fig. 8.1). This seismic activity is the result of movement along various active faults in the region. Thus it may be seen that the collisional mountain ranges, where active faults are common, are endowed with high seismicity whereas the more stable Indus Platform zone is characterised by relatively low* seismicity. In Pakistan (Fig. 8.1), the following zones of high seismicity may be identified.

Hindu Kush – West Karakoram Seismic Zone

This zone comprises the Tirich Mir region and the adjacent Afghan territory. It is seismically hyperactive with frequent earthquakes of magnitude 5–7 and focal depths of 100-200 km (Kazmi 1979b). Focal mechanism solutions and spatial distribution of earthquakes and their hypocenters indicate the presence of a fault plane down to about 200 km depth and a 25 km thick contorted Benioff zone within the upper mantle. According to Billington et al. (1977), these features are typical of regions of active lithospheric subduction. This zone coincides with the Akbaytal Fault of Boulin (1990), which is the eastward extension of Herat Fault or the Waser-Rushan Pshart Suture of Sengor et al. (1988). Focal-mechanism solutions by Verma et al. (1980) indicate that thrust and strike-slip movements are equally prevalent in the Hindukush and in addition to the

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*Here we have followed Quittmeyer et al. (1979) in the use of the term "low" and "high" in a relative sense, to give a qualitative indication of the relative number of earthquakes per unit area per unit time in a given locality.

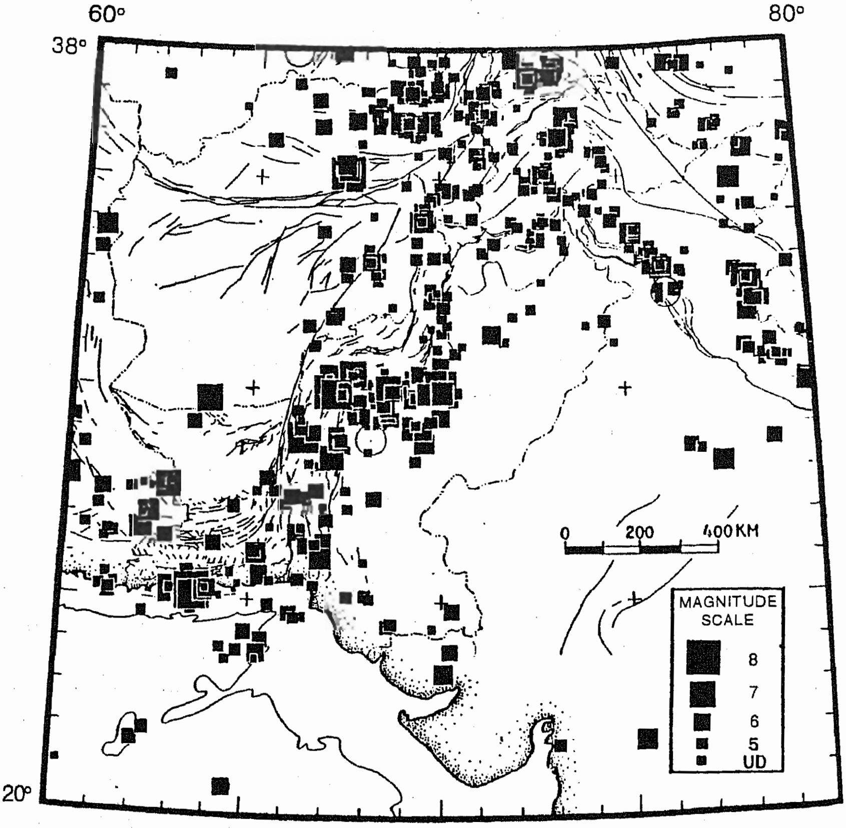


Fig. 8.1. Teleseismic map of Pakistan and adjacent regions. (From Seeber et al. 1980).

NE lineament (Akbaytal Fault) characterised by strike-slip movement, they mention a NW-lineament also which is dominated by thrust faulting (Fig. 8.2).

Yasin Seismic Zone

The region around the town of Yasin is seismically active and there is a cluster of several earthquake epicenters of magnitude 3-5 and focal depth of less than 50 km (Fig. 8.3). The Main Karakoram Thrust passes through this zone and it may be the source for this seismic activity.

Hamran Seismic Zone

South of Gupis, along the Hamran Valley which is in the center of the Ghizar Range, there is a particularly active seismic zone with frequent earthquakes of 3 to 5 magnitude and with hypocenters ranging from < 50 to 100 km depth (Fig. 8.3). In recent years these earthquakes have caused much damage in Gupis and adjacent areas. Though no distinct active fault has yet been mapped in this region, focal mechanisms for some events from this zone (Chandra 1978) show thrusting on northwest trending nodal planes. Seismic data collected by the International Karakoram Project shows that in this region seismic deformation is confined to the upper 20 km of the crust both during important earthquakes and during background activity (Yielding et al. 1984).

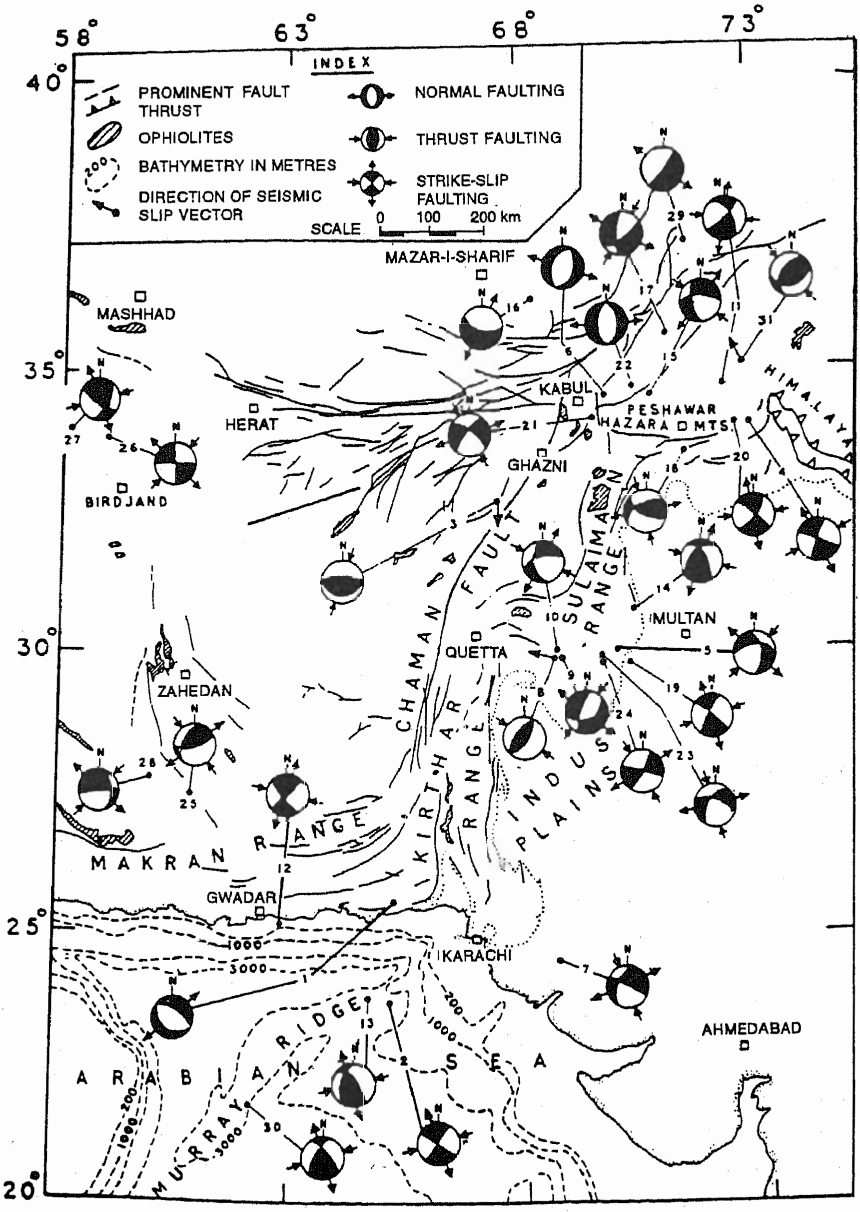


Fig. 8.2. Map showing major faults and focal mechanism of a few selected events in Pakistan and adjacent regions. Dark area comprises zone of compression, blank area is the zone of dilatation. (From Verma et al. 1980).

Indus-Kohistan Seismic Zone (IKSZ)

In northern Pakistan, a telemetered seismic network operated from 1973 to 1977, has recorded data from approximately 10,000 earthquakes, covering the area between longitudes 69° and 75° and latitudes 30° 30' and 35° 30' (Seeber and Armbruster 1979). This

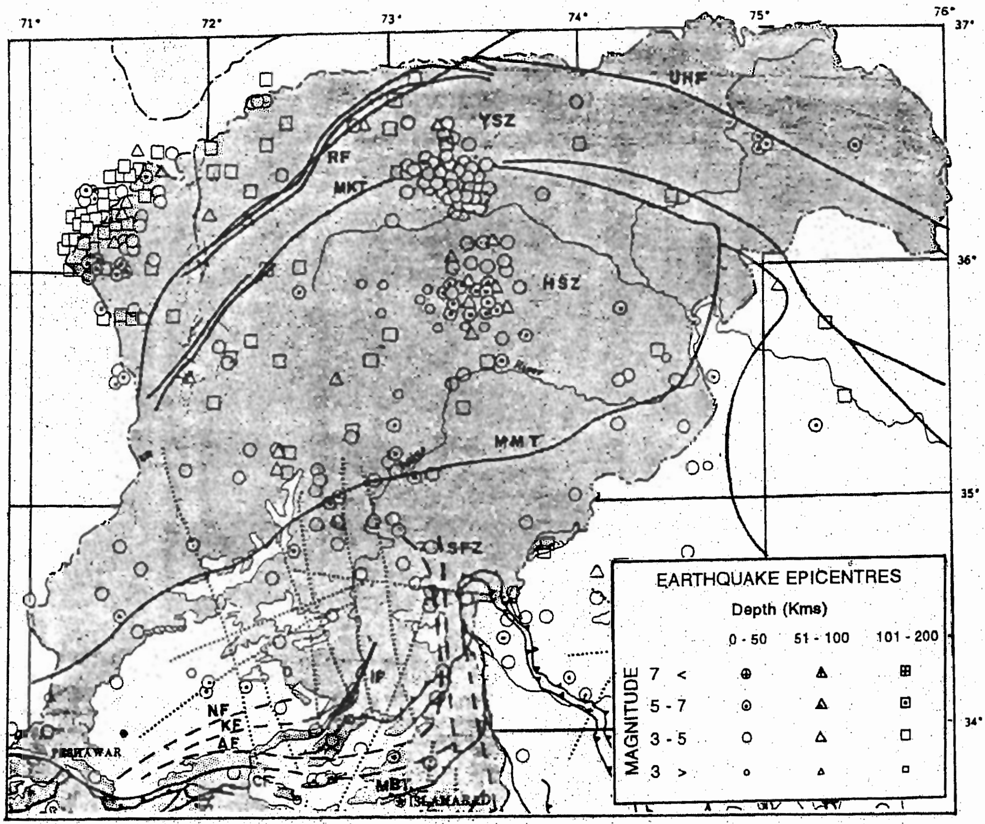
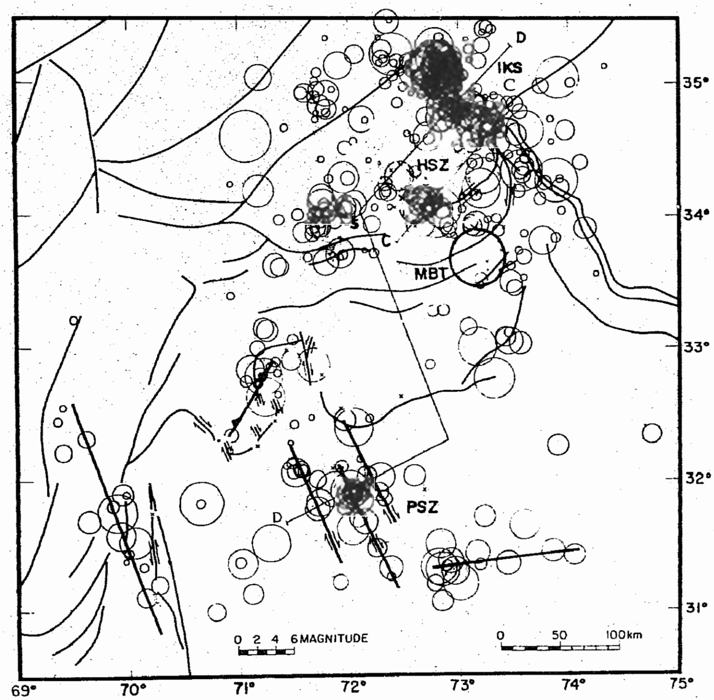
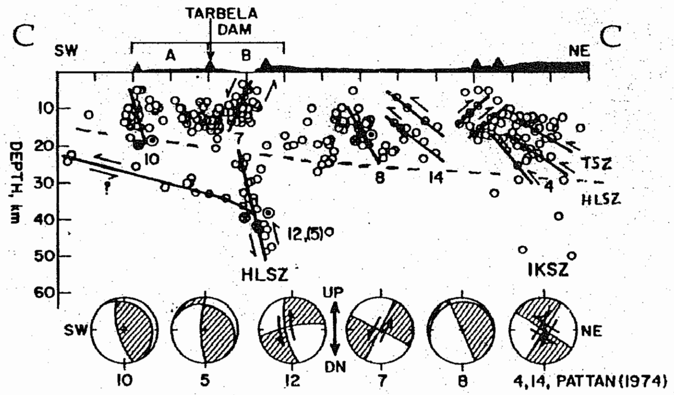


Fig. 8.3. Seismotectonic map of northern areas. Note the location of Yasin (YSZ) and Hamran (HSZ) Seismic Zones. Faults UHF=Upper Hunza, RF=Reshun, MKT=Main Karakoram Thrust, MMT=Main Mantle Thrust, NF=Naushera, KF=Kund, AF=Attock, IF=Indus, MBT=Main Boundary Thrust. (From Kazmi 1979).

study shows that the northwestern margin of the Indian plate north of the MBT, comprises a zone of high seismicity whereas the Salt Range and the Trans-Indus Ranges reflect moderate seismicity. In the study area, three main zones of high seismicity have been identified. North to south these are (1) the Indus-Kohistan seismic zone, (2) the Hazara lower seismic zone and (3) the Punjab seismic zone (Fig. 8.4) The Indus-Kohistan seismic zone has a NW-SE orientation and lies between the MMT and the Hazara-Kashmir Syntaxis. It overlaps the crystalline nappes of Swat, Besham and Hazara and parts of the Kohistan magmatic arc along the MMT. Tectonic activity in this region generated the 1974 Pattan earthquake (Jackson and Yielding 1983). According to Seeber et al (1979) in this region most of the seismicity is associated with a 10 to 12 km deep, wedge shaped feature, which has a horizontally disposed upper surface and a northeasterly dipping lower surface (Fig. 8.4). Seismic data shows that the upper surface conforms to a widespread seismic zone interpreted as a detachment zone along which the decollement of sediments and metasediments is decoupled from the basement. Below the decoupling surface, fault plane solutions indicate northeast and southwest dipping thrust faults (Chandra 1978, Seeber and Armbruster 1979, Pennington 1979). This structure is consis-





C–C. Vertical projection of the seismicity along a NE-trending strip across the Tarbela seismic network. The hypocenters cluster along well-defined features. Faults based on hypocenter alignments. Fault-plane solutions from Armbruster et al. 1978. HLSZ–Hazara Lower Seismic Zone, TSZ–Tarbela Seismic Zone. **D–D.** Section showing the structure beneath the seismic network deduced from the seismicity. UT, MT, LT–upper, middle, lower Tertiary; MZ–Mesozoic; PZ–Paleozoic; PC–Precambrian; TG–Tertiary granite.

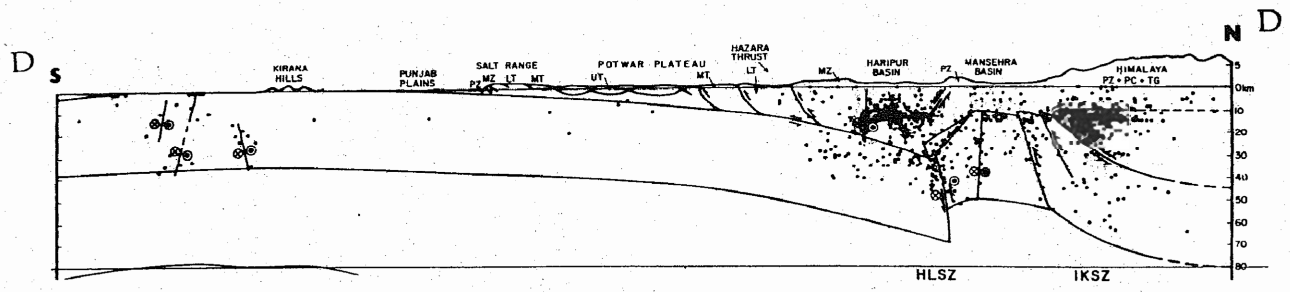


Fig. 8.4. Map showing six months of epicentral data from Tarbela (February–August 1976) and Chasma (May–November 1976) networks. Only earthquakes of magnitude 0.5 or greater shown. Thin lines show main structural features; faults shown by thick lines. HSZ–Hazara Lower Seismic Zone, IKS–Indus Kohistan Seismic Zone. MBT–Main Boundary Thrust, PSZ–Punjab Seismic Zone.(From Sebeer and Armbuster 1979).

tent with the crystalline thrust-stacking model of Treloar et al. (1991 c) for the crystalline nappes of Swat, Besham and Hazara (Fig. 8.4). According to Treloar et al (1991d) if this SW-verging, NE-dipping blind thrust detaches in mid to lower crust, it should structurally underlie both the Hazara Syntaxis and the Besham antiform.

Tarbela and Hazara Lower Seismic Zones

This zone is located about 60 km southeast of IKSZ and runs parallel to it (Fig. 8.4). The nearly horizontal aseismic detachment zone of IKSZ may be traced into this zone also. Here the detachment zone forms a kind of divide, with a distinct overlying zone of shallow seismicity and a deeper seismic zone beneath it (Fig. 8.4). The shallow seismic zone is known as the Tarbela seismic zone (TSZ) and comprises a set of steeply-dipping NW and NE striking faults which are seismically very active at 8 to 18 km depth (Seeber and Armbruster 1979, Seeber et al. 1980). Underlying the TSZ there is the deeper Hazara lower seismic zone (HLSZ) in which the hypocenters cluster along a steeply dipping basement fault with reverse and right-lateral strike-slip motion.

The aseismic detachment zone or the decoupling surface between the sedimentary wedge and the basement is primarily a sub-horizontal boundary to seismic zones and extends from IKSZ to south of the Salt Range. In the Salt Range and the Potwar it is associated with Infracambrian salt. North of the Potwar Plateau continuation of the salt is rather speculative (Chapter 4).

Punjab Seismic Zone

In the northern part of the Indus Platform, the high-gravity zone corresponding to the buried Sargodha-Shahpur ridge comprises a region of relatively high seismicity (Fig. 8.4). Fault-plane solution of earthquakes from this region indicate extensional faulting in the upper crust due to downward bending of the crust (Molnar et al. 1973). Strike-slip active faults, transverse to the Himalayas, occur along the Ganga Platform in India (Valdiya 1976) as shown in Figure 8.5 and similar structures are inferred in the Punjab seismic zone (Fig. 8.4). A large number of very extensive lineaments, oriented longitudinally or transversely (NNW to NE) towards the mountain front, have been mapped throughout the Indus Platform zone (Kazmi 1979a, 1979b, Kazmi and Rana 1982) and are likely to be associated with deep crustal faults like the ones revealed by seismic data in the Ganga Platform and the Punjab seismic zone and by seismic surveys in the southern part of the Indus Platform (Chapter 4).

Sulaiman Seismic Zone

This zone covers the entire Sulaiman fold-and-thrust belt and is characterised by shallow earthquakes of moderate to high magnitude (5–7). Within this zone there is the Harnai–Kohlu high seismicity belt, which is arcuate and 25 to 50 km wide. It follows the general EW orographic and structural trend and lies between the Gumbaz Valley to the north and the Kaha Valley to the south (Fig. 8.6). From the town of Khost in the west, this belt extends 250 km eastward up to Rakhni and it is bounded by the Khalifat Fault in the north, the Tatra Fault in the south and the Kingri Fault in the east. This high seismicity belt contains a number of large NE, EW and NW trending faults. According to Quittmeyer et al. (1979) the earthquakes in this belt may be the result of movement on single, continuous fault or else they may be due to activity on a number of smaller but similarly oriented faults. Kazmi (1979a, 1979b) has identified some of these faults (e.g., the Khalifat, Kohlu, Tatra and Kingri Faults) and considers them to be active owing to the concentration of earthquake epicenters along the fault traces. Focal-mechanism solutions of a number of events in the Harnai–Kohlu belt suggest the predominance of strike-slip faulting (Fig. 8.2) though thrust faulting is also indicated (Molnar et al. 1873, Chandra 1978). The devastating Sharigh earthquake of 1931 having a magnitude of over 7 caused much damage in this belt. Another severe earthquake, the 1966 Duki earthquake, occurred in this region (Kazmi 1972, Quittmeyer et al. 1979).

The Sulaiman Range comprises another high seismicity belt characterised by a cluster of recent seismic events along two narrow, N-S trending, over 180 km long bands along the eastern and western margins of the main Sulaiman Range (Figs. 8.4 and 8.6). These high seismicity bands coincide with left lateral active faults and lineaments (Gawad 1971, Kazmi 1979a, 1979b). Fault plane solutions on a few earthquakes from the Sulaiman Range indicate strike slip faulting (Fig. 8.2).

North of the Harnai–Sharigh seismic belt there is another narrow, arcuate zone of high seismicity centered along the Loralai–Mekhtar Valley. It is characterised by shallow (<50 km), 3–6 magnitude events attributed to the active Mekhtar Fault (Kazmi 1979a). In other areas of the Sulaiman seismic zone, the seismicity is rather diffused as suggested by the wide scatter of the seismic events (Fig. 8.6).

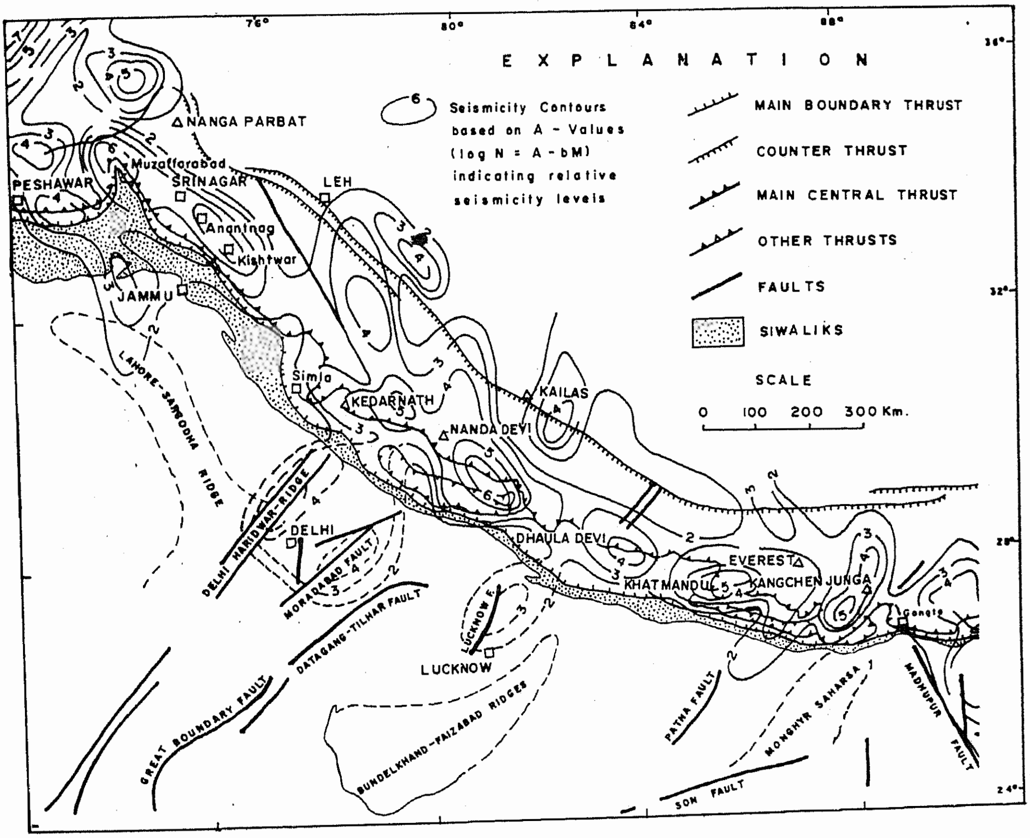


Fig. 8.5. Sketch map showing the seismicity in northern Pakistan and India and the some of the deep crustal faults in the Ganga Platform (from Kaila and Narain 1979).

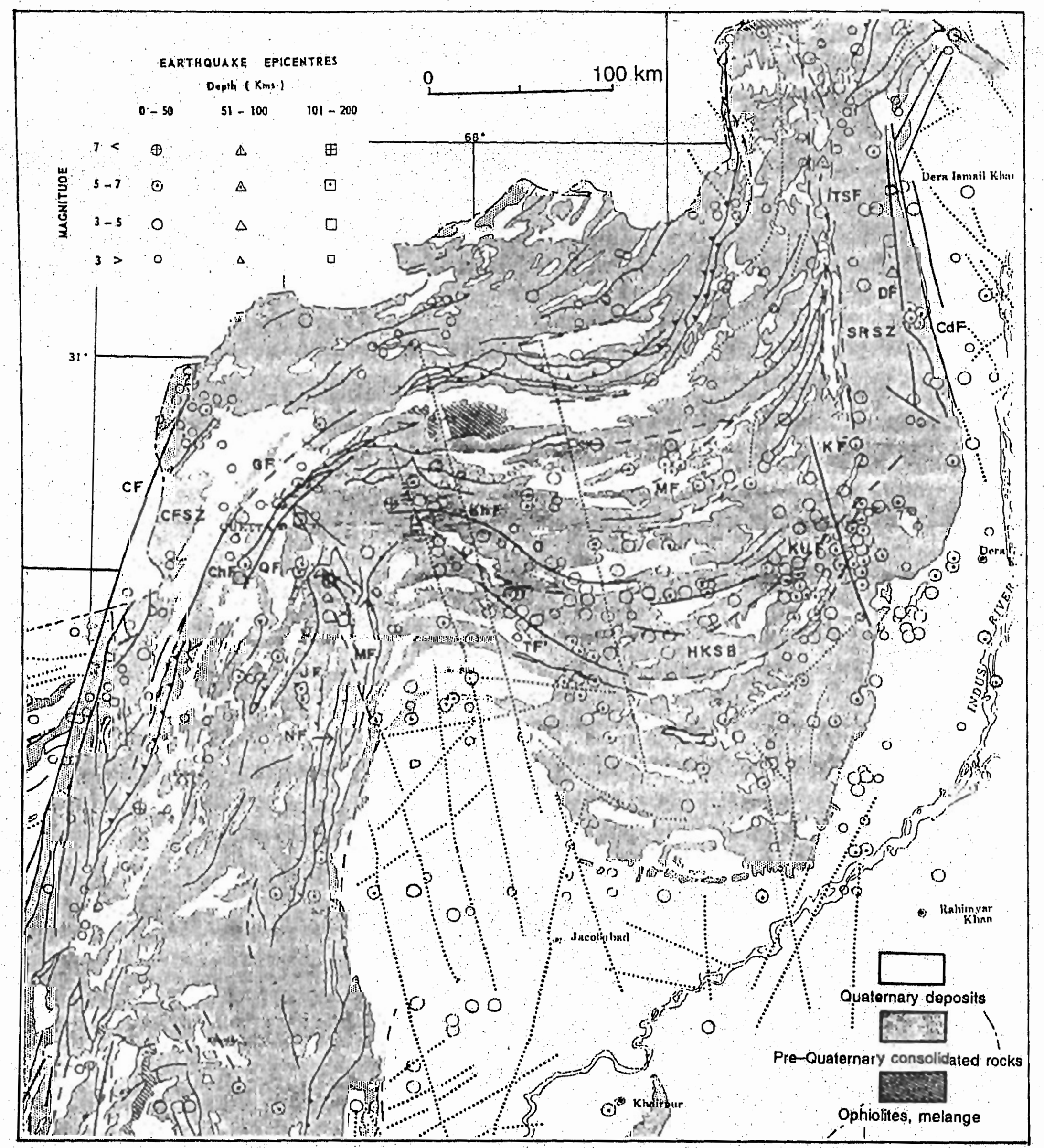
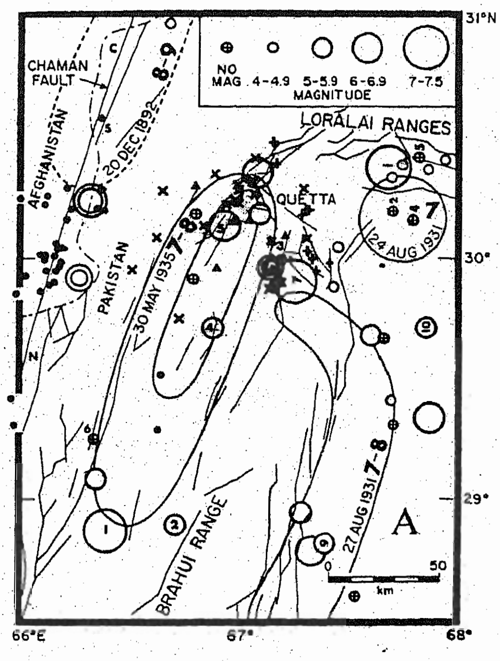


Fig. 8.6. Seismotectonic map of the Sulaiman and Northern Kirthar seismic zones. CF–Chaman Fault, CFSZ–Chaman Fault seismic zone, ChF–Chiltan Fault, GF–Ghazaband Fault, HKSB–Harnai-Kohlu seismic belt, JF–Johan Fault, KF–Kingri Fault, KhF–Khalifat Fault, KuF–Kohlu Fault, MF–Mach Fault, NF–Nagau Fault, TF–Tatra Fault. (From Kazmi 1979).

North Kirthar Seismic Zone

This zone comprises the northern portion of the Kirthar fold-and-thrust belt lying between Quetta and Kalat and bounded to the east by the Kachhi Plain and to the west by



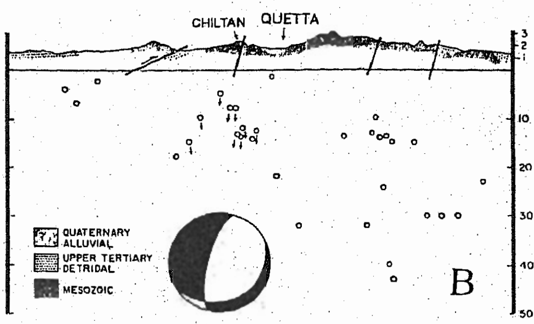


Fig. 8.7.B. Generalised geological cross-section through the Quetta area. Small circles indicate hypocenters determined by the Quetta seismic network. Focal mechanism for the fault responsible for 1935 earthquake is also shown. Shallow dipping fault to the west (Ghazaband Fault) is also active (From Armbruster et al. 1980).

Fig. 8.7.A. Map showing earthquake epicenters in the Quetta area from 1914 to 1977(open cicles) and modifiedMercalli isoseismals for the largest three earthquakes. Sequence of events indicated by number: 1935 Quetta = numbers upright,1931 Mach = numbers facing right, Triangles–Quetta network stations; X = hypocentral depth 0–15 km, + =15–30 km, filled circles = depth fixed at 15 km; double circle = epicenters 1975–1978. Note increase of seismicity on southern section of Chaman Fault since 1975. (From Armbruster et al. 1980).

the Ghazaband Fault. The latter separates it from the Chaman Fault siesmic zone (Fig. 8.6). The high seismicity of this region is indicated by several teleseismic events from 1914 to 1975 and also by 63 small earthquakes recorded between October and November 1978 by the Quetta seismic network (Armbruster et al. 1980). Two major earthquakes of magnitudes greater than 7 devastated this region (1931 Mach and 1935 Quetta earthquakes). Lesser magnitude events occurred in later years e.g., 1955 Quetta earthquake of magnitude 6.2 and 1977 Kolpur earthquake of magnitude 5.5. Seismicity in this zone is rooted at shallow levels as several hypocenters are within 5–15 km depth in Quetta–Mastung region. It deepens eastward to depths of 30–40 km (Fig. 8.7).

Seismic events and associated ground-ruptures at various locations indicate that the seismicity of this region is centered around a few N-S trending faults (Ghazaband, Quetta, Chiltan, Johan, Mach and Nagau Faults). The 1935 Quetta earthquake formed a 150 km long NNE oriented rupture (Quetta Fault of Armbruster et al. 1980), parallel to the Chiltan Fault. Since the 1935 main event, this ruptured zone is quiescent and recent seismicity is now centered at the extremities of this zone, in the vicinity of Quetta and Mastung. The fault plane solution for the 1935 Quetta earthquake shows thrust faulting along a plane having strike N 33° E, dip 68° NW and a focal depth of 20 km (Gupta and Singh 1980). Composite fault mechanism of recent events, however indicate left-lateral strike slip motion in the northern portion of the Quetta Fault. Fault plane solution for events related to other active faults in this region indicate strike-slip faulting, though the composite data are equally compatible with normal faulting (Armbruster et al. 1980).

Chaman Fault Seismic Zone

This zone forms a narrow seismic strip around the 900 km long Chaman Fault and has been the site of infrequent; moderate to large earthquakes, though the more recent events have small magnitudes. The short term data from the Quetta network also suggest that this a most active seismic zone (Figs. 8.6 and 8.7). Prior to 1900, two large earthquakes occurred in the vicinity of Chaman Town and produced surface ruptures (Oldham 1982, Griesbach 1893). Other significant events occurred in 1975 (Spin Tezha, M 6.4 and 6.7) and in 1978 (Nushki, M 5.9). Besides these three larger events, several small earthquakes have occurred in this 60 km long section of the Chaman Fault between Spin Tezha and Nushki since 1975 and this indicates an abrupt increase in the seismicity on the fault, south of the 1892 events (Armbruster et al. 1980). Focal mechanism for the 1975 event indicates left-lateral strike slip motion which is consistent with the geological field evidence (Quittmeyer et al. 1979, Lawrence and Yeats 1979).

Makran Coast Seismic Zone

This zone comprises a region of active subduction and is characterised by shallow seismicity. Most of the earthquakes in this region are located off the coast. Two offshore earthquakes occurred in this region in 1945 (M_8=8.0) and 1947 (M_8=7.3) having epicenters in approximately the same area (Fig. 8.1). The 1945 event severely damaged coastal towns, formed vertical displacements of 4.5 m on the shore near Pasni, produced upheaval of the sea floor in the form of a row of off-shore islands or mud volcanoes and generated a large tsunami (HSC 1960, Sondhi 1947, Gutenberg et al. 1954, Geller et al. 1977). Several high level coastal terraces and platforms provide geomorphological evidence of vertical uplift and high seismotectonic activity in this region.

Inland, a series of shallow 3–7 magnitude events have occurred along the arcuate Hoshab–Mashkai Valley and are attributed to the activity along Hoshab Fault zone (Kazmi 1979a). The earthquake hypocenters from Makran indicate a northward dipping seismic zone which is shallow (15–20 km) in the off-shore region and deepens inland to about 80 km (Fig. 4.44). Focal mechanism for two earthquakes in the Makran region suggest down-dip tension in a subducting slab (Nowroozi 1973).

Seismicity in other regions

In the remaining areas of Pakistan, namely Chagai, Bannu, Potwar, Bela-Khuzdar and southern Kirthar fold belt, the level of seismicity is relatively low and rather diffused. Surface faults and lineaments are mapped in these regions (Kazmi and Rana 1982) but the modern teleseismic activity does not align with any particular feature. The central and southern part of the Indus Platform zone is devoid of any modern teleseismic activity, except in the Sulaiman and Kirthar foredeeps and the Rann of Cutch and adjacent areas. In the Kirthar fore-deep, shallow low-magnitude teleseismic activity has been recorded. This seismicity may be related to bending of the lithosphere and active basement faults or development of a new frontal thrust (Quittmeyer et al. 1979). A number of extensive NNW trending lineaments have been mapped in this region and are thought to reflect some of the basement faults (Kazmi 1979a, 1979b). At least one large-magnitude event, the

1909 Kachhi earthquake, has occurred in this region and its epicentral tract had a northwesterly orientation.

The Rann of Cutch region is characterised by a low level seismic activity. However one large-magnitude earthquake (Allah bund earthquake) occurred in 1819 in this area and produced a vertical displacement of 7-9 metres at one location, forming a "bund"-like east-west oriented feature, which dammed the local drainage and formed a vast lake upstream (Oldham 1926).

Though apparently the present day aseismic feature of a large part of the Indus Platform zone is in keeping with the general seismic characteristic of most of the ancient shield areas of the world, it does not necessarily suggest that the Indus Platform comprises a truely aseismic zone free from earthquake hazards. Installation of modern sensitive seismic networks, like the ones operated for short periods in the Tarbela area, would be able to monitor the very low level seismicity of the Indus Platform zone and provide data for the location of buried active faults. Presence of active basement fractures, with large recurrence intervals for fault movement, may be expected as has been the case with the 1819 Allahbund earthquake or the 1909 Kachhi earthquake.

Active Faults

Active faults are discernible through offsets of topography, deformation of Late Quaternary deposits, offset of cultural features, seismicity and historic records. A number of active faults have been identified in Pakistan and according to Kazmi (1979a) they comprise four main types: (a) transform faults that mark the western boundary of the Indo-Pakistan subcontinent, (b) strike-slip faults which occur mainly along the margins of the structural arcs and oroclines and are to a large extent involved in the formation of various syntaxes, (c) thrust or reverse faults which are commonly arcuate faults and traverse the zone between each set of strike slip faults and (d) traces of basement geofractures which appear as lineaments of various dimensions (Fig. 8.8). A brief account of these features is given in the following section.

Karakoram Block

Topographic features, prominent lineaments noted on aerial photographs, and mapped fault traces, coupled with teleseismic activity along these features reveal a few active faults in this block. We have earlier mentioned the Akbaytal Fault associated with the high seismicity zone in Western Karakoram. East of this zone there is the **Reshun Thrust Fault** (Calkins et al. 1981) which extends from near Chitral up to Baroghil Pass towards the northeast (Chapter 4). A number of teleseismic events (magnitude 3-5, focal depth less than 50 km to more than 100 km) have been recorded along this fault zone in the upper Yarkhun Valley (the Dobargar-Lash region), and south-westward near Drasan, and west of Chitral (Figs. 8.1 and 8.9). There is therefore, a strong probability that this fault zone is an active feature.

In the Central Karakoram the E-W trending **Upper Hunza Fault** (Zanchi 1973, Kazmi 1979b, Gaetani et al. 1995) is an east-west oriented thrust fault that runs along the depression formed by the Chapursan and Upper Yarkhun Valleys. In this region it runs parallel to the northern margin of the Karakoram Axial Batholith. Southeast of Morkun (upper Hunza) three teleseismic events of magnitude 5-7 and focal depths of less than 50 km have been recorded. These epicenters form a cluster very close to the mapped trace of the

Upper Hunza Fault. It therefore appears that the northern and western margin of the Karakoram Axial Batholith is characterised by an active zone of faulting along the Upper Hunza-Reshun Fault zones.

The strike-slip **Karakoram Fault** forms the eastern termination of the Karakoram Block and comprises a nearly 500 km long fracture that appears vividly on satellite images as a spectacular linear feature. Right-lateral displacement of 200 to 250 km has been suggested along this fault (Burtman et al. 1963, Srimal 1983). A number of teleseismic events have occurred along the fault (Fig. 8.9). Interpretation of seismic data from 1975 Kinnaur earthquake confirms right-lateral strike slip motion on this fault (Ni and Barazangi 1985). During August-September 1980 a series of portable seismic stations were operated from a total of 24 sites in the region between the Nanga Parbat and Karakoram Ranges, by the International Karakoram Project (IKP) in order to record possible subcrustal earthquakes and to monitor crustal activity (Yielding et al 1984). During this short period three events were recorded from the Karakoram Fault Zone at the head of the Nubra Valley at approximately 35° 3' N: 77° 5' E (Fig. 8.9).

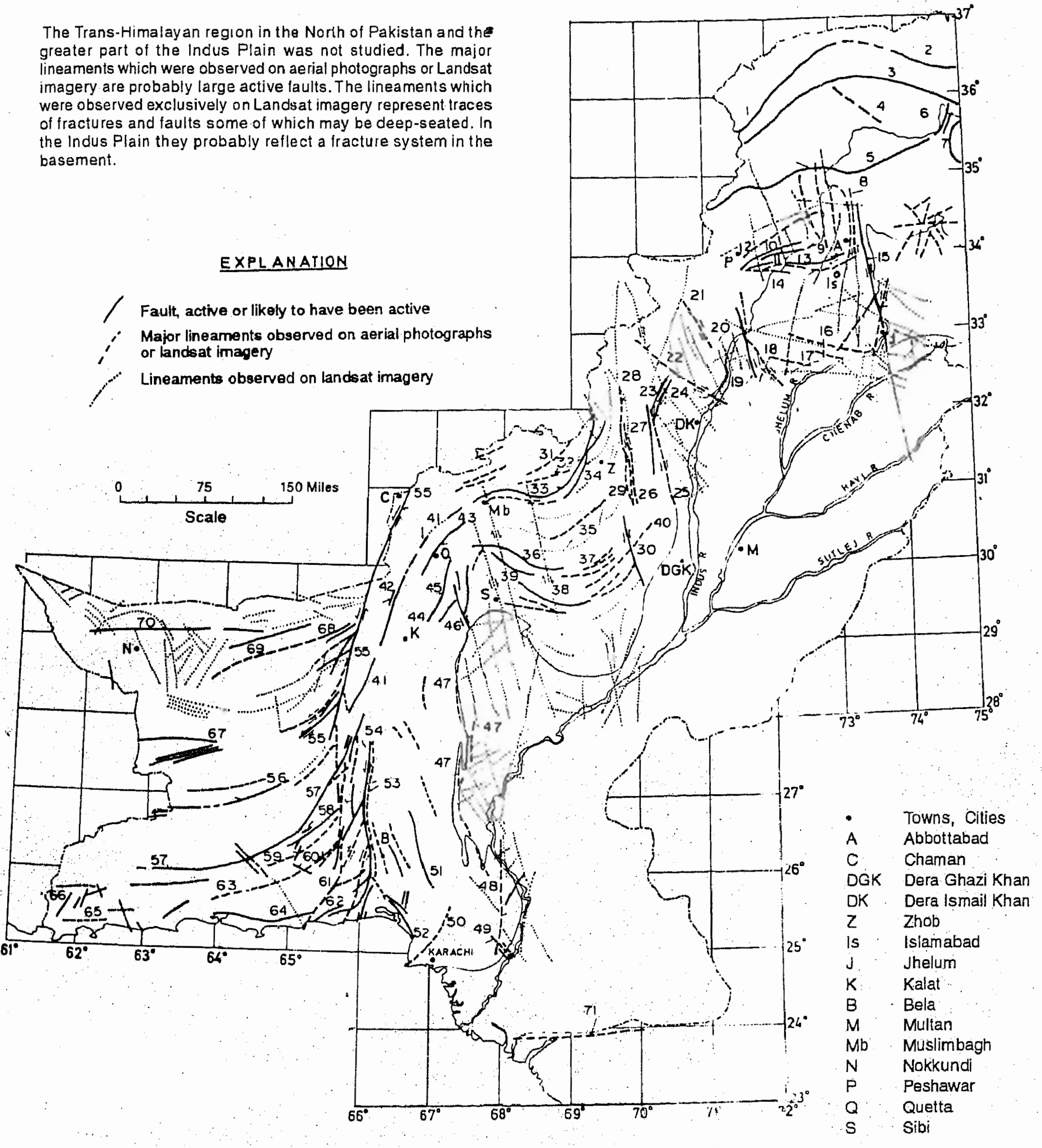
The International Karakoram Project also recorded nine events in the vicinity of Skardu most of which occurred along prominent lineaments seen on satellite imagery and on the ground. These are likely to be active faults, particularly the lineaments along Shigar, Thalle, Hushe and Lower Shyok Valleys (Fig. 8.9).

Kohistan Magmatic Arc

Teleseismic data suggest that the geofractures surrounding this tectonic block comprise a system of active faults (Figs. 8.1, 8.3 and 8.9). We have earlier mentioned that the **Main Karakoram Thrust** (MKT) or the Northern Suture which juxtaposes the Karakoram Block with Kohistan is the source of high seismic activity in the vicinity of Yasin. Also to the southwest in Chitral, and to the east between Ishkuman and Shigar Valleys, a few earthquake epicenters are located along the MKT (Figs. 8.3 and 8.9). **The Main Mantle**

ACTIVE FAULTS SHOWN IN FIG.8.8

1. Reshun
2. Upper Hunza
3. MKT
4. Hamran
5. MMT
6. Raikot–Sassi
7. Harban
8. Sassi–Dassu
8. Shinkiari
9. Indus (Darband)
10. Nowshera
11. Kund (Manki)
12. Peshawar basin F.
13. Attock (Khairabad)
14. MBT (Parachinar–Hissartang–Cambellpur–Murree F.)
15. Jhelum
16. Kallar Kahar
17. Uchchali
18. Salt Range Thrust
19. Kalabagh
20. Surghar Thrust
21. N. Bannu basin F.
22. S. Bannu basin F.
23. Sora Rogha
24. Mandana Kach
25. Chaudhwan
26. Domanda
27. Takht-e-Sulaiman
28. Moghalkot
29. Manikhawa
30. Kingri
31. N. Kakar Khorasan F.
32. S. Kakar Khorasan F.
33. Chukhan Manda
34. Zhob
35. Mekhtar
36. Khalifat
37. Kohlu
38. Tatra
39. Harnai
40. Barkan
41. Ghazaband
42. Bhalla Dor
43. Chiltan–Takhatu
44. Quetta
45. Johan
46. Mach
47. Kirthar
48. Surjan
49. Jhimpir
50. Hab
51. Pab
52. Sonmiani
53. Ornach–Nal
54. Hudishi
55. Chaman
56. Panjgur
57. Hoshab
58. Awaran
59. Bazdar
60. Jhal Jhao
61. Ras Malan
62. Aghor
63. Nai Rud
64. Ormara
65. W. Makran coast F.
66. Kulmir Sunt
67. Ladgasht
68. Ahmadwal
69. Dalbandin
70. Mashki Chah
71. Rann of Kutch F.



The Trans-Himalayan region in the North of Pakistan and the greater part of the Indus Plain was not studied. The major lineaments which were observed on aerial photographs or Landsat imagery are probably large active faults. The lineaments which were observed exclusively on Landsat imagery represent traces of fractures and faults some of which may be deep-seated. In the Indus Plain they probably reflect a fracture system in the basement.

EXPLANATION

- Fault, active or likely to have been active
- Major lineaments observed on aerial photographs or landsat imagery
- Lineaments observed on landsat imagery

•	Towns, Cities
A	Abbottabad
C	Chaman
DGK	Dera Ghazi Khan
DK	Dera Ismail Khan
Z	Zhob
Is	Islamabad
J	Jhelum
K	Kalat
B	Bela
M	Multan
Mb	Muslimbagh
N	Nokkundi
P	Peshawar
Q	Quetta
S	Sibi

Fig. 8.8. Active-fault map of Pakistan (modified from Kazmi 1979).

Thrust, which forms the southern margin of Kohistan, also bears a few epicenters of earthquakes (magnitude 3-5 and focal depth less than 50 km) along its trace between Chilas and Mingora (Fig. 8.3). The topographic depression occupied by the Swat River between Mingora and Chakdarra is along the MMT. The MMT is a complex fault zone, comprising a number of thrust sheets (Chapter 4). It is therefore likely that along this zone there are faults that may be active.

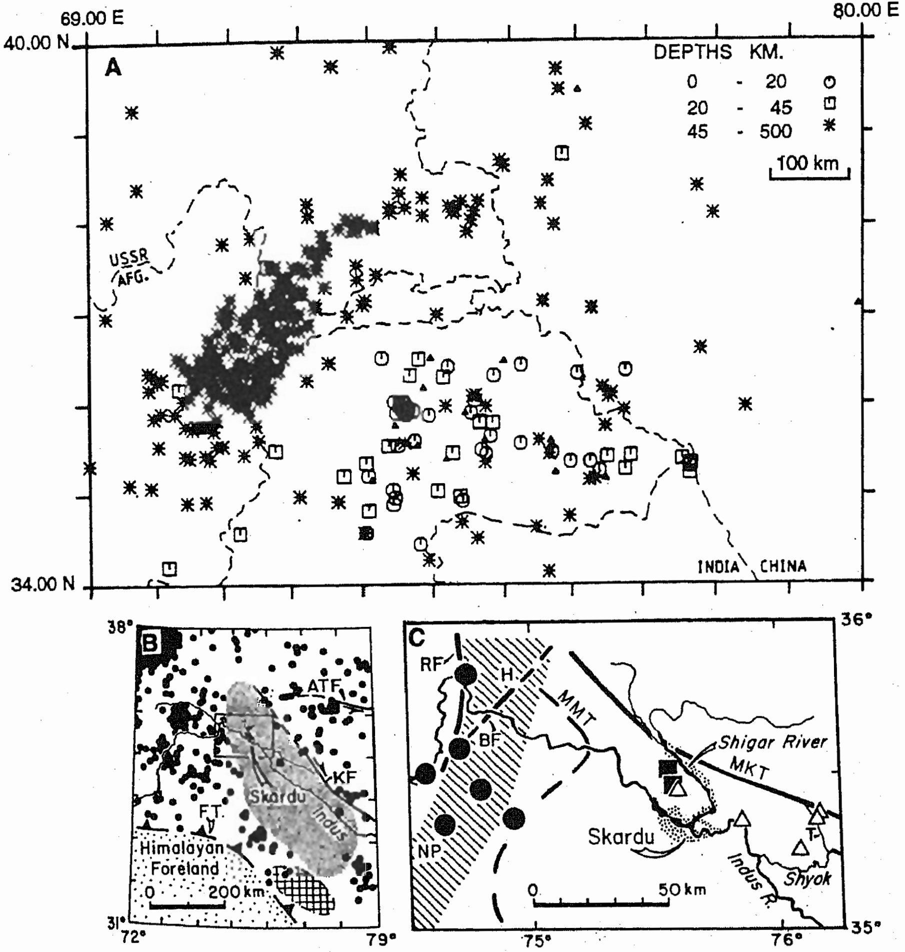


Fig. 8.9. Seismicity in northern Pakistan and adjacent regions. A. Map showing events recorded from 17 August to 14 September 1980 by the International Karakoram Project (IKP) seismic network (from Yielding et al. 1984). B. Seismicity in the NW Himalayas and eastern Karakorams (1914–1981). Grid pattern shows area affected by Kangra earthquake of 1905; grey half-tone shows area which is seismically quiescent. ATF–Altyn Tagh Fault, KF–Karakoram Fault, FT–Frontal Thrust. C. Seismicity within the Skardu region. Solid circles are hypocenters <70 km, solid squares are hypocenters >70 km, open triangles are epicenters recorded by IKP, ruled band is axis of maximum relative uplift along Nanga Parbat–Haramosh Massif. RF-Raikot Fault, BF-Baraluma Fault, MMT-Main Mantle Thrust, MKT-Main Karakoram Thrust (from Cronin 1989).

The Hamran seismic zone is located along the deep, linear, northwest trending Hamran Valley, south of Gupis. Even during the short 3-4 week operation of its seismic network, the International Karakoram Project recorded 13 low magnitude events along the Hamran Valley (Fig. 8.10). In addition, two seismic stations, one at the head of Tangir Valley and the other near the confluence of Gupis and Ishkuman Rivers (Fig. 8.10), recorded several small local earthquakes, the source of which could not be pinpointed as they were not recorded elsewhere (Yielding et al. 1984). The 1972 Hamran and 1981 Darel earthquakes of focal depths 8-10 km and magnitude 6 occurred in the vicinity of the Hamran Valley. Fault plane solutions show high angle reverse fault mechanism for these earthquakes (Jackson and Yielding 1984). Inasmuch as geographical positioning of epicenters based on international seismic network may be in error by 10–15 km, northward shifting of the 1972 and 1981 events by about 15 km would bring them in line with the events recorded by the IKP (Fig. 8.10). There is thus sufficient evidence to suggest that an active fault - the **Hamran Fault,** runs along the Hamran Valley.

The NW Himalayas

The northernmost promontory of the Himalayas, the Nanga Parbat-Haramosh Massif, is tectonically an active feature, characterised by present day uplift rates of > 7 mm/yr (Zeitler 1985), high seismicity (Yielding et al. 1984) and active faulting (Lawrence and Ghauri 1983, Madin et al. 1989, Treloar et al. 1991). Regardless of the controversy whether along the western margin of the Nanga Parbat-Haramosh Massif, the MMT is offset by the Raikot Fault of Madin et al. (1989), or whether the MMT is present along much of the length of the **Raikot-Sassi Fault** zone (Treloar et al. 1991), there is good field evidence to show that this zone is an active fault feature. Near Raikot, the Nanga Parbat Gneisses are thrust over Holocene gravels, in which a 5–10 km thick breccia zone has formed along the Karakoram Highway. Slickensides have developed in this fault zone (Lawrence and Ghauri 1983). Farther to the north, the Raikot-Sassi Fault zone is comprised of a number of parallel or en echelon faults. In the vicinity of Sassi, two of these faults, the **Harban** and **Sassi-Dassu Faults**, cut through the Holocene colluvium and Late Quaternary glacial deposits (Shroder et al. 1989). Thrusting as well as right lateral strike slip faulting is indicated in this fault zone.

Along the western margin of the Hazara-Kashmir Syntaxis, in Shinkiari-Abbottabad region, an active fault zone - the **Shinkiari Fault** zone, has been identified by Kazmi (1979a). He reports two N-S trending parallel fault traces which have dislocated strata, caused stream offsets and on aerial photos and satellite imageries they may be traced across the alluvium. This is apparently a sinistral wrench fault zone and a few earthquake epicentres are located in this zone (Fig. 8.3).

Farther to the southwest, in the Tarbela region, the 1973-1976 record of the Tarbela seismic network (Seeber et al. 1980) has revealed the active **Indus Fault system** comprising a number of steeply dipping parallel faults along the Indus River (Fig. 8.11). The main fault in this region is the northeast trending **Darband Fault** (Calkins et al. 1975) which has formed an overhanging escarpment at the base of Indus River gravels. This 140 to 210 m escarpment, which was clearly exposed during excavations for Tarbela dam foundation, apparently represents a 65° NW dipping reverse fault. The Quaternary deposits in the vicinity of Tarbela have been tilted 20° S to 40° N and are cut by NE trending faults having a few centimeters of displacement. Pre-Tarbela air-photographs also show a fault

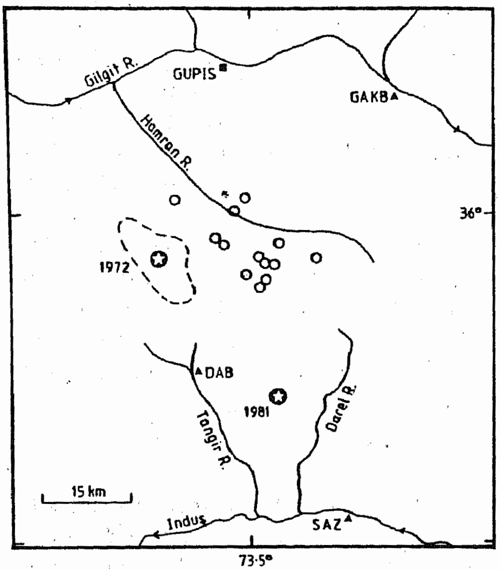


Fig. 8.10. Earthquake locations in the Hamran Seismic Zone recorded by the IKP (1980). Stars show 1972 Hamran and 1981 Darel earthquakes. From Yielding et al. 1984.

with left-lateral stream offsets of 200 to 350 m (Yeats and Hussain 1989).

In Pakistan, the Lesser Himalayas, which include the Attock-Cherat, Kalachitta and Margalla Hill Ranges, lying to the north of the MBT, are very young tectonic features. They are characterised by thrusting along MBT as late as 2.1-1.8 Ma, and movement on the Attock Thrust being as late as 0.5 to 0.1 Ma (Burbank and Beck 1989). It is, therefore, not surprising that this is a region of moderate to high seismicity having a number of active faults.

Along the northern margin of the Attock-Cherat Range, two active E-W trending and apparently right lateral faults– the **Nowshera** and **Kund Faults,** have been mapped by Kazmi (1977, 1979a), largely based on photogeological interpretation. Yeats and Hussain (1987, 1989) have shown that the Nowshera Fault extends about 20 km eastward from Nowshera up to Misri Banda. The fault cuts Holocene lacustrine sediments, and is marked by slickensides suggesting strike-slip motion.

Farther to the south, along the foot-hill region of Attock-Chert Ranges, the **Kund Fault (Manki Fault** of Yeats and Hussain 1987) has formed three en echelon E-W trending left-stepping pressure ridges near Garhi Chandan, Utch Khattak and Walai Villages. The fault cuts Holocene alluvial fan deposits. The fault terraces north of the fault comprise the upthrow side. The eastward diversion of the streams (Kazmi 1979a) and the apparently eastward offset of the young alluvial fans (Yeats and Hussain 1989) suggest right-lateral movement.

According to Nakata et al. (1989) in Pakistan active faults are rather densely distributed on and in the vicinity of the **Main Boundary Thrust** (MBT). The MBT zone is truely comprised of a series of parallel or en echelon faults. In this zone an active fault - the **Parachinar-Murree Fault** of Kazmi (1979b) and Kazmi and Rana (1982) and the Parachinar Fault of Nakata et al. (1989), has been delineated. Yeats and Hussain (1989) mention that lineations and south facing scarps occur throughout the Attock-Cherat Range and Jurassic limestone is faulted over gravels at the western end of the Nizampur basin **(Hissartang Fault)**. Earlier, Kazmi (1979a, 1979b) had mapped two east-west trending parallel active fault lineaments, the **Attock** and **Campbellpur Faults**, extending from the Attock-Cherat Range up to the Haripur Basin. These two faults correspond to the **Khairabad** and **Hissartang Faults** of Yeats and Hussain (1989). A number of teleseismic events are located along these fault zones (Fig. 8.3).

Salt Range and the Trans-Indus Ranges

Along the eastern margin of the Potwar Plateau and the Salt Range, the **Jhelum Fault** (Chapter 4) is apparently an active feature and is responsible for the deep linear gorge of the Jhelum River between Muzaffarabad and Jhelum (Kazmi 1979a). South of Jhelum the fault trace appears as a lineament through the alluvium and has been traced up to the Ravi River (Fig. 4.10). A number of teleseismic events are located along this lineament.

Photogeological interpretation has revealed two WNW trending faults- the **Kallar Kahar** and **Uchhali Faults** in the vicinity of Kallar Kahar and Uchhali Lakes. These are apparently fault-dammed lakes (Kazmi 1979a).

The **Salt Range Thrust** that forms the southern margin of the Salt Range, apparently continues westward along the southeastern margin of the Khisor-Marwat Range and is largely covered by alluvium. However, in the western Salt Range near Kalabagh this thrust is exposed and juxtaposes Paleozoic rocks over Late Pleistocene conglomerates (Gee 1989). A few teleseismic events are located along the southern margin of the central and eastern Salt Range (Kazmi 1979b). According to Allen (1976) there is good evidence for ongoing thrust faulting at the base of Khisor Range. The short term teleseismic data from Tarbela-Chasma network shows that the Salt Range, particularly the Trans-Indus Salt Range is an active structure (Fig. 8.11). Nakata et al. (1989) report E-W trending dip-slip faults in some parts of Kohat Plateau and Surghar Range. The zone of **Surghar Thrust Fault** is characterised by high teleseismic activity (Fig. 8.4).

The western margin of the Salt Range has been truncated by a spectacular right-lateral, strike-slip fault, the **Kalabagh Fault**. Kazmi (1979a,b) indicated that it is an active fault and McDougall and Khan (1990) have shown that this fault has displaced and uplifted the Holocene terrace deposits and splays into smaller, northward trending **Dinghot**, **Ainwan** and **Surghar Faults** (Fig. 4.28).

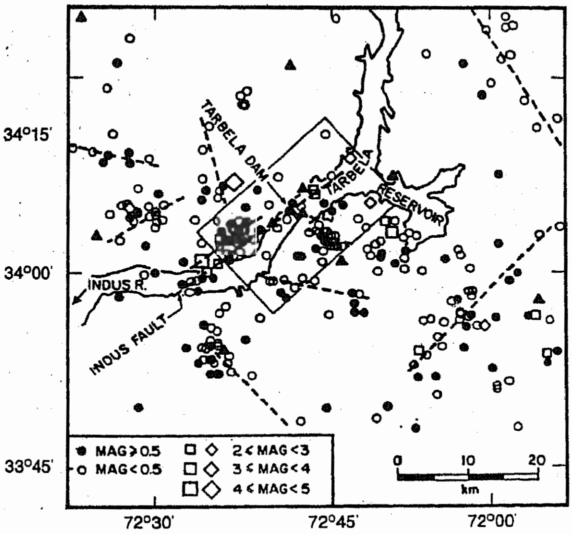


Fig. 8.11. Indus Fault and seismicity in the Tarbela area. Circles are for earthquakes between Aug. 1973–Aug.1976 (depth 15–20 km); squares indicate hypocentral depth 0<h ≤ 20 km; diamonds are epicenters (Aug. 1976 Jan. 1978) with depth 20<h < 30; dashed lines are faults revealed by fault plane solutions and associated seismic patterns. (From Seeber et al. 1980).

Kazmi (1979a) and Nakata et al. (1989) mention NW trending right-lateral, strike-slip faults and dip-slip faults in N-S direction in the Bannu Basin. West of the Bhittani Range, parallel to the Bhittani syncline, there are two parallel NE trending faults–the **Mandanna Kach** and **Sora Rogha Faults** (Hemphill and Kidwai 1973). Stream offsets suggest that these may be active faults.

The Sulaiman Fold-and-Thrust Belt

A number of active faults are located in this zone (Fig. 8.6). In its eastern part, the Sulaiman Range is characterised by a number of N-S oriented strike-slip faults. The left-lateral **Chaudhwan Fault** traverses the eastern margin of the Sulaiman Range and cuts through the Recent alluvium (Gawad 1971, Kazmi 1979a). West of this fault and subparallel to it is the **Domanda Fault** which passes close to Domanda village (Hemphill and Kidwai 1973, Kazmi 1979a). A number of teleseismic events are located along these faults (Figs. 8.4 and 8.6). Nakata et al (1989) show that the Chaudhwan Fault continues southward along the Sulaiman Range up to D. G. Khan and then swings to the SW.

Along the western margin of the Sulaiman Range there is another zone of left-lateral strike-slip faulting - the **Takht-e-Sulaiman Fault** zone which runs parallel to the Chaudhwan Fault (Gawad 1971, Kazmi 1979a). Within the Takht-e-Sulaiman zone there are the **Manikhawa** and **Moghalkot Faults** (Fig. 8.8). This fault zone also has been the site of a number of earthquakes (Fig. 8.6). Farther to the south, a NW trending left-lateral wrench fault, the **Kingri Fault**, is apparently the most active tectonic feature in this region as shown by the concentration of several teleseismic events (Fig. 8.6).

Other active faults in the Sulaiman belt are largely arcuate, convex to the south, thrust or reverse faults which follow the general structure and orographic features. They have been largely located and identified on the basis of photogeological interpretation and teleseismic data (Kazmi 1979a, Nakata 1989). From north to south these are the **Kakar Khorasan, Chukhan Manda, Zhob, Mekhtar, Khalifat, Kohlu, Tatra, Harnai** and **Barkan Faults** (Fig. 8.8).

The Kirthar Fold-and-Thrust Belt

As we have mentioned earlier, the northern part of this belt is characterised by high seismic activity and a number of active faults have been identified. West of Quetta, the **Ghazaband** and **Chiltan-Takhatu** are north-south, parallel, west dipping thrust faults (Kazmi 1979a). Parallel to these is the left-lateral, strike-slip **Quetta Fault** (revealed by the ground rupture during 1935). Other similar faults have been mapped by Kazmi (1979a) and Nakata et al. (1989) in the region between Quetta and Kalat (Fig. 8.8).

In the Mach region, three N-S, NW-SE and NE-SW trending faults (the **Mach Faults**) which affect Holocene alluvium have been identified by Kazmi (1979a). Along the eastern foot-hills of the Kirthar Range three active N-S en echelon faults (North, East and South Kirthar Faults) have been mapped (Kazmi 1979a,b). According to Nakata et al. (1989), along the Kirthar Range Front, there are N-S trending active faults. They are dip-slip or bedding-plane faults. Farther to the south and west of Lakhra, the north-south **Surjan Fault** cuts across the Quaternary deposits. West of Jhimpir, the southern end of this fault is intersected by the northwest trending **Jhimpir Fault**. The intersection of these two faults is characterised by at least four teleseismic events of shallow focal depth and magnitude 3-6.

In the western part of this fold belt, the **Pab** and **Hab** Faults are other major active faults (Kazmi 1979a,b, Kazmi and Rana 1982, Nakata et al. 1982). The Pab Fault has dislocated vertically the Quaternary alluvial fans. The Hab Valley is traversed by the Hab Fault.

Chaman and Ornach–Nal Faults

These faults comprise the western margin of the Indo-Pakistan plate and have been discussed in detail in Chapter 4. The teleseismic events along the Chaman Fault have been briefly reviewed earlier in this Chapter.

Makran Accretionary Zone

The Chaman Fault zone continues into the NE corner of this zone. On aerial photographs the active trace of the Chaman Fault may be seen cutting the Holocene alluvium and curving southwestward along the margin of the Kharan basin (Fig. 8.8). Field reconnaissance and photogeological interpretation (HSC 1960, Kazmi 1979b) show that south of Kharan, this fault zone becomes diffused and splits into several strands some of which continue southward. The Panjgur Valley is straddled by several such arcuate faults, which link up with the Chaman Fault zone. This valley also is characterised by a number of teleseismic events. According to Kazmi (1979a,b), it is a fault valley and the Panjgur Fault zone is an active feature. Farther to the south, there is the Hoshab Fault zone which runs parallel to the Panjgur Fault zone. A number of earthquake epicenters occur in this region along these neotectonic features.

The left-lateral Ornach–Nal Transform Fault forms the eastern margin of the Makran zone. Based on photogeological interpretation, Kazmi (1979a,b) has inferred three NE-SW striking left-lateral active faults- the **Aghor**, **Ras Malan** and **Nai Rud Faults** which link up with the Ornach–Nal Fault to the east (Fig. 8.8). Dislocation of strata, stream deflections and indentation of coastline are observed along these faults.

A number of smaller faults, with signs of recent activity, largely in the form of fault scarps in Recent to Sub-Recent alluvial deposits, have been mapped in the Makran region (Fig. 8.8). These are the **Hudishi**, **Awaran**, **Bazdar**, **Jhal Jhao**, and **Kulmir Sunt Faults** (HSC 1960, Kazmi 1979a).

Kharan Basin and Chagai Magmatic Arc

Southwest of Ahmadwal, the NE-SW trending **Ahmadwal Fault** runs along the northern margin of the Ras Koh and cuts the Subrecent alluvium (HSC, 1960, Kazmi 1979a,b, Nakata 1989). Photogeological interpretation has revealed several other features which may be active (Fig.8.8).

9

Mineral Deposits

Minerals have been mined in the territory that now comprises Pakistan at least since the eighteenth century. There are reports of old mines and workings of arsenic deposits in Tirich Valley (Chitral); antimony deposits in Karangali Hill in Salt Range, Zaimukht Hill near Thal and at Krinj in Chitral; copper deposits near Robat (Chagai District) and in south Waziristan; galena deposits in Chitral, Shekhan (Khuzdar), Saindak and Miran Koh (NW of Chagai); manganese ore near Las Bela; and rock salt deposits at Khewra, Sardai and Warcha in the Salt Range. Regular mining of coal began in the 1870's and the Hindubagh (now Muslimbagh) chromite mines were opened up in 1903 (Heron et al. 1954). Reconnaissance surveys during the nineteenth century (Chapter 1) provided considerable data on mineral occurrences which were summarised by La Touche (1918). The Geological Survey of India regularly updated the information on mineral deposits through its publication–'Quinquennial review of the mineral production of India' starting from 1898-1903. The position concerning mineral deposits of Pakistan at the time of Independence was summed up by Gee (n.d.). This paper may be taken as a bench-mark paper on mineral deposits of Pakistan. It shows that only antimony, coal, gypsum, rock salt, sulphur and limestone were mined, whereas a number of occurrences of antimony, arsenic, asbestos, barite, bauxite, bentonite, beryl, bismuth, copper, iron ore, lead, magnesite, manganese and mica were known. A subsequent review by Heron et al. (1954) shows that within six years several new mineral occurrences were recorded as a result of reconnaissance surveys by the Geological Survey of Pakistan though no new minerals were mined. Since the early fifties several new minerals have been discovered and evaluated, and a number of these are now being mined as shown in Table 9.1.

METALLOGENIC ZONES

Minerals occur in distinct belts or metallogenic zones. Each zone is characterised by its unique geological environment. In Pakistan these zones broadly conform to the tectono-stratigraphic zones discussed earlier in Chapter 4. From the standpoint of mineral occurrences they may be broadly grouped as follows (Fig. 9.1):

1. *Indus basement zone.* The basement rocks crop out in the Sargodha–Shahpur region and Nagar Parkar area. They comprise metasediments, metavolcanics, and magmatic sequence of the Indian shield, characterised by gold-bearing hematite deposits.
2. *Indus platform and foredeep zone.* It contains the sedimentary cover with major hydrocarbon deposits–coal, oil and gas.
3. *Foreland sedimentary fold-and-thrust belt.* It is largely characterised by sedimentary, non-metallic mineral deposits and hydrocarbons. However, iron ores and bauxite deposits occur along major unconformities.

Table 9.1. Economic minerals of Pakistan.

LARGE DEPOSITS	MEDIUM TO SMALL DEPOSITS	SHOWINGS WITH GOOD PROSPECTS
*Barite	*Antimony	Antimony
*Bentonite	*Asbestos	Arsenic
Beryl	*Bauxite	Bismuth
*Chromite	*Celestite	Cadmium
*Copper	*China clay	Cobalt
*Coal	*Fluorite	Chromite
*Dolomite	*Gemstones	Copper
*Fireclay	Gold	Lead
*Gypsum	Graphite	Iron ore
*Iron ore	*Magnesite	Gold
Lead	*Manganese	Graphite
*Limestone	Mica	Manganese
*Marble	*Petroleum	Nickel
*Natural gas	Potash salt	Platinum
*Nepheline syenite	*Radio active minerals	Silver
*Ochre	*Rock phosphate	Tungsten
*Quartz	Silver	Zinc
*Rock salt	*Soapstone	
*Silica sand	Sulphur	
Zinc	Vermiculite	

Asterisks indicate minerals in production.

4. *Ophiolitic thrust belt.* It contains chromite, manganese, talc, soapstone, nickel, platinum and Mississippi and Sedex type lead-zinc deposits.
5. *Himalayan crystalline zone.* It comprises a magmatic and metamorphic belt, largely associated with (a) epigenetic, polymetallic deposits generated by metamorphic fluids (Cu, Fe, Pb, Ba), (b) exhalative type, volcanogenic polymetallic sulphides (Pb-Zn in Precambrian metavolcanics), (c) mineralisation affiliated with pegmatites and anatectic granites (gemstones, scheelite, fluorite), and (d) Rare-earth minerals in carbonatites and alkaline complexes.
6. *Kohistan magmatic arc.* Mineralisation in this arc consists mainly of minor chromite deposits, contact-metasomatic iron and Pb-Zn deposits, and volcano genic polymetallic sulphides characterised by extensive alteration zones and gossans (possibly underlain by massive sulphide bodies).
7. *Karakoram block.* It comprises a highly tectonised, magmatic, sedimentary and metamorphic terrain which contains polymetallic sulphides (Cu, Pb, Sb, As, Zn) with gold and silver, and deposits of scheelite. They occur largely as fissure fillings or replacement deposits along major faults and are related to pegmatites and granitic intrusions. There are extensive pegmatite fields with gem-bearing pegmatites.
8. *Balochistan flysch basin.* Apart from possible prospects for oil and gas, present data suggest that metallogenically it is largely a barren zone, except for the occurrence of antimony in the Chaman Fault zone (Qila Abdullah) and showings of Sb, Pb, Zn, Au and Ag in ferruginous quartz-calcite stockworks in Eocene-Oligocene shales and sandstones.
9. *Chagai magmatic arc.* This zone comprises a volcanogenic, polymetallic sulphide belt characterised by Cretaceous to Neogene intrusive and extrusive calc-alkaline magmatism. It contains porphyry copper, Kuroko-type massive sulphides, vein type Cu, Pb, Zn, Ag and Mn deposits, volcanogenic, contact-metasomatic massive magnetite-hematite deposits and solfataric-type sulphur deposits.

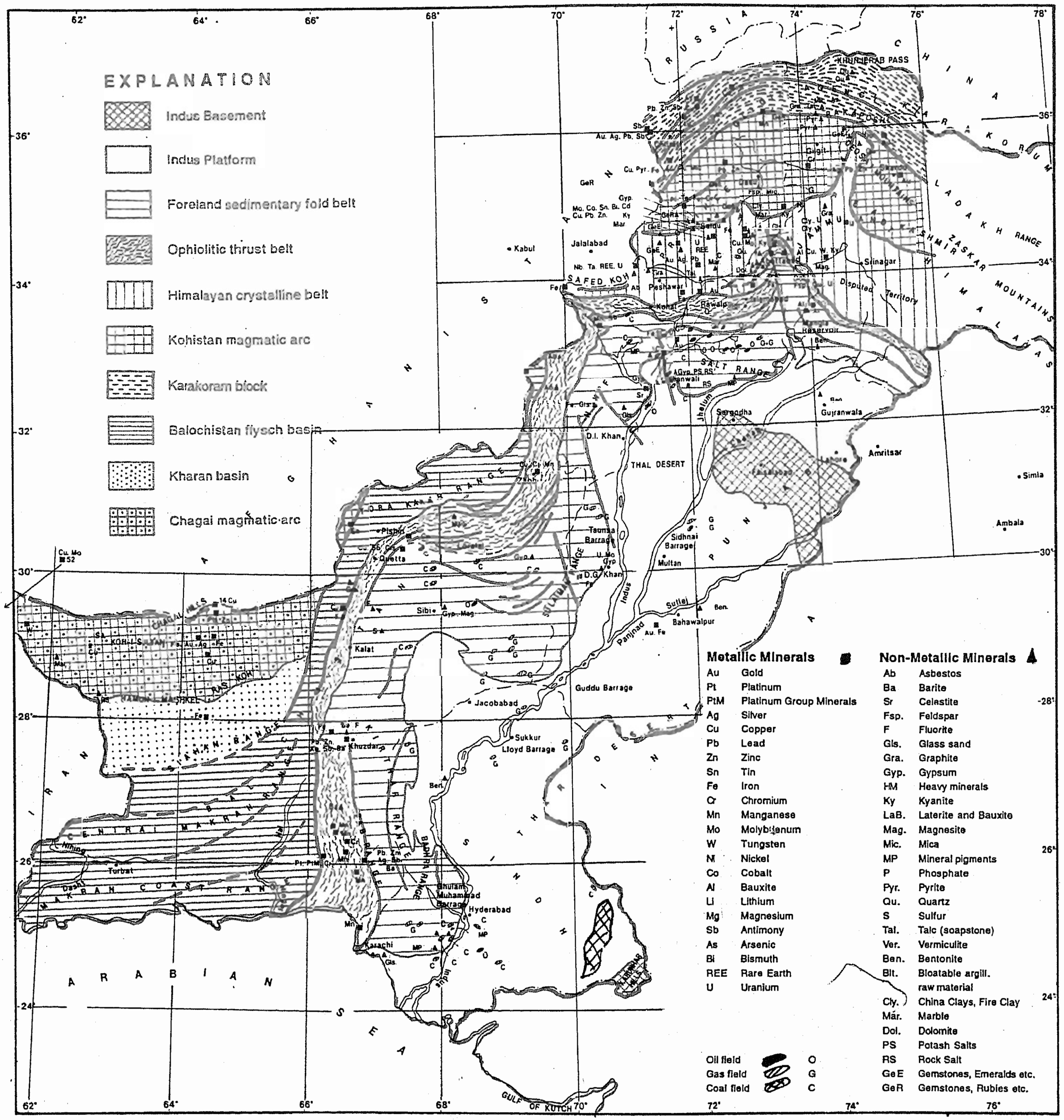


Fig. 9.1. Minerallogenic map of Pakistan. (Compiled from Kazmi and Rana 1982 Ahmed and Abid 1983, Bender and Raza 1995).

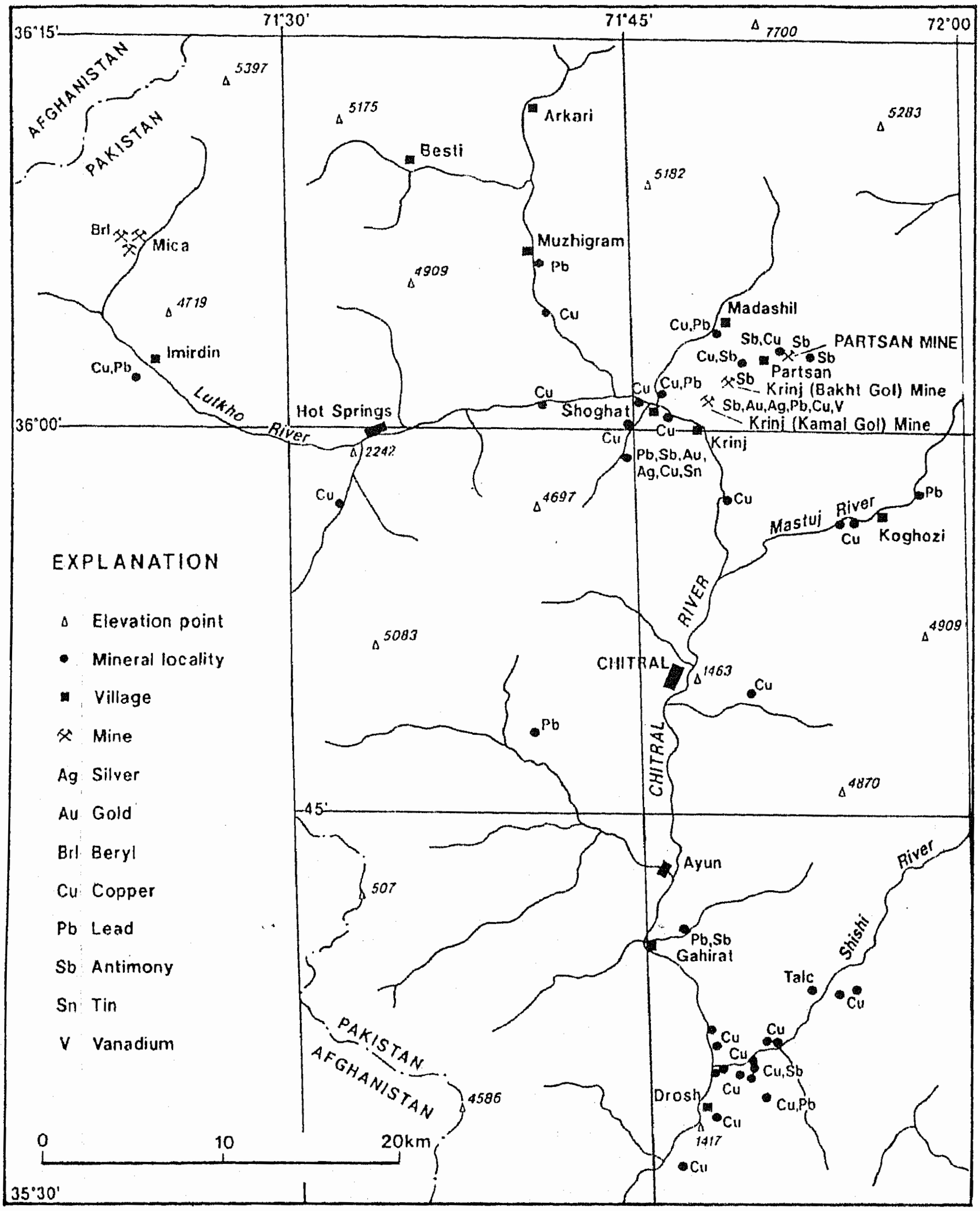


Fig. 9.2. Mineral localities in the Chitral-Partsan area, northern Pakistan (from Calkins et al. 1981).

Relevant details of the mineral deposits in different metallogenic zones have been given in Tables 9.2 to 9.15 and a brief account of the more significant mineral deposits is given in the following pages.

METALLIC MINERALS

Antimony

Showings of antimony have been reported from Salt Range (Karangali-Hill), Kurram Valley (Zaimukht Hills), Khuzdar (Shekran), Qila Abdullah and Krinj–Partsan region of Chitral (Table 9.2). Only the Krinj deposit (Fig. 9.2) is of some significance where 50–120 tonnes of the ore has been mined annually. The ore contains 29–37.6% Sb and significant amounts of silver, gold, and vanadium (Ahmad 1969, Calkins et al. 1981).

Chromite

Alpine-type chromite deposits occur in the suture zones along the northwestern and western margins of the Indo-Pakistan plate. Chromite is found as pods, lenses, pipes and irregularly-shaped bodies in dunites within ultramafic tectonites and ultramafic cumulates associated with ophiolites. It has been mined in the Muslimbagh area since 1903. In subsequent years several showings and deposits have been discovered (Table 9.2). The main deposits are located in the Jijal (Besham, Fig. 9.3), Harichand–Skhakot (Dargai), Muslimbagh (Zhob), Sonaro (Wad), and Bunap (Ras Koh) areas. The average annual production since Independence has been about 14,000 tonnes, though in some years it has exceeded 30,000 tonnes (Butt and Latif 1992). The entire production is exported. The exported ore contains over 46% Cr_2O_3, 10–15 Al_2O_3, less than 10% SiO_2, and Cr:Fe ratio of over 2.8.

The total resources of high-grade chromite are estimated at about 3.6 million tonnes (Islam 1993), whereas from Muslimbagh alone more than 0.8 million tonnes have been already mined. The Muslimbagh deposits are likely to yield a sustained annual production of 40,000 tonnes for several years (Working Group on Minerals 1978).

Copper ore

Deposits and showings of copper ore are largely confined to the ophiolitic thrust belt and suture zones, Kohistan magmatic arc, Karakoram block and the Chagai magmatic arc (Table 9.2). In these regions copper ore is intimately related to igneous intrusions and volcanic rocks, and occurs as (a) vein deposits with silver, gold and other metals (e.g., epigenetic, polymetallic and metamorphic deposits in Chitral), (b) strata-bound disseminations in volcanogenic sequences (e.g., Manto-type copper deposit at Talaruk), (c) massive sulphide bodies in volcanics (e.g., Kuroko-type Ann Dhora deposit in Las Bela and others in the Gilgit region), (d) contactmetasomatic-type small deposits (e.g. Kundi–Baluchap and Mashki Chah in Chagai area), (f) porphyry copper-molybdenum deposits (e.g. Saindak and Dasht-e-Kain in Chagai area), (g) complex, multiphase, ophiolite-associated massive sulphide (e.g., Boya in Waziristan), and as a minor contituent in volcanogenic magnetite deposits (e.g., Dommel Nissar in Chitral and Chilgazi in Chagai area).

Associated with the magmatic and volcanic sequences in the Bela ophiolite belt, Waziristan ophiolite belt, Chagai magmatic arc, Kohistan island arc and the Tirich Mir zone of the Karakoram block, there are numerous and extensive alteration zones and gossans. Some of these contain traces of copper and other metals (Table 9.2) and

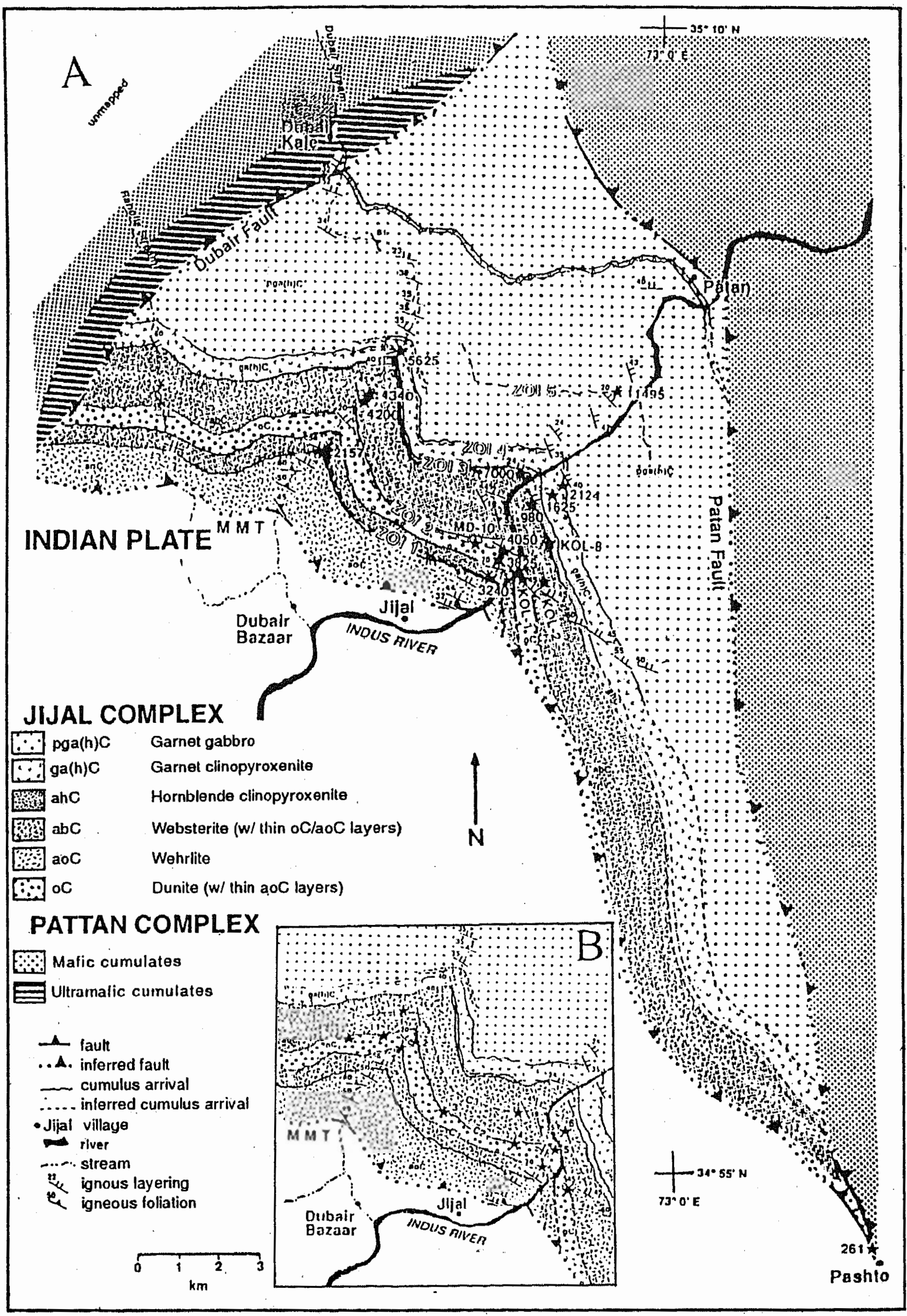


Fig. 9.3. Geological map of the Jijal Chromite deposits. In (A) stars with numbers indicate sites of precious metal-mineralised samples; zol= zone of interest for precious metals. In (B) stars with numbers show chromite mines/prospects: 1= Shungial, 2= Kuroo, 3= Gabara, 4= Manidara, 5= Kokial, 6= Serai; C1, C2, C3= chromite bearing units. (From Miller et al. 1991).

Table 9.2. Metallic deposites : location, geological setting and potential.

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Antimony ore.	Foreland fold & thrust belt	-Karangali Hill 32°45':73°05' -Salt Range	Galena with small amount of stibnite	Not known	Not known	Trivial	Minor mining reported	Heron et al. 1954.
	Ophiolitic thrust belt.	-Zaimukht Hills 32°22':70°35' -Kurram Valley	—	—	Not known	Trivial	Ancient mine reported	Heron et al. 1954.
		-Shekran 27°85':66°38' 12 Km-NW of Khuzdar	—	Antimony in slags of ancient lead mines reported. Iron, lead and barite mineralisation in Jurassic limestone of Shirinab Fm.	—	Minor	As minor constituent in iron, lead baryte ore.	Heron et al. 1954.
	Karokaram block	-Krinj-Partsan 36°00':71°50' 18 km N of Chitral Town		Mineralisation in the vicinity of Pasti Fault and Reshun Thrust. Largely fracture fillings. Metamorphic origin.			Promising prospect.	Sillitoe 1979.
	″	-(a) Awireth Gol.	Very fine grained Pb-Sb sulphides, with chalcopyrite, pyrite & boulangerite : quartz, dolomite gangue.	Mineralised vein along a fault breccia at contact of Cretaceous Reshun Fm. and Paleozoic (?) phyllites. The fault is the eastern boundary of a horsetail structure.	Sb 12.8-17%, Pb 45.2 -56.6%, Ag(av) 980 ppm, Au 77 ppm.	Minor	—	Calkins et al. 1981.
	″	(b) Krinj.	Stibnite, Quartz.	Quartz-stibnite veins emplaced in Chitral Slate, at the faulted contact between the slate and overlying limestone and marble of Reshun Fm.Mineralised zone brecciated.	Sb 29-37.6%, Ag < 2- 22 ppm, Au 0.6- 4.3 ppm, V. 15-35ppm.	Medium (60,000 tonnes)	Annual production . 50-120 t. Maximum production 650 t. in 1940.	Calkins et al. 1981. Austromineral 1978.
	″	(c) Partsan	Stibnite, Quartz.	Mineralised quartz veins in pyritised phyllite of Lun (Sarikol) Shale or in dolomites of Shogram Fm. adjacent to altered, faulted contact between dolomites and Lun Shale.	Sb 4.8-50.5%, Cu 0.24-2.8 %, Ag <2-750 ppm, V. 25-38 ppm.	Minor	At one locality 1,000 ppm Ag, 16 ppm Au, 0.1 % Zn, 7 % Pb.	Calkins et al. 1981.

(Continued)

Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit, and location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Antimony ore (contd.)	Chaman Fault zone.	Qila Abdullah 30°44':66°40'	Stibnite & valentinite in gangue of quartz calcite, limonite, & hematite.	In slate and sandstone of Khojak Fm. Fissure fillings in oxidised zones. Mineralised quartz veins along en echelon faults which are part of Chaman Transform Fault system. Product of dynamo-thermal metamorphism.	Sb 5-30%, Ba 10.73%, Fe 3.36 %, Pb 0.3 %, As 0.22 %.	Minor	Traces of Hg, Cu, Zn. Sporadic small scale mining.	Ahmad 1975. HSC 1960. Sillitoe 1975.
Arsenic	Kohistan magmatic arc.	Dainyor Nala 15 km NE of Gilgit. Bagrot Nala 20 km NE of Gilgit.	Arsenopyrite, chalcopyrite, malachite, pyrite.	In extensive alteration zones, sulphide mineralisation at contact of metavolcanics (Rakaposhi Volcanic Complex) and Early Eocene diorites.	Not known	Un-explored.	Promising prospect for massive sulphide.	Kazmi 1951.
	Karakoram block	Lundku-Mirgasht 36°26':72°17' (Tirich Gol Valley).	Orpiment, realgar	Mineralisation hosted in Permian Lst (Lun Shale?), stratabound replacement deposit. Mineralisation related to dolerite dykes cutting limestone and calc. shales.	Not known.	Un-explored.	Deposits at altitudes of 11,000 to 15,000 ft. Small scale sporadic mining in the past.	Tipper 1921. Searle 1991.
Bismuth Cadmium Cobalt	Himalayan crystalline zone (Besham nappe).	Lahor & Pazang (3 & 4 km N & SE of Besham 34°56':72°52')	—	Associated with Pb-Zn deposits in altered pegmatite and granite complex enclosing Proterozoic metasediments. For detail see Pb-Zn deposits in text.	Bi- 0.2-0.8%, Cd- 0.1-0.2%, Co- 0.005%, Sn-0.12%, W-400 ppm.	Un-explored.	—	Chaudhry et al. 1983. Ashraf et al. 1980a.
Chromite	Ophiolitic thrust belt and suture zone-Dargai klippe.	Harichand-Sakhakot-Qila, west of Dargai (34°28':71°54')	Chromite	Occurs in irregular veins and lenses in the pyroxenites and dunites of the Ultramafic Complex (ophiolites and melanges of the Dargai klippe) in the Malakand region.	Cr_2O_3- 24.3-64.1%, Al_2O_3- 6.5-42.7%, Cr/Fe -1.4 to 4.5:1. Refractory grade.	Several small deposits (more than 50,000t. estimated).	Intermittent mining.	Ahmad 1969. Ahmed 1983.
	Ophiolitic thrust belt and suture zone.	Boya 32°57':69°50' (North Waziristan).	Chromite.	Occurs as disseminations and small iregular pods in dunites of the Waziristan ophiolitic complex. Occurrences at Sherkai, Madar algad and Tut Nari.	Apparantly low grade.	Showings.		Khan et al. 1982.

(continued)

Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Chromite (contd.)	Ophiolitic thrust belt.	-Naweoba, E of Zhob town 31°20':69°27'	Chromite	In ultramafic rocks of the obducted masses of ophiolites and melange.	Cr_2O_3-36.7-46.5%, Al_2O_3-20%, Cr/Fe-2.9:1, Refractory grade.	Several minor deposits	Alpine type deposits	WPIDC 1970c. Bilgrami 1964a. Bilgrami 1964b.
		-Zizha 24 km NE of Zhob	Chromite	" "				
	"	-Khanozai 30°36':67°20' 46 km WSW of Muslimbagh	Chromite	Chromite occurs as steeply dipping lenticular bodies of varying thickness in ultramafic rocks of the obducted masses of ophiolites.	Cr_2O_3-49.3-52.6%, (Av-45%), Cr/Fe-2.7:1 to 3.5:1.	Small	Mined sporadically.	Ahmad 1975. Bilgrami 1964 a,b. Ahmad 1969.
	"	-Jang Tor 9 km S of Muslimbagh 30°50':67°44'	Chromite	Chromite occurs as irregular veins and lenses in dunites of the ultramafic complex that forms a part of the Zhob Ophiolite.	High grade ore Cr_2O_3-48%+, Cr/Fe-3:1.	Several small deposits.	Being mined.	Ahmad 1975. Bilgrami 1964 a,b. Ahmad 1969.
	"	-Saplai Tor 8 km SE of Muslimbagh	Chromite	Chromite occurs as irregular pipes or lenses or as disseminations in layers traceable for several thousand feet; layers folded at places. Occurs in dunites as at Jang Tor.	Med. grade ore Cr_2O_3-44.0-52.5%, Cr/Fe-3:1.	Several small deposits.	Being mined.	Ahmad 1975. Bilgrami 1964 a,b. Ahmad 1969.
	"	-Nasai 35 km SE of Muslimbagh	Chromite	Chromite occurs largely as layered disseminations in dunites of the Zhob Ophiolites.	Cr_2O_3-39-49 %, Cr/Fe- 2.1 to 3.1:1. Low grade ore.	Small	Mined sporadically.	Ahmad 1975. Bilgrami 1964 a,b. Ahmad 1969.
	"	-Sonaro 26°21':62°28' (Pat Nadi & Dirya deposits) near Wad.	Chromite	Small lenses to massive tabular bodies associated with harzburgites and dunites which from part of ophiolitic tectonites in Kanar Melange.	—	Large, probable reseres 1.3 m.t.	Being mined, production up to 10,000 tonnes p.a.	Islam et al. 1993.
	Indus suture zone.	-Jijal 35°01:72°52' N of Besham	Chromite	Along the northern margin of the MMT zone chromite occurs largely as strata-bound lenses associated with a repetitive dunite-wehrlite-websterite sequence in the Jijal Complex. This complex forms a part of the layered cumulate complex developed towards the base of the Kohistan magmatic arc.	Cr_2O_3-22-55%, Cr/Fe-2.8:1 to 3.6:1. Metallurgical grade.	Several dozen chromite pods and lenses reported. Identified reserves over 0.6 m.t.	Large reserves may be expected. Mining in progress.	Ashraf et al. 1982. Miller et al. 1991.

(continued)

Table 9.2. Metallic mineral deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Chromite (contd.)	Kohistan magmatic arc.	-Chilas 35°26:74°04'	Chromite	Chromitite bands in dunite occur in the ultramafic rocks of the stratiform? Chilas Complex. This Complex comprises gabro-norites and UMA association of ultramafic, mafic and anorthositic rocks. The latter forms lenses of layered cumulates in gabronorite bodies. The Chilas Complex forms the medial zone of the Kohistan arc and has intruded the basal Kamila amphibolites.	Low grade, Cr_2O_3-26 %, Al_2O_3-26 %, FeO-37 %, MgO-9 %.	Not known	Unexplored	Jan et al. 1984. Khan et al.
	Chagai magmatic arc.	-Ras Koh (a) Nag-Bunap 29°50':65°18' (b) Rayo Nai, 28°58':64°43' (S of Nok Chah)	Chromite	Chromite occurs as disseminations, nodules, lenses and veins in the ultramafic rocks of Ras Koh Range; 9 deposits occur in Nag-Bunap area and 7 in Rayo Nai region.	Cr_2O_3- 47-57 %, Cr/Fe-3:1.	Small, over 30,000+ tonnes.	Mined sporadically.	HSC 1960. Ahmad 1975.
Copper ore.	Himalayan crystalline zone.	-Babusar 34°08':74°02'	Chalcopyrite, pyrrhotite, fluorite.	In quartz veins cutting Precambrian metasediments. Mineralisation associated with Late Mesozoic to Early Tertiary granites and diorites.	Not known	Trivial	Un-explored	Ahmad 1969.
		-Phalkot 34°09':73°22'	Malachite, chalcopyrite.	In veins cutting Hazara Fm.	Not known	Trivial	Un-explored	Ahmad 1969.
		-Galdanian 34°15':73°19'	Malachite, chalcopyrite.	In sandstone associated with hematite deposits.	Not known	Trivial	Un-explored	Ahmad 1969.
	Foreland fold-and-thrust belt.	-Katha 32°39':72°26' -Musa Khel 32°38':71°45' -Nilawahan Gorge 32°39':72°36' -Warcha 32°29':71°59'	Malachite, cuprite, native copper.	Native copper and copper minerals occur as nodules, specs and stains disseminated in in Sardhai Shale and Warcha Sandstone (Late Paleozoic).	Low grade, traces to 110-1,800 ppm. Cu; some samples contain 0.2-0.7% copper.	Not known	Un-explored	Memon 1965. Qureshi 1980. WPIDC 1970b. Ahmad 1969.
Copper ore	Ophiolitic thrust belt.	-Gujarghuna 33°59':69°56' (2 km NNW of this village) Parachinar Dist.	Chalcopyrite (altered to goethite, hematite, & malachite).	Mineralisation occurs in quartzites, underlain by shale and limestone having a "baked" appearance. Shale intruded by basalt sill. This mineralisation is along major fault, the melange boundary zone (MBZ) of Beck et al. 1995.	Not known	Not known	Un explored. (also referred to as Neemtota Cu-Pb mineralisation of Piewar).	Meissner et at. 1975. Beck et at. 1995. Badshah 1983.

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Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Copper ore (contd.)	Ophiolitic thust-belt and suture zone.	Shinkai (Boya) 32°55':69°52' (Waziristan)	Chalcopyrite, bornite, pyrite, malachite, azurite, cuprite, tenorite, brochantite, covellite.	Several copper deposits occur in the Shinkai and adjacent regions, in the Khost suture zone (Beck et al. 1995), associated with the ophiolites and volcanics of the Waziristan igneous complex and the melanges. The igneous complex has been thrust eastward on Mesozoic shelf deposits and includes a sequence of basal ultramafics, an intermadiate zone of gabbros and dolerites and an upper part of volcanics, interspersed with tectonic slices of melanges and intruded by granites and diorites (calc-alkaline magmatism). Mineralisation largely along "Boya-Spinkamar-Preghar line". 15 mineralised bodies occur in this zone as sheet breccia in contact with dolerite sill along margin of ultramafics. Copper in quartz veins occur at Kambat, Gabbari, Mami Rogha, Sarpunga, Papure Tip; in alteration zones at contact of sedimentary and volcanic rocks at Papure Tip; in diorite, granodiorite, and volcanic rocks at Lwargi and Dangarkonr.	Average Cu content in 6 bore holes: 0.37 %, 0.89 %, 0.11 %, 0.43 %, 0.52 % & 0.79 %.	Inferred reserves down to 150 m are 120 m.t. of 0.3 to 0.5% Cu	Promising deposit. Breccia-pipe-type mineralisation.	Badshah 1983. Beck et al. 1995. Khan et al. 1982.
	"	-Zhob 31°21':69°26' -Sange Gar 19 km N of Zhob. -Zizha, 24 km NE of Zhob. -Shin Ghar 14 km SE of Zhob. -Otman, near Jalat Killi.	Copper sulphides & carbonates with manganese & pyrrhotite.	Associated with chromite bearing ultramafic rocks in the Zhob Melange.	Not known	Trivial		Ahmad 1969. Heron et al. 1954.
		-Nasai 30°50':68°02'	Copper sulphide.	Mineralisation in the contact zone, between Dungan Limestone (Paleocene) and ultramafic rocks.	Not known	Trivial		Ahmad 1969. Heron et al. 1954.

(continued)

Table 9.2. Metallic deposites : location, geological setting and potential. (Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Copper ore (contd.)	Ophiolitic thrust belt & suture zone.	-Tor Tangi 30°33':67°47'	Copper minerals with magnetite.	Occurs as lenses in serpentinised ultramafic, mafic rocks.	Not known	Trivial	—	Ahmad 1969.
	Kohistan magmatic arc.	-Dainyor Nala NW of Gilgit 35°55':74°17'		A number of gossans and alteration stains occur along fault zones and contact of metavolcanics and diorites/ and granodiorites.	—	—	Unexplored. May reveal massive sulphide bodies.	Bughio et al. 1970. Kazmi 1951b.
		(a) Barit	Chalcopyrite, pyrite.	In quartz veins ; fissure fillings in diorite.	0.4-0.6% Cu.	Spread over 60 acres.	Not explored.	Bughio et al. 1970. Kazmi 1951b.
		(b) Bulashgah 2 km E of Barit.	Malachite, azurite, chalcopyrite, bornite, chrysocolla, pyrite.	An extensive gossan occurs within highly altered metavolcanics at their contact with diorite. Trenching and pitting shows a wide zone of copper dissemination. A 23 m thick alteration zone contains on an average 0.86% Cu.	0.2-2.5% Cu, (average 0.7% Cu).	Probable reserves 0.5 m.t. Mineralised zone 700 m long 25 m thick.	Promising prospect.	Bughio et al. 1970. Kazmi 1951b.
	Kohistan magmatic arc.	(c) Majadar 3 km E of Bulashgah.	Malachite, azurite, chrysocolla, magnetite.	The mineralised zone follows a steeply dipping fracture zone which is weathered brownish yellow on the surface. The host rock is an ultramafic pod in metavolcanics and contains olivine and pyroxene. It has been invaded by quartz veins bearing green tourmaline. About 1.5 km S of Majadar there is an ultramafic pod in diorites containing a 0.8 m thick mineralised zone with 1.08 % Copper.	0.4 % Cu, 8.4 % Fe, (Only one surface sample tested).	Small	Unexplored.	Bughio et al. 1970. Kazmi 1951b.
	Kohistan magmatic arc.	(d) Bora Nala 3 km NE of Barit.	Pyrite, malachite, azurite.	Mineralisation characterised by a reddish weathered zone in metavolcanics with quartz stockwork. This zone is 70 m thick and 330 m long and largely contains disseminated pyrite grains and copper stained bands a few cm to a few metres across.	Traces to 1.3 % Cu.	Not estimated	Unexplored, Promising prospect.	Bughio et al. 1970. Kazmi 1951b.

(continued)

Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Copper ore (contd.)	Kohistan magmatic arc.	-Bagrot Nala (18 km E of Gilgit).	Malachite, chalcopyrite, pyrite.	Similar to Balushgah, mineralisation in metavolcanics.	Not known	—	Unexplored. Promising prospect. Massive sulphide type.	Kazmi(Verbal Communi-cation).
	"	-Nomal 16 km N of Gilgit.	Pyrite, chalcopyrite, malachite, arsenopyrite.	Extensive weathered zones, comprising a powdery yellow mass, cover the hill slopes on right bank of Hunza River. The weathered zone contains sulphur and alum. Scattered rock outcrops beneath the weathered zone contain sulphide mineralisation. There is widespread silicification and sericitisation and the mineralised zone occurs at the contact of metavolcanics and granodioritic intrusions.	Average of several samples: Fe 4.17 %, Mn-944 ppm, Cu-353 ppm, Zn-108 ppm, Pb-550 ppm, Two samples: Au-0.65 & 1.03ppm, Ag-69 ppm, Mo-Traces.	Not known	Good prospect for massive sulphide, gold and silver.	Kazmi 1977.
	"	-Matumdas 25 km N of Gilgit.	—	Similar to Nomal.	A random sample of the gossan yeilded 100 ppm Cu & 60 ppm Zn.	—	Good prospect for massive sulphide.	Kazmi 1977.
	"	-Henzel 10 km NW of Gilgit.	Chalcopyrite, bornite, magnetite, garnet, epidote, actinolite.	Mineralised quartz veins more than 135 m long and 4 m wide, along faulted contact of of marble and biotite schist, close to a granitic intrusion.	Random samples contain 2.5 to 7% Cu.	Small	—	Kazmi 1977. Sillitoe 1979.
	"	-Sher Qila 33 km NW of Gilgit.	Malachite, azurite.	There is a 100 ft thick red, brown, and yellow gossan zone in metavolcanics, close to their contact with granitic intrusions.	Not known.	Not known.	Not explored.	Kazmi 1977.
	"	-Singal (45 km NW of Gilgit).	Oxidised Cu minerals.	Quartz veins in granite, along minor faults.	Not known.	Trivial		Sillitoe 1979.

(Continued)

Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Copper ore (contd.)	Kohistan magmatic arc.	-Nazbar Valley 22 km W. of Yasin 36°22':73°20'	Pyrite, pyrrhotite, (Sulphide mineralisation).	Sulphide mineralisation occurs in metasediments of the Darkot Group, close to the Main Karakoram Thrust. Southward there is an intrusion of granodiorite. There are 10 sulphide zones within 10-12 km distance.	Analysis of random samples:- Fe-30-56%, Cu-0.2-1.3%, Zn-0.1-1.2%, Co-0.03-1.5%.	Not known	Unexplored.	Kazmi 1977.
	"	-Drosh 35°33:71°45' several showings in the vicinity, noteably:-			(Based on random samples).			Pudsey et al. 1985.
		(a) Gawuch Gol	Malachite, azurite.	The MKT(Shyok suture zone) runs along the Shishi River and close to the Drosh town. It contains melanges and ophiolites. Its northern contact is with the Kesu-Buni Zom pluton of the Karakoram block. Southward the suture zone rocks have been thrust over Drosh volcanics, Purit and Gawuch Fms. Along the Shishi Valley these rocks are wedged in between the suture zone and the Kohistan Batholith. A number of Cu occurrences have been noted in the volcanic and sedimentary rocks along the suture zone, E and NE of Drosh. Mineralisation is in quartz-calcite veins (fissure fillings) or disseminations in volcanics.	(a)Cu-3.0%, Ag-150 ppm.	Trivial?	With small amount of Sb, Zn, V.	Calkins et al. 1981.
		(b) Kaldam Gol 5 km E of Drosh	Chalcopyrite, malachite, galena, pyrite.		(b)Cu-8.9%, Pb-39.5%, Sb-5.6%, Ag-0.17%, Au-4.3 ppm	Not known	Promising prospect	
		(c) Shishi near Shishi village	Malachite, sulphides disseminated in porphyritic greenstone.		(c)Cu-7.0%, Ag-15ppm.	—	—	Calkins et al. 1981.
		(d)Drosh Gol 1 km E of Drosh	Malachite, quartz, calcite.		(d)Cu-1000 ppm.	—	—	
	"	-Dir (a)Ashnamal 35°13':72°14' (b)Lal Qila 34°55':71°45' (c)Barwa Kambat 34°59':71°40' (d)Dommel Nissar 35°22':71°39' (e)Mirkhani 35°27':71°45'	Chalcopyrite, azurite, malachite, bornite, pyrite.	Mineralised quartz veins in diorites and amphibolites.	Not known	Not known	—	Hussain 1974. Ahmad 1969.

(continued)

Table 9.2. Metallic deposites : location, geological setting and potential. (Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Copper ore (contd.)	Kohistan magmatic arc.	(f) Pana Kot	Malachite, azurite and chalcopyrite.	Mineralisation in skarn, dissemination and secondary stainings along fractures.	Not known	—	Reportedly a large showing.	WPIDC 1970a.
		(g) Usheri Region	Bornite, chalcopyrite, malachite.	Mineralised veins in granite. Veins small but widespread (9x0.8 km area).	Average Cu content 2%.		Showings near Nashnamal, Tarpatar & Shadia villages. Promising prospect.	WPIDC 1970a.
	Karakoram block.	-Yarkhun Valley 36°35':72°53' (near Kanhur)	Chalcocite, azurite.	Mineralisation near contact of limestone and granite gneiss.	Not known	Not known	—	Ahmad 1969.
		-Mastuj (a) Chapali 36°20':72°36' (b) Chapchirag 36°20':72°40'	Azurite	Disseminations in white quartzite.	Not known	Not known	—	Ahmad 1969.
		(c) Pakhturi 36°22':71°72' 22 km W of Mustuj.	Copper sulphides & galena.	Numerous mineralised quartz veins cut quartzite, phyllite and limestone.	Not known	Not known	—	Ali 1950. Ahmad 1969.
		(d) Rain 36°24':72°23'	Copper sulphide minerals & galena.	Mineralised quartz vein cutting shale, quartzite and limestone.	"	"	—	Ali 1950. Ahmad 1969.
	"	-Imirdin 36°03':71°23' 3 km SW of the village.	Chalcopyrite, galena.	Mineralisation in quartz vein and as stringers in quartzite and slate.	"	"	—	Ali 1950. Ahmad 1969.
		-Madashil 36°04':71°50'	Malachite, azurite, galena, pyrite.	Mineralisation along fault zone, quartz-veins in phyllite and phyllitic dolomite.	Cu-0.5%, Pb-0.72%, Sb-1.0%. Au-0.6 ppm (Random)	"	—	Calkins et al. 1981.

(continued)

Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Copper ore (contd.)	Karakoram block.	-Shoghot 36°01':71°46'	Chalcopyrite, galena, malachite.	In the vicinity of Shoghot village there are a number of showings along a narrow zone bound by two parallel faults- the Pasti and Reshun Faults. The fault zones are mineralised. Mineralisation is along brown-weathering, gossan-like (14-15 m thick) breccia zone.	-2sp*. of gossan-like breccia indicated -0.05% Cu, 0.78% As, 2 ppm gold. Sp* from Reshun Fault zone shows 3.6 ppm Ag.	—	Promising prospect.	Calkins et al. 1981.
	"	-Prince Burhanuddin Locality. 35°58':71°48'	Chalcopyrite, malachite.	Mineralised quartz veins along fractures in Chitral Slate.	One sp. shows Cu-3%, Ag-4.7ppm, & smaller amounts of Pb, Zn & Au.	Trivial	—	Calkins et al. 1981.
	"	-Koghozi 35°50':71°50'	Chalcopyrite, malachite.	Disseminations in metavolcanics, showing alteration zones with malachite stains and calcite veinlets with malachite stains.	Random sp. analysis Cu-0.01-1.0% Au-up to 0.2 ppm.	Not known	—	Calkins et al. 1981.
	"	-Mogh 36°01':71°40'	Malachite, chalcopyrite.	Disseminations in garnet-biotite-quartz schist.	Random sp. analysis Cu-2%, Ag-15 ppm, Au-0.2 ppm.	Not known	—	Calkins et al. 1981.
	"	-Kukil Gahirat 6.4 km E of Chitral	Chalcopyrite, galena, pyrite.	Mineralised quartz vein in Chitral Slate.	Cu-0.33%, Zn-0.7%, Pb-0.7%, Ag-15 ppm.	Trivial	—	Calkins et al. 1981.
	Chagai magmatic arc.	-Saindak 29°18':61°33'	Chalcopyrite, chalcocite, covellite, digenite, bornite.	The porphyry type sulphide ore body is in hydrothermally altered and mineralised sequence known as Saindak alteration zone, developed in siltstone, sandstone and tuff of Amalaf Fm. Composite in nature, the mineralisation is related to typical zonal patterns centered on three related porphyry stocks of Mid Miocene age.	Cu-0.336-0.44%, (Average -0.38%) with recoverable quantities of Mo, Ag, & Au.	Large, 412 m.t. (proved)	Mining in progress.	Sillitoe et al. 1977. Ahmed et al. 1972. Bizanjo 1994.

(continued)

* Sp=Specimen

Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Copper ore (contd.)	Chagai magmatic arc.	-Dasht-Kain 29°33':64°29'	Malachite, chrysocolla, goethitic limonite, chalcocite, covellite, chalcopyrite, magnetite, molybdenite.	Porphyry type copper mineralisation associated with two tonalite porphyry stocks which intrude a monzonite and diorite batholith. This batholith has intruded Sinjrani volcanics.	Cu-0.09 -1.2%, Mo-21 ppm.	Large. 400 m.t. probable reserves.	Estimates based on 8 test holes.	Ahmad 1980.
	"	-Darband Chah 29°27':63°44'	Sulphide minerals.	Porphyry type deposit similar to Dasht-e-Kain.	—	—	—	Siddiqui 1984.
	"	-Ziarat Pir Sultan 29°22':64°10'	Sulphide minerals.	Porphyry type deposit similar to Dasht-e-Kain	—	Rough estimate 200 m.t.	—	Islam et al. 1993. Kazmi et al. 1991.
	"	-Kabul Koh	Sulphide minerals	Porphyry type deposit.	—	Rough estimate 50 m.t.		Islam et al. 1993.
	"	-Missi	Sulphide minerals	Porphyry type deposit.	—	Rough estimate 100 m.t.		Islam et al. 1993.
	"	-Humai -Max G. White -Kangord -Ziarat Malik Karkam -Bazgawanan	Sulphide minerals.	Porphyry type deposit.	—	Not known		Islam et al. 1993.
	"	-Talaruk 29°45':61°02'	Chalcocite, pyrite.	Stratabound accummulation of hypogene chalcocite disseminated in dacite pyroclastics (Sinjrani Volcanics). Syngenetic Manto-type mineralisation.	—	Small	—	Sillitoe 1975.
	"	-Amuri 29°15':63°35'	Chrysocolla, malachite, chalcocite, chalcopyrite, native copper.	Mineralisation along fissures and as disseminations in andesitic Volcanics (Sinjrani Volcanics). Syngenetic Manto-type deposit.	Random samples contain up to 6.0% Cu.	Small	—	Ahmad 1975. Sillitoe 1975.

(continued)

Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Copper ore (contd.)	Chagai magmatic arc.	-Kundi-Baluchap. 29°12':64°28' -Mashki Chah 29°01':62°26'	Chalcopyrite, magnetite.	Small magnetite and chalcopyrite deposits of contact-metasomatic type in Cretaceous limestone at the contact with dioritic and granodioritic intrusives.	Not known	Small	—	Sillitoe 1975.
	"	-Bandegan 28°49':65°03'	Chalcopyrite magnetite.	Small magnetite and chalcopyrite deposit in andesitic tuffs in contact with syeno-dioritic intrusion in Ras Koh region.	Cu-0.5-1.0% Fe-40-45%.	Small, (32,000 tonnes).	Electro-magnetic survey & 9 shallow holes drilled.	Ahmed 1964. Ahmad 1975.
	"	-Robat 29°27':65°56'	Copper carbonate, sulphide & silicate.	Mineralised veinlets in basic dyke.	—	Trivial	—	Ahmad 1975.
		-Amir Chah 29°01':62°38'	Malachite, galena, pyrite.	Mineralised shear zones in granite and granodiorite.	—	Trivial	—	Ahmad 1975.
	"	-Dalbandin 28°52':64°24'	Copper sulphide, & carbonate.	Mineralised quartz vein filling shear zone in syenite.	—	—	—	Ahmad 1975.
	"	-Nok Chah 28°57':64°45'	Copper sulphide, carbonate, silicate, magnetite, hematite.	Mineralised shear zone at contact of volc. tuff and syenite, monzonite, diorite.	—	—	—	Ahmad 1975.
	"	-Pakus Nala 28°51':65°06'	Chalcopyrite, sphalerite, pyrite.	Quartz veins in diorite and shear zones at contact of metavolcanics and diorite.	—	—	—	Ahmad 1975.
	"	-Koh Marani 29°28':64°25'	Chalcopyrite, galena, hematite.	Quartz veins in andesite porphyry, grano-diorite and tuff.	—	—	—	Ahmad 1975.
Iron ore	Indus Platform (Shahpur buried Ridge).	-Kirana Hills (Sargodha) 31°58':72°34'	Hematite, magnetite.	Massive beds of hematite interlayered with metavolcanics (Precambrian Kirana Gr) and sills of dacite.	Fe_2O_3 75-95% Si_2O-0.5-3.8% with traces of gold.	Not known	Likely to contain large reserves.	Kazmi and Abbas 1991.

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Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Iron ore (contd.)	Foreland fold-and-thrust belt. (NW Himalaya)	Abbottabad (E of the town, along hill-crest).	Hematite.	Sedimentary ore in the Paleozoic sequence (Abbottabad Group).	Fe-14-46%.	2.6 m.t.	—	Ahmad 1969.
		Galdanian 34°1':73°19' 16 km NW of Abbottabad.	Hematitic claystone.	Sedimentary ore, between a sequence of dolomite and limestone (Jurassic). Several lenticular, discontinuous beds of hematitic claystone (30-50 m) thick.	Average Fe -20%, SiO_2 -9%, P -0.3%.	60 m.t.	20 m.t. ore with about 40% Fe.	Klinger et al. 1963.
	″	Langrial 33°55':73°07' 32 km S of Abbottabad.	Chamosite-limonite, hematite-chamosite.	Sedimentary ore comprising lateritic and oolitic hematite occurs along the unconformity between Paleocene Lockhart Limestone and Cretaceous Kawagarh (?) sandstone.	Fe-9-50% (Av=30%) SiO_2-0.0-60.3%, Al_2O_3-5.7-30.05%, CaO-1.0-12.0%.	30 m.t.	—	Khan and Ahmad 1966. Asrarullah 1978. Kazmi and Abbas 1991.
	″	Mazar Tang 33°45':71°55' 64 km NE of Kohat.	Oolitic hematite.	Sedimentary ore, lenticular beds of massive, oolitic hematite in Jurassic limestone.	Fe_2O_3-50.4-57.6%, SiO_2-5.6-5.8%, Al_2O_3-3.6-14.6%.	0.5 m.t.	Chem. quality based on 2 random samples only.	Asrarullah 1978. and 1979.
	″	Kalabagh 32°55':71°32'	Chamosite-siderite, glauconite-siderite.	Sedimentary ore, occurs in sandstones in upper part of the Chichali Fm. (Jurassic Cretaceous).	Fe-32-36%, SiO_2-20-26%, Al_2O_3-5-13%, P-0.2-0.3%, S-0.1-0.5%.	350 m.t.	Proved reserves 165 m.t.	Ahmad 1969. Asrarullah 1978. WPIDC 1970b.
	″	Pezu (SE of the town) 32°20':70°44'	Limonite, siderite, glaucophane, chamosite.	Sedimentary oolitic ore in Cretaceous Lumshiwal Formation.	Fe-31.3%, SiO_2-19.8%, Al_2O_3-5.5%, Ca/MgO-8.4%, TiO_2-0.4%.	66 m.t.	Sulphur & Phosphorus below 0.62%.	WPIDC 1970b.
	″	Samana Range (16 km from Hangu)	Hematite, limonite	Sedimentary oolitic ore in Cretaceous Lumshiwal Formation.	Fe_2O_3-55.85%, Al_2O_3-9.4%, TiO_2-0.35%, SiO_2-20.64%.	Small		Ahmad 1969.
	″	Rakhimunh 52 km W of D. G. Khan.	Limonite, siderite.	Sedimentary lateritic ore at the base of Oligocene Nari Fm.	Fe-37.5%, Al_2O_3-7.1%, CaO/MgO-6.6%, Mn-1.5%.	Small, (14.5 m.t) ore.	Medium grade	Asrarullah 1979.

(continued)

Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Iron ore (contd.)	Himalayan crystalline zone.	-Besham	Magnetite.	Skarn-hosted magnetite concentrations in Precambrian metasediments (Karora Gr), which are enclosed in basement gneissic granite (Besham Group).	Fe-up to 40%, Zn-2.0%, Sn-0.2%, W-0.05%, Mo-0.01%.	Small (6.8 m.t)	—	Ashraf 1980.
	Ophiolitic thrust belt (Zhob).	-Naweoba 31°35':69°22' -Inzarkai 31°35':69°25'	Magnetite.	Minor occurrences associated with Zhob ophiolites.		Small		HSC 1960.
	Ophiolitic thrust belt, (Bela-Khuzdar).	-Shekran (Khuzdar) 27°51':66°23' (A) 21 km NW of Khuzdar. (B) 24 km NW of Khuzdar (C) 1.6 km NE of B above.	Siderite, limonite, hematite, calcite, quartz.	Hydrothermal, replacement type veins and stock works in impure siliceous limestone (Jurassic Zidi Lst.).	Fe-35-40%,	Small, (10 m.t. down to 16 m depth).	—	HSC 1960. Ahmad 1975.
	"	-Monar Talar 27°44':66°35' 9.5 km SW of Khuzdar.	Hematite, siderite.	Irregular netwark of ferrugenous veins in limestone (Jurassic) occurs below the barite deposits.	Not known	Not known.	—	Ahmad 1975.
	Indus suture zone.	-Sherkot-Kolai. 34°57':73°02'	Magnetite.	Associated with amphibolites of the Jijal Complex which is the lowermost stack of layered cumulate complexes developed along the frontal (lower most) part of the Kohistan magmatic arc and precipited from subduction-related tholeiitic magmas.	Fe-38%	Small.	—	Miller et al. 1991. Ashraf 1980.
		-Jijal 34°58':72°00'	Magnetite		Fe-38%.	Small.	—	Ashraf 1980.
	Kohistan magmatic arc.	-Ghazanosar 24 km NW of Shah Dheri (34°53':72°54')	Magnetite	Associated with amphibolitic rocks of Kohistan magmatic arc.	Fe-30%	V. small, (20,000 t)	—	Asrarullah 1979.

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Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Iron ore (contd.)	Kohistan magmatic arc.	-Manarai 25.5 km W of Saidu (34°78':72°37')	Magnetite, ilmenite, spinel, hematite, pyrite.	Disseminated magnetite in amphibolite, with small lenses of massive magnetite.	Av. comp. of crushed ferro-magnetic con-centrates (%) Fe_2O_3-80, SiO_2-8, Al_2O_3-4, MgO-4, CaO-2.5, TiO_2-2.8, (Iron 14-46).	Not known	There are vast reserves of magnetic amphi-bolite in the region.	WPIDC 1970a.
	"	-Munda (Dir District)	As above	As above	Percent Fe-10, TiO_2-0.8, MgO-16, Mn-1.2, Ni-0.25, Co-0.15.	Not known	—	WPIDC 1970a.
	"	-Dommel Nissar (35°22':71°39') Chitral Dist.	Magnetite, specularite, pyrite, chalcopyrite.	Contact-metasomatic magnetite deposit in garnet-epidote meta-volcanosedimentary rocks intruded by granodiorite-quartz diorite, close to the northern suture (MKT).	Fe 50-65%, (Av=48.6%) SiO_2-4-15%, CaO-1-5%, Al_2O_3-0.5-4%, Mn>0.5%.	Small, (6.5 m.t.)	—	WPIDC 1970a. Austromineral 1978.
	Chagai magmatic arc.	-Saindak 29°17':61°32'						
		(a)3.5 km SE	Hematite.	Volcanogenic deposit in Amalaf Fm.	Not known	Small	—	Ahmad 1975.
		(b)5.6 km NE of Saindak.	Hematite.	Massive bodies in volcanics of Juzzak Fm. or as low-grade hematite with galena in quartz veins cutting sedimentary rocks.	Not known	Small	—	Ahmad 1975.
	"	-Mashki Chah 29°02':62°24'	Hematite, magnetite.	There are several small contact-metasomatic deposits within andesitic Sinjrani volcanics and intercalated sediments, close to quartzporphyry intrusions. Iron ore occurs in veins as disseminations in the host rocks.	Fe-30-50%, Al_2O_3-2.8%, TiO_2-0.13%, Cu-0.01%.	Small (0.43 m.t.)	—	HSC 1960. Asrarullah 1978, 1979.

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Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Iron ore (contd.)	Chagai magmatic arc.	-Durban Chah 15 km N of Mashki Chah.	Magnetite, hematite.	Mineralised veins in Sinjrani Volcanics-Contact-metasomatic deposit.	Not known	Small, (1.125 m.t.).	—	Asrarullah 1978.
		-Amir Chah 29°13':62°27'	Hematite, magnetite.	Contact-metasomatic deposit in Sinjrani Volcanics intruded by small granitic intrusions. Hematite veins and a layer of hematite-epidote rock at contact of limestone and volcanics.	Not known	Small, (1.125 m.t.).	—	HSC 1960. Ahmad 1975.
	"	-Chilgazhi 29°08':64°14'	Magnetite (Gangue:pyrite, chalcopyrite, calcite, epidote, pyrophyllite, apatite).	Massive to disseminated, layered deposit in Sinjrani Volcanics which contain several small intrusions of diorite and granodiorite. The host rock is andesitic to epidote-rich volcanic rock. According to Shcheglov (1969) the deposit is endogenous and volcanogenic.	Fe-32-55%, Cu-0-0.39%, Au-tr-2.8g/t.	Small, 23 m.t.	—	Farooq et al. 1970. Shcheglov 1969. Ahmad 1975. Kazmi and. Abbas 1991.
	"	-Gorband-Kasanen Chapar 29°06':64°18'	Hematite, magnetite.	Mineralisation due to contact metamorphism in Cretaceous limestone at intrusive contact of syenite and Sinjrani Volcanics.	Not known	Very small (50,000t.)	Two deposits NNW of Dalbandin.	HSC 1960. Ahmad 1975.
	"	-Kundi-Baluchap 29°08':64°30' 30 km NE of Dalbandin	Hematite, magnetite, siderite.	Two deposits in Sinjrani Volcanics. (a) Siderite-calcite vein in basalts intruded by diorite and granodiorite (contact metasomatic). (b)Hematite-magnetite concentration in garnet-epidote rock (contact metamorphic).	Fe-43-64%	Small, (0.13 m.t.)		Ahmad 1969. Asrarullah 1978.
	"	-Pachin Koh (88 km NW of Nokkundi)	Magnetite, hematite.	Magnetite intercalated with andesite (Sinjrani Volcanics). Magnetite-actinolite plugs accompany two principal magnetite flows. Volcanogenic, contact-metasomatic, hydrothermal.	Fe_2O_3-67-82%, SiO_2-9-22%, Al_2O_3-1.4-4.4%, CaO-1.2-2.2%, (Av. of 11 channel sps.)	Medium, 45 m.t. (30 m.t. proved).	27 small magnetite-hematite bodies at the locality. 78 test holes (19,700m) drilled.	Asrarullah 1978. Ahmad 1978. Kazmi and Abbas 1991.
	"	-Chigendik (40 km NW of Nokkundi).						
	"	Bandegan 28°51':64°03' (Ras Koh)	Magnetite, hematite.	3 deposits in Kuchakki Volc. Gr. One is a magnetite vein, others hematite-magnetite layers in indurated shale and volcanics, metamorphosed to siliceous and hornfelsic rocks. Kuchakki Volcs. are intruded by diorite-syenite.	Fe-24-54%,	Small (0.18 m.t.)	Contact-metasomatic to pyrometasomatic.	Ahmad 1975. Asrarullah 1978.

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Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Iron ore (contd.)	Chagai magmatic arc.	-Nok Chah (Ras Koh)	Hematite, magnetite.	Mineralisation in Kuchakki Volcs. intruded by syenite stock. Contact metamorphism. Deposits in limestone with epidote, garnet, diopside.	Not known	Small	Mined sporadically.	Ahmad 1975.
Lead-Zinc ore	Foreland fold-and thrust-belt.	-Faqir Mohd. 33°57':73°09' -Hazara Dist.	Galena.	Galena in quartz-barite veins, in Eocene limestone.	Uneconomic	Trivial	—	Ahmad 1969.
	Himalayan crystalline zone.	-Hal 34°11':73°03' -Kokal 34°27':73°26'	Galena.	These showings are in Hazara Dist. Galena bearing quartz veins (fracture filling) occur in Precambrian metasediments of Hazara Fm. At Paswal the veins also contain pyrite, sphalerite and chalcopyrite.	Uneconomic	Trivial	—	Ahmad 1969.
	"	-Mihal 34°09':73°08' -Paswal 34°12':73°07'	Galena.	As above.	—	Trivial.	—	Ahmad 1969.
	"	-Lahor 6 km N of -Besham (34°56':72°52')	Galena, sphalerite, melnikovite, chalcopyrite, pyrite, pyrrhotite.	Zoned polymetallic sulphide mineralisation occurs in Precambrian metavolcanics and metasediments of the Pazang Fm. where sphalerite-pyrite-galena and sphalerite-pyrrhotite-melnikovite-galena occur in veins or as disseminations. The deposits are volcanogenic exhalative type affected by subsequent regional metamorphism.	-Pb-3.1%, Zn-4.2%, (average).	-Medium, 0.5 m.t.	—	Ashraf et al. 1979. Khan 1983. Siddiqi et al. 1988.
	"	-Pazang 3 km E of Besham (34°56':72°52')			-Pb-3.45%, Zn-6.2%, (average).	-1.0 m.t.		
	Ophiolitic thrust belt. (Bela).	-Sekran 27°88':66°38' (NW of Khuzdar)	Galena, siderite, limonite, calcite.	Stratabound mineralisation occurs in Jurassic Loralai Lst. The western part comprises a 1.5 km E-W iron-rich gossan along a thrust fault. The eastern part is 400 m long, 3-10 m thick with fissure fillings.	Pb-0.02%, Zn-3.15%, (from gossan zone).	Not known.	—	Ahsan et al. 1994. Ahmad 1969.
		-Malkhor-Ranj Laki 20 km NW of Khuzdar (27°49':66°35')	Galena, siderite, limonite.	Mineralisation characterised by E-W gossan zone comprising 7-8 bedded ore bodies in Jurassic Loralai Lst.	Malkhor- 2.3% Pb+Zn, (Pb dominant) Ranj Laki- 0.3% Pb, 2.61% Zn, 7 g/t Ag.	Not known	—	Ahsan et al. 1994. Ahmad 1969.

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Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Lead-Zinc ore (contd.)	Ophiolitic thrust belt (Bela).	-Gunga 27°44':66°33' 11 km SW of Khuzdar	Sphalerite, galena, pyrite, baryte, siderite.	Pb-Zn-Ba mineralisation occurs as strata-bound-hydrothermal deposit (SEDEX type) in Anjira Member of Shirinab Fm. The mineralised zone is straddled by faults and characterised by silicic gossan. Upper part is dominantly composed of barite, in the lower part Pb-Zn is dominantly stratiform comprised of disseminations and thin layers of sulphides.	Zn-5.36%, Pb-1.43%, Ag-up to 2,500 ppm.	Large, 3.0 m.t. (in addition the Zn-Pb-Ba ore reserves are estimated at 9.9 m.t.)	—	Jankovic 1984a,b. Ahsan et al. 1994. UNDTCD 1990.
	〃	-Surmai (1 km S of Gunga)	Sphalerite, galena, pyrite, siderite, marcasite, chalcopyrite.	Mississippi type mineralisation occurs in Loraiai Lst. (lower stratigraphic position than Gunga), along bedding plane and also in fissure filling. Outcrop characterised by reddish to yellowish brown gossan.	Zn+Pb<5%, Ag-15g/t.	Medium 0.76 m.t.		Ahsan et al. 1994. Subhani and Durrazai 1989.
	〃	-Duddar 26°35':66°50' (135 km NNW of Karachi).	Pyrite-marcasite, sphalerite, galena, barite.	Sedex type mineralisation occurs in Anjira Member of Shirinab Fm. and is constrained between 2 major faults. It is stratiform and conformable to bedding.	Zn-11.43%, Pb-2.1%, Fe-3.89%, Ag-<2-7 ppm	V. large, 10.29 m.t.	—	Jones et al. 1994. Azam et al. 1989.
		-Mithi (15 km E of Bela).	Galena, baryte.	Mineralisation is hosted in Jurassic Loralai Lst. and is marked by extensive jasperoid gossan. Mineralisation is epigenetic, with veins along bedding planes, joints and cross-cutting fractures.	Zn-up to 2.7%, Pb-up to 2.6%, (Random chip samples)	Not estimated.	—	Ahsan et al. 1994.
	Kohistan magmatic arc.	-Ushu 35°44':72°40' (Swat)	Galena, sphalerite, pyrite, Chalcopyrite.	Fracture filling, quartz vein in diorite.	Zn-12-26%, Pb-14-57%, Sb-tr.	Not estimated.	—	Tahirkheli 1959.
	Karakoram block.	-Tirich Mir Zone i) Parabeck 35°59':71°24' ii) Imirdin 36°03':71°23' iii) Muzhigram 36°06':71°40' iv) Tashker 36°03':71°48' v) Pakhturi 36°22':72°18'	Galena, sphalerite, pyrite, chalcopyrite.	Scattered small showings in the form of quartz veins in quartzite and slate (Arkari Fm.). The deposits are within short distance of Hot Spring and Tirich Mir pluton.	Not known	Not estimated.		Calkins et al. 1981.

(continued)

Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Lead-zinc ore. (contd.)	Karakoram block.	Baig 36°09':72°01'	Galena.	Galena in quartz veins cutting slates and phyllites (Lun Shale).	Not known	Not estimated.	—	Ahmad 1969. Calkins et al. 1981.
		Madashil 36°03':71°49'	Galena, malachite, azurite.	Mineralised breccia and quartz veins in Lun Shale, close to Pasti Fault.	Pb-53.5%, Au-0.3-0.6 ppm.	Not estimated.	—	Ahmad 1969. Calkins et al. 1981.
		Awireth 35°58':71°44'	Fine-grained Pb-Sb sulphides.	Mineralised vein in breccia zone along fault separating Cretaceous Reshun Formation and Paleozoic Lun Shales.	Pb-45-51%, Sb-12-17%, Ag-980 ppm, Au-77ppm.	Small	Old mine workings	Ahmad 1969. Calkins et al. 1981.
	Kohistan magatic arc.	Rain 36°24':72°23'	Galena.	Mineralised veins in shale, limestone and quartzite of Lun Shale.	Not known	Not estimated.	—	Ahmad 1969.
		Awi 36°16':72°20'	Jamesonite.	In quartzite and dolomite of Darkot Group, close to Buni Zom pluton and MKT.	Not known	Not estimated.	Limited mining.	Ahmad 1969.
	″	Gahirat 35°41':71°46'	Galena, stibnite.	Mineralised quartz vein in Gahirat Lst. near contact with Buni Zom pluton.	Not known	Not estimated.	—	Ahmad 1969.
	″	Drosh (35°57':71°80') 5 km to the E	Galena, chalcopyrite, pyrite.	Mineralised breccia zone between marble and siltstone, near contact of Purit and Gawuch Fms close to MKT and Kohistan Batholith.	Pb-39.5%, Cu-8.9%, Sb-5.6%, Ag-0.17%, Au-4.3 ppm.	Not estimated.	—	Calkins et al. 1981. Pudsay et al. 1985.
	Chagai magmatic arc.	Saindak 29°17':61°34'	Galena.	Mineralised calcite veins in basaltic dykes that intrude Paleocene volcanics.	Pb-12-22%.	Trivial.	—	Ahmad 1975.
		Koh Marani 29°28':64°25'	Galena.	Small quartz veins in andesite porphyry.	Pb-29.59%, Zn- 4.24%, (one sample).	Trivial.	—	Ahmad 1975.
	″	Dirang Kalat 29°28':64°33'	Galena, sphalerite, limonite, siderite.	Mineralised veins along fault in andesite.	Pb-42-51%, Zn-2.4%, Ag-0.8-1.44 oz/t. Au-0.1-0.02 oz/t.	V. small.	Worked sporadically.	Ahmad 1975.
	″	Makki Chah 4 km SE of Talaruk 29°45':61°02'	Sphalerite, pyrite.	Massive sulphide mineralisation occurs along the upper part of a sequence of dacite and tuffs. Three massive lenticular ore bodies are enclosed in a gypsum bed overlying the dacite. Mineralisation is volcanogenic and of syngenetic sub-marine exhalative origin (Kuroko type).	Zn-3.8%, Ag-15g/t.	Small, 0.5 m.t.	—	Saigusa 1977.

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Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

	Tectono- zone	Name of & location	Minerals		Quality/			
Lithium	Himalayan crystalline zone. (Nanga Parbat Massif).	-Shengus 35°43':74°48'	Lepidolite.	This region comprises a major pegmatite field with thousands of pegmatites cutting the Proterozoic Nanga Parbat Gneisses. A large number of these contain various gemstones along with lepidolite.	Up to 4.24%, Li_2O.	—	—	Butt et al. 1987. Kazmi et al. 1985.
Manganese ore	NW Himalayan fold-and-thrust belt.	-Galdanian 34°16':73°19' 16 km NE of Abbottabad.	Pyrolusite, psilomelane.	Manganese-iron bearing red-bed formation occurs within a Cambrian sequence of limestone, shale, sandstone, quartzite and dolomite (Abbottabad Fm.). Mn ore occurs in thin, discontinuous lenses (2-7m long). There are several lenses in the region.	Mn-6.7%, Fe-20.4%, (Av. of 19 channel sp.).	Small	—	Quraishi & Imam 1960. WPIDC 1970a.
		-Chur Gali 34°18':73°38' 20 km NE of Abbottabad.	Pyrolusite, psilomelane		Mn-7-22%, Fe-28.6% (Av. of several channel sp.).	Small 180,000 tonnes proved.	—	WPIDC 1970a. Quraishi & Abdullah 1960.
	Ophiolitic thrust belt and suture zone.	-Shangla 30 km NE of Mingora (Swat)	—	Veins and lenses of manganese ore have been reported in metavolcanics of the Indus suture melange.	MnO_2-32.46%.	Not known.	—	Hussain et al. 1990.
		-Kassai (Mohmand Agency).	—	Veins and lense of manganese ore have been reported in tectonic blocks of qaurtzite in the Kot-Pranghar Melange.	MnO_2-60%.	Not known.	—	Hussain et al. 1990.
	"	-Thal 33°22':70°32' (2.5 km SW of Town).	Pyrolusite, hematite, magnetite, barite.	Manganese ore occurs in a cherty hematitic shale in a Cretaceous volcanosedimentary sequence (Kahi Melange). The ore is lenticular and the mineralised zone is fairly extensive.	Not known	Not known.	—	Ahmad 1969. WPIDC 1970a. Beck et al. 1995.
	"	-Shinkai area 32°55':69°52' Waziristan.	—	Widespread occurrence of lenses and stringers of Mn ore has been reported from Waziristan. Mn occurs at the contact of volcanic flows and cherty shale and limestone (Kahi Melange).	Not known	Not known	Main localities: Garang Dala algad, Barazai, Ser Kour, Ghundai, algad & Saidgi.	Khan et al. 1982.

(continued)

Table 9.2. Metallic deposites : location, geological setting and potential. (Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/Grade	Size	Remarks	References
Manganese ore. (contd.)	Ophiolitic thrust belt.	-Naweoba 31°33':69°22' N of Zhob -Waltoi Kud 30°40':68°01' S of Nasai.	—	Mangnese ore occurs in Zhob Melange. Veins and layers of Mn are found in Parh Limestone which is intercalated with tectonic blocks of volcanics and ophiolites.	Not known.	Not known.		Ahmad 1975.
	"	(a)Kohan Jhal 26°37':66°19' (64 km N of Bela).	Psilomelane, pyrolusite.	Mn ore occurs in Kanar Melange in a sequenceof pillowed basalt, basalt, dolerite, shale and limestone. The mineralised zone comprises irreglar layers and lenses of Mn ore in a jasperoid layer, underlain by pillow basalts and commonly overlain by red shales. These are largely stratabound volcanogenic deposits. However a few (e. g., Siro Dhoro) are replacements and veinlets in basalt and dolerite dykes and probably represent feeders to the syngenetic accumulations.	MnO_2-36.85%.	Small, 11,000 t.		HSC 1960. Ahmad 1969, 1975. Sarwar 1984, 1992. Sillitoe 1975.
	"	(b) Haji Khan Bent. 26°31':66°24'						
	"	(c) Sanjro 26°28':66°27'	Oxysilicate of manganese, brunite.		Mn-0.09-30.02%.	21,000 t.		
	"	(d) Khabri Dhora 26°28':66°25'	Psilomelane, pyrolusite.		Mn-17.99-23.99%.	65,000 t.		
	"	(e) Siro 26°17':66°33'			Mn-2.88-31.72%.	45,000 t.		
	"	(f) Khan Kheo Nai. 26°11':66°34'			MnO_2-40%	5,800 t.		
	"	(g) Kharrari Nai 25°54':66°45'			MnO_2-42%	40,000 t.		

(continued)

Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Manganese ore. (contd.)	Chagai Magmatic arc.	-Nushki 29°43':66°01' (9 k m N of town)		Mn ore occurs as nodules in the Cretaceous Humai Formation.	Mn-40.78% (one sp only).	Trivial	—	Ahmad 1969.
		-Sotkinoh Hill 29°14':65°51' (Ras Koh)		Mn ore occurs as fissure fillings in quartz veins which cut the volcano-sedimentary sequence of the Paleocene Rakhshani Fm.	MnO_2- 40-50%.	Small, 3,200 t. down to 7 m depth	Large reserves possible.	Ahmad 1975.
Mercury	Bela ophiolitic thrust belt.	-Gunga, Duddar.		Mercury occurs in very small quantities in the Pb-Zn ore.	0.017-89.6 ppm.		—	Azam et al. 1989. Jankovic 1984a,b.
Rare Earths	Himalayan crystalline zone.	-Koga (Swat Dist.)		In nepheline syenite and alkaline granites of Koga alkaline complex. Also in Koga carbonatites.	(Random samples) La-0.01-0.02%, Y-0.005%, Ce-0.079%.	Not estimated.	—	Ashraf et al. 1987.
	"	-Sillai Patti 30 km W of Dargai.		Associated with carbonatites in alkaline complex.	La-0.04%, Y-0.01-0.02%.	Not estimated.	—	Ashraf et al. 1987.
	"	-Loe-Shilman (Khyber Agency).		In carbonatites associated with the alkaline complex.	La-0.005-0.055%, Y-0.004-0.006%, Nb-0.007-0.2%, Ta-0.006-0.01%.	Not estimated.	—	Ashraf et al. 1987.
	"	-Skhakot-Qila (Malakand Dist.)		In monozite-bearing chlorite schists.	Very small amount of Ce.	Not estimated.	—	Zulfiqar 1986.
Nickle	Indus suture zone.	-Souch 34°55':73°40' (Kaghan)	Heazlewoodite, pentlandite.	In ophiolites along the suture zone.	Ni-0.28-0.32%.	Not estimated.	—	Ahmad 1981.
		-Swat, Malam Jabba-Shangla-Alpurai area.	Pentlandite, milerite, polydymite.	Anomalous values of Ni have been reported from the serpentinites and talc-carbonate rocks of the Mingora ophiolite melange.	Ni-0.4-0.8%.	Not estimated.	—	Chaudhry et al. 1980.
		-Muslimbagh 30°51':67°40'		In ultramafic rocks of Zhob ophiolites.	Ni up to 0.85%.	Not estimated.	—	Shams 1995.

(continued)

Table 9.2. Metallic deposites : location, geological setting and potential.(Contd.)

Mineral	Tectono-metallogenic zone	Name of deposit & location	Minerals present	Geological setting	Quality/ Grade	Size	Remarks	References
Niobium and Tantalum.	Himalayan crystalline zone.	Loe Shilman (Khyber Agency.)	Pyrochlor, betafite.	Carbonatites in alkaline complex.	0.01-0.03% Nb.	Small showing.	—	Butt 1981b.
Platinum	Kohistan magmatic arc.	Chilas	Moncheite, michinerite, merenskyite, platinum arsenide, palladium telluride.	In dunites of the Chilas ultramafic complex.	PGE+Au-800 to 2,800 ppb.	—	—	Ashraf and Hussain 1994.
		Jijal 34°97':72°00'	Sperrylite, melonite, atheneite, tetraauri-cupride, electrum, hessite.	PGE and gold mineralisation occurs in association with disseminated accessory sulphides in the Jijal ultramafic complex. Five precious metal zones have been identified and in the layered complex stratigraphically they lie one above the other (Table 9.3).	Pt-56-723 ppb. Pd-40-2,275 ppb.	Not estimated.	—	Miller 1991.
	Kohistan arc (Dargai Klippe)	Sakhakot-Qila (West of Dargai 34°47':71°90')	Awaruite, irridian.	Pt, Pd and Rh have been reported in dunite, harzburgite and chromitite samples from the Dargai Complex. Ir-Ru-Os-Ni-Fe reported from one sample.	Pt+Pd+Rh = 5-25 ppb.	Not estimated.	—	Page et al. 1979. Ahmed and Bevan 1981.
	Zhob ophiolitic thrust belt.	Muslimbagh 30°51':67°40'	Chromitites.	Associated with chromite deposits, in the Zhob ultramafic complex. ppb.	Pt+Pd+Rh up to 375 mated.	Not esti-	—	Page et al. 1979. Ahmed and Bevan 1981.
Tungsten ore.	Himalayan crystalline zone.	Oghi 34°10':74°19'	Scheelite, powellite.	In pegmatites and aplites traversing the Precambrian Susagali granite.	3.2% W.	Trivial	—	Shams 1995.
	Karakoram block.	Miniki Gol 35 km NW of Chitral (35°50':71°48')	Scheelite.	Stratabound, skarn- type scheelite occurs in the Arkari Fm at several locations in a 7 km long mineralised zone. W is concentrated in only parts of the horizon. The mineralised zone (host rock) comprises calc-silicate quartzites, tourmaline-quartz gneiss, quartz-calcite schist, albite quartzite and graphitic schist. The Lutkho River, Miniki Gol and Arkari River sediments contain significant amounts of W (180-2,500 ppm), suggesting much wider mineralisation than known presently.	0.08 to 0.85% W.	Not known. Likely to be at least a medium size deposit.	—	Leak et al. 1989.
	Chagai magmatic arc.	Amalaf 29°18':61°37'	Scheelite, tungstite.	Xenothermal alterations in Saindak Fm which comprises pyroclastics intruded by quarz porphyry.	Not known	Not known.	—	Shams 1995.

detailed exploration may reveal large economic deposits. However, so far only the Saindak (Fig. 6.17), Dasht-e-Kain and Boya deposits have been explored. At Saindak open-cast mining has started. The proven ore reserves are about 412 million tonnes with 0.38% copper and recoverable quantities of molybdenum, silver and gold. The proposed annual production from Saindak is about 15,800 tonnes of copper blister, 1.47 tonnes of gold, and 2.76 tonnes of silver (Bizanjo 1994).

Gold and Silver

Gold and silver occur in association with copper, iron and lead-zinc ores, in the inliers of the buried shield in the northern part of the Indus platform, the Himalayan crystalline zone, Bela ophiolite belt, Kohistan magmatic arc (Table 9.3), Karakoram block and the Chagai magmatic arc (Table 9.4). The Chaman Fault zone and its southwestern extension in Kharan (the region between Kharan and Panjgur Faults) contains stocksworks of ferruginous quartz and calcite veins bearing small amounts of antimony, lead, zinc, mercury, traces of gold and up to 12 ppm silver (Ahsan et al. 1991, Hussain et al. 1995, Haq et al. 1995, Younas et al. 1995). Apart from these primary deposits, placer gold occurs extensively in the upper reaches of the Chitral, Gilgit, Hunza and Indus Rivers. These placers are largely thin pockets of heavy mineral concentrations that are removed by floods every summer season and redeposited as the floods recede. Gold washers have panned and produced gold on a very small scale. A small, economic gold placer deposit has been proved through drilling near Chilas (Austromineral 1976).

Mining of the gold-silver bearing Saindak copper deposit has started in 1995 and it is expected that processing of this ore would produce 1.47 tonnes of gold and 2.76 tonnes of silver annually. The silver bearing Duddar Pb-Zn ore is expected to be mined in the near future and a small production of silver is likely.

Table 9.3. Characteristics of PGE and gold mineralisation in the Jijal layered ultramafic complex (data from Miller et al. 1991).

ZONE	MINERALS	HOST ROCK	PRECIOUS METALS (ppb)				
			Au	Pt	Pd	Rh	Total
V	Pyrite, chalcopyrite, bornite, millerite, pentlandite.	Sulphide cumulate in garnet-gabbro with 5-10% sulphide.	339	723	1730	< 2	279
IV	Chalcopyrite, bornite clinopyroxene cumulate	Garnet-hornblendite	7 to 2130	64 to 221	88 to 417	< 2	647 to 2284
III	Chalcopyrite, pyrrhotite, pentlandite; lesser pyrite and millerite, tetraauricupride and atheneite.	Hornblende-clinopyroxenite – ilmenite + postcumulus garnet.	76 to 2457	15 to 715	144 to 2275	< 2 to 3	293 to 3597
II	Pyrrhotite, pentlandite, chalcopyrite as inclusions in chromite and clinopyroxene.	Wehrlite and chromite-dunite.	35 to 275	99 to 402	40 to 720	< 2 to 10	365 to 1405
I	Chalcopyrite, bornite, pentlandite, supergene violarite, magnetite.	Clinopyroxenite with 10-15% bronzite, 2-3% disseminated sulphides.	86 to 146	56 to 90	47 to 70	< 2	246 to 249

Table 9.4. Gold and silver deposits and showings (also see Table 9.2 under Cu, Fe and Pb-Zn ores).

TECTONO-METALOGENIC ZONE	DEPOSIT	GRADE (PPM)		REMARKS
		GOLD	SILVER	
Indus Platform (buried shield).	Kirana Hills (near Sargodha)	up to 6	–	Associated with hematite deposits in metavolcanics of the Kirana Group (Shah 1973).
Himalayan crystalline zone.	Hall 34°11':73°03".	1.7	–	Associated with galena.
Bela ophiolitic thrust belt.	Malkhor-Ranj Laki.	–	7	
	Gunga.	–	2,500	Associated with Pb-Zinc ore.
	Surmai.	–	15	
	Duddar.	–	<2-7	
Kohistan magmatic arc.	Nomal.	0.65-1.03	69	
	Gawuch.	–	150	Associated with Cu ore.
	Drosh.	4.3	1700	
Karakoram block.	Kaldam Gol.	4.3	100	
	Shishi.	–	5	
	Madashil.	0.6	–	
	Shogot.	2.0	3.6	Associated with Cu ores.
	Pr. Burhanuddin.	–	4.7	
	Kogozi.	0.2	15	
	Baig.	0.3-0.6	–	Associated with Pb-Zn ore.
	Awireth.	5	980	
Chagai magmatic ore.	Saindak.	0.15	0.17	Associated with Cu ore.
	Chilghazi.	2.8	–	Associated with Fe ore.
	Dirang Kalat.	0.56-2.8	22.6-40.8	Associated with Pb and Zn ores.
	Makki Chah.	–	15	
	Ushu.		127-834	
	Jijal.	0.0035-2.45	–	Associated with ultramafic complex

Iron Ore

Iron ore resources of Pakistan are estimated at about 658 m.t. (Table 9.2). Iron deposits are sedimentary, volcanogenic or associated with igneous intrusions (metamorphic, contact-metasomatic or hydrothermal). Iron ore occurrences are largely confined to the Foreland sedimentary fold-and-thrust belt, Kohistan island arc and the Chagai magmatic arc, though a few deposits occur in other tectono-metallogenic zones also (Table 9.2). Sedimentary ores characterise the Foreland sedimentary fold-and-thrust belt and the largest deposits are also found in this zone (e.g., Kalabagh, Pezu and Langrial deposits). The large proven reserves at Kalabagh have attracted much attention. The ore is of low grade and comprises three types as shown in Table 9.5.

Table 9.5. Kalabagh ore-types, their composition (%) and reserves (m.t.).

Type	Fe	SiO_2	Al_2O_3	CaO	P	TiO_2	S	Reserves
1. Kuch (chamosite-siderite)	34.2	21.8	12.1	2.7	0.25	0.68	0.33	34.60
2. Chichali (glauconite-siderite)	33.0	24.0	6.0	3.5	—	—	—	206.00
3. Chuglan (transitional between 1 & 2)	32.0	25.0	6.8	2.8	0.30	0.34	0.26	52.00

(Modified from Ahmad 1969).

Tests on Kalabagh ore (by Koppers of U.S.A., U.S. Bureau of Mines, Hogahna of Sweden, Carpo and M. W. Kellogg of U.S.A. etc.) indicate that this complex ore is not amenable to physical beneficiation. Krupps of Germany tested the ore in 1959 by Krupp-Renn process and found that the ore burden adheres to the kiln wall, threatening damage to the kiln. In 1964 Institute de Recherches de la Sierurgi Francaise (IRSID) conducted semi-commercial test on 1,500 tonnes of Chichali ore after sintering by blending it with imported high-grade ore in a conventional blast furnace at Liege, Belgium. The results suggested that the Acid Blast Furnace Process would suit the ore and may be considered as the exclusive feasible method of reducing the Chichali ore (PMDC 1970b). Since then no serious effort has been made about the development of this ore.

Several showings and small deposits of various types of iron ore occur in the Kohistan island arc terrane and in the Chagai magmatic arc (Table 9.2). Amongst these the Dommel Nissar, Chilgazi and Pachin Koh-Nok Kundi deposits have been explored in detail and economic reserves have been established. These are high-grade magnetite deposits with no deleterious impurities to cause any problem in their industrial use. According to Hussain (1983) the Pachin Koh-Nok Kundi ore is suitable for direct reduction plus electric arc furnace process combination. Lack of infrastructure and long distance from industrial centres are the main impediments to their exploitation. A few showings and very small deposits occur in Bela-Zhob ophiolite belt and in the Indus Suture zone. These are mainly hydrothermal vein-type deposits or minor occurrences associated with ophiolitic complexes. Amongst these the Jijal deposits in the ultramafic layered cumulate complex are of considerable academic interest, besides the possibility that detailed exploration may reveal economic reserves (Ashraf et al. 1980, Miller et al. 1991).

An interesting gold-bearing volcanogenic iron ore deposit has been discovered in the Precambrian metavolcanics of the Kirana Group which comprises the Shahpur buried ridge (Kazmi and Abbas 1991). Presently it is being explored. Large deposits of iron ore, estimated at 200m.t. have been reported from Dilband area os Kalat District by Abbas et al. (1997). The ore consists of hematite and the iron oxide content varies from 52 to 62.25%. The ore occurs at the unconformity on top of the Chiltan Limestone.

Lead-Zinc Ore

Deposits and showings of lead-zinc ore are largely confined to the Himalayan sedimentary fold-and-thrust belt, Himalayan crystalline zone, Bela ophiolite belt, western part of the Karakoram block and the Chagai magmatic arc (Table 9.2). Some of the deposits in the Himalayan crystalline zone (Lahor–Pazang), Bela ophiolite belt (Gunga, Surmai and Duddar) and in the western Karakoram are of considerable geological and economic interest.

The Lahor–Pazang deposits of Besham are located within a fault zone in a highly tectonised region close to the MMT. There are a number of N-S trending strike-slip faults in the vicinity of the deposits (Siddiqi et al. 1988). The ore occurs in veins and as dissemination in the Proterozoic Pazang Formation (part of Besham basement complex) which is comprised of a metavolcanic sequence of tremolite- and diopside-bearing marbles, quartzites, schists and psammitic gneisses. The Besham region is intruded by various granites ranging in age from Proterozoic to Permian. Mineralisation is associated with the Proterozoic Lahor granite complex and according to Butt (1981a) it is characterised by ore zonation ranging from a molybdenite-bearing core to vein type Pb-Zn mineralisation. Varied ore assemblages have been reported from Lahor–Pazang deposits. These are:-

1. Magnetite-barite
2. Magnetite-barite-fluorite-pyrite
3. Scheelite-molybdenite
4. Molybdenite-pyrrhotite-pyrite-galena-sphalerite
5. Sphalerite-pyrite
6. Sphalerite-pyrite-galena
7. Sphalerite-pyrite-melnikovite-galena
8. Pyrite-pyrrhotite-uraninite-chalcopyrite-galena
9. Pyrite-galena-uraninite
10. Crystalline magnesite-talc
11. Barite-marble

These assemblages suggest that the Lahor–Pazang deposits are metallogenically complex and formed in stages. It is likely that they are the product of a number of overlapping or superimposed processes, such as an early volcanogenic exhalative type mineralisation (Pb-Zn-Ba) followed by magmatic and metamorphic processes resulting in the formation of pegmatitic/hydrothermal veins (Mo, Fl, Cu, W, U). The inferred and indicated reserves of Pb-Zn ore with average Pb content of 3.1–3.4% and Zn 4.2 to 6.2% are estimated at 1.5 million tonnes (Siddiqi et al. 1988). Further exploration may reveal larger reserves.

Extensive lead-zinc-baryte zone characterises the Bela ophiolite belt, between Khuzdar and Karachi. Mineralisation is entirely in the upper part of the Lower Jurassic Shirinab Formation. Main deposits are located at Shekran, Ranj Laki, Mal Khor (NW of Khuzdar), Gunga, Surmai (SW of Khuzdar), and Duddar (SE of Bela). In addition to these at least 27 showings of Pb-Zn ore have been located in the Piaro and Mor Ranges, east of Bela (Fig. 9.4). Geochemical exploration in these ranges has further delineated areas with high Pb-Zn anomalies (Akhtar et al. 1989).

The Gunga, Surmai and Duddar deposits have been explored and evaluated in detail. Surmai is a stratabound Mississippi-type deposit hosted in the middle part (Loralai Member) of the Shirinab Formation. The Gunga and Duddar deposits occur in the upper part (Anjira Member) of the Shirinab Formation and are of the sedimentary-exhalative (sedex) type. The deposits are constrained between major faults and have been intersected by smaller faults (Fig. 9.5). The Duddar deposit shows a well-defined stratigraphy. The lower part comprises a Zn-Fe sulphide zone, followed by a middle high-grade Zn-Pb sulphide zone which is overlain by the upper Fe-sulphide zone (Fig. 9.5). Barite is independent of the sulphides and forms a separate body. The Duddar deposit also bears evidence of multiphase mineralisation and overprinting of earlier by later phases. The following paragenesis has been proposed (Jones et al. 1994).

$$FeS_2 - Ba - (SiO_2) - FeS_2 - ZnS - PbS - SiO_2 - \text{Chalcedony} + Ba + Ca.$$

At Duddar baryte is probably exhalative and formed on the ocean floor, whereas the sulphide mineralisation is syn-diagenetic and formed by displacement or replacement of the host from siliceous fluids. Deformation of sulphide layering and pressure solution features show that the ore was emplaced before early deformation and is, therefore, pre-Tertiary.

The Pb-Zn ores in the Karakoram block and the Chagai magmatic arc have not been explored and evaluated in detail. These are, however, of some economic interest owing

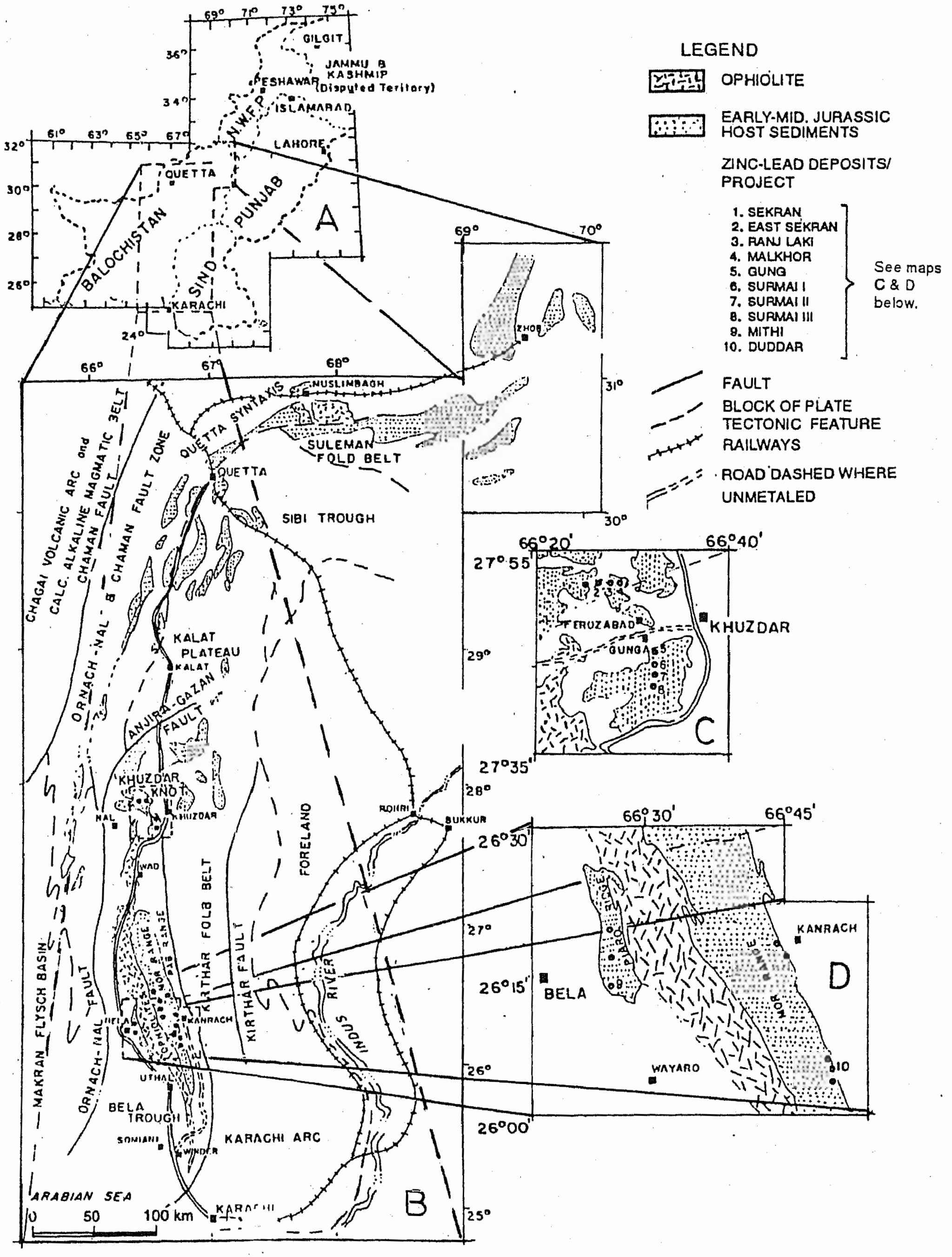

Fig. 9.4. Map showing distribution of Lead-Zinc deposits and prospects in the Bela-Khuzdar area (modified from Ahsan and Khan 1994).

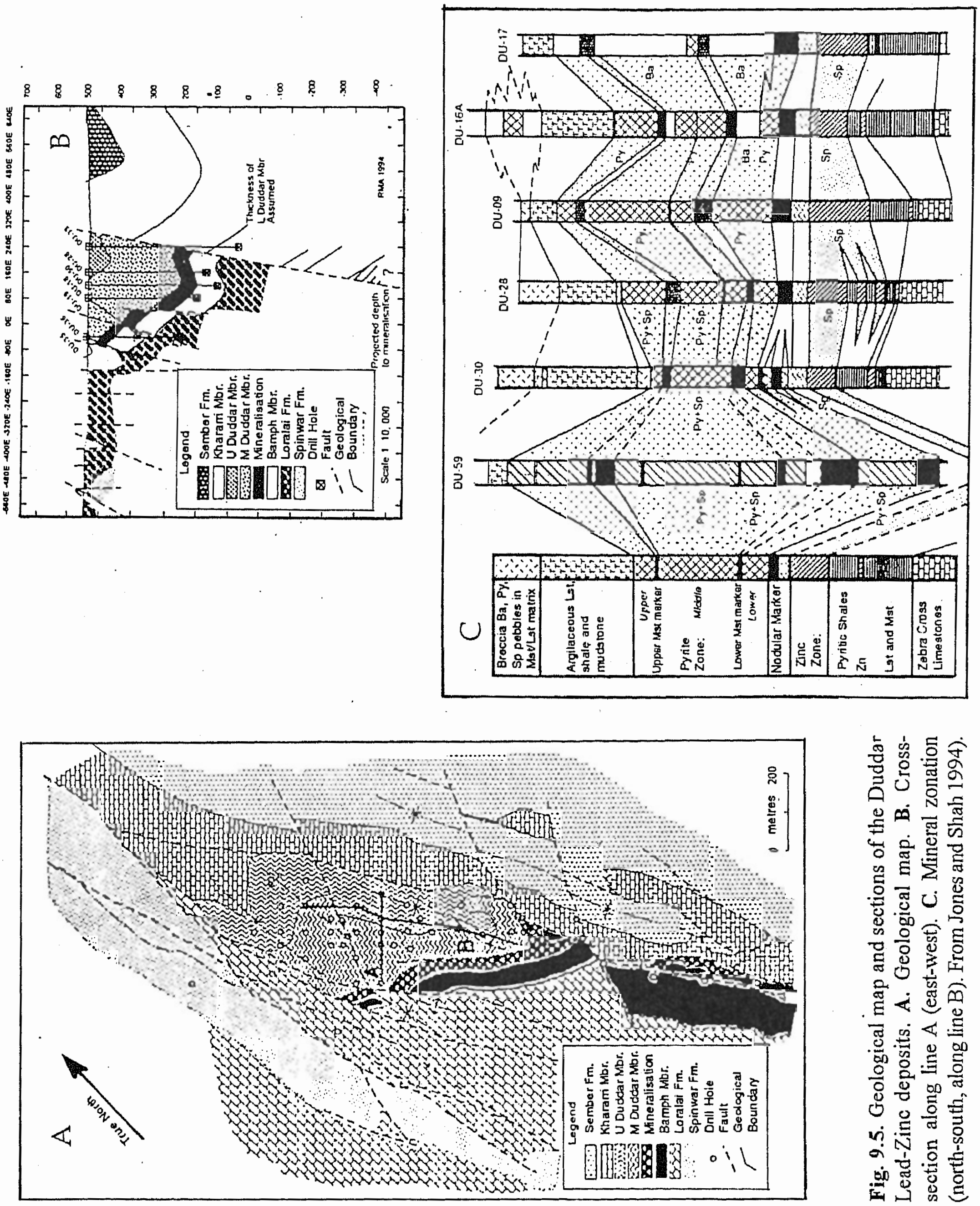

Fig. 9.5. Geological map and sections of the Duddar Lead-Zinc deposits. **A.** Geological map. **B.** Cross-section along line A (east-west). **C.** Mineral zonation (north-south, along line B). From Jones and Shah 1994).

to their gold and silver content. Only small. sporadic mining of Pb-Zn ore has been done in Chitral and Chagai Districts. The estimated reserves of Pb-Zn ore in Pakistan are over 16 million tonnes (Table 9.2).

Manganese Ore

Several small deposits of manganese ore occur in Pakistan, mainly associated with the volcanic rocks in the ophiolitic thrust belt extending from Bela through Zhob and Waziristan to Parachinar (Table 9.2). In this belt, veins and lenses of manganese ore occur in the volcano-sedimentary sequences of various melanges, commonly at the contact of volcanic flows and the overlying sedimentary rocks. The occurrences are largely characterised by cherty or jasperoid layers. Only two minor showings are reported from the Chagai magmatic arc and two small deposits are located in the NW Himalayan fold-and-thrust belt near Abbottabad. In this region several discontinuous lenses occur in a red-bed hematitic sequence in the Cambrian Hazara Formation.

The total estimated reserves of manganese ore in Pakistan are about 0.5 m.t. (Table 9.2) and the annual production has varied from 100 to 600 tonnes (Butt and Latif 1992). Most of this production comes from the Las Bela region.

Radioactive Minerals

Radioactive minerals, mainly those containing uranium, occur as showings or small deposits at several localities in the Foreland sedimentary fold-and-thrust belt and the Himalayan crystalline zone. They are linked with a wide range of metallogenic environments, which include occurrences in granitic rocks, alkaline magmatic complexes and carbonatites, migmatites and pegmatites, graphitic schists and Neogene to Recent placers (Table 9.6). The main economic deposits are, however, located at Baghal Chur in Dera Ghazi Khan District.

The fluviatile, cross-bedded sandstones of the Dhok Pathan and Nagri Formations host the uraniferous placers. Radioactive anomalies have been traced along a 190 km long stretch of the N-S oriented outcrop of the Dhok Pathan Formation along the foothills of the Sulaiman Range, in Dera Ghazi Khan and Dera Ismail Khan Districts. Seven of these anomalies have shown strong radioactivity and one of these, at Baghal Chur, has been developed as a producing mine (Moghal 1974b).

The mineralisation largely occurs in small lenses or pockets which are commonly located in cross-beds. The main source of uranium are the interstraitified ash beds, which contain uranium bearing feldspars (Moghal et al. 1997). The main radioactive mineral is tyuyamunite – a hydrated vanadate of calcium and uranium. It occurs in the interstices of sandstone as coating on sand grains and pebbles. Uraninite and coffinite also occur in the proximity of water table (Basham et al. 1974). The ore reserves are reportedly equivalent to 100 tonne uranium, though the overall resource potential is likely to be much higher (Nuclear Fuel 1986).

In the northern areas, uranium occurs extensively, though in minute quantities, in the placers of the Indus and its tributaries. Ahmad et al. (1976) have reported uraniferous placers from the following localities (U_3O_8 in %):–

Indus River-	Skardu-Oldhang section	(0.045).	**Hunza River-**	Hunza section	(0.0215).
	Shigar Valley	(0.062).	**Chitral River-**	Kunhar Valley	(0.0025).
	Astor Valley	(0.015).		Chitral Valley	(0.0019).

Table 9.6. Uranium occurrences and their geological setting.

Metallo-genic Zone	Name of deposit & location	Minerals present	Geological setting	Quality/grade	References
Himalayan crystalline zone	Karakar (Swat Dist.)	Monazite, uranothorite	Quartz-uranothorite-sulphide-monazite veins within the contact aureole of the Cambrian Ilum granite gneiss.	—	Butt 1989a.
	Ahl (34°32':73°09')		Secondary uranium mineralisation in shear zones in Cambrian Mansehra granite gneiss. Also Pleistocene placers overlying the granite.	U_3O_8- 0.03-0.2%	Butt 1989b.
	Rajdhwari (34°40':73°08')	Billibinite, beryl, mica,	In pegmatites cutting Mansehra granite. Uranium		Khan 1964.
	Loe Shilman (30 km W of Warsak).	pyrochlore	occurs as disseminations in isolated patches, in amphibole-carbonatite associated with the alkaline magmatic complex.	U-200 ppm	Butt 1989a. Shabbir et al. 1976.
	Silai Patti (35 km W of Dargai- 34°28':71°54')	Pyrochlore	In carbonatites associated with Malakand granite.	U_3O_8- 0.1-0.2%	Butt 1989a.
	Thakot (near Besham)	Uraninite	U occurs as disseminations in biotite pegmatites and sulphide bearing veins near the contact of Thakot Fm. and Lahor granite gneiss (both Proterozoic).	Low grade	Butt 1989a. Baig 1990.
Foreland fold and thrust belt	Reshian, ESE of Muzaffarabad	Brannerite, xenotime, uraniferous limonite.	In pyrite bearing graphitic schist, interbedded with pelitic schists and quartzitic schist of Precambrian Salkhala Fm.	Low grade	Butt 1989a.
	Parachinar Kurram Agency		Uranium mineralisation in accessory allanite in anatectic granites and pegmatites.	Low grade, small showings.	Butt 1989a.
	Baghal Chur 30°15':70°17' (NW of Dera Ghazi Khan)	Metatyuyamunite, uraninite, coffinite	Placer-type uraniferous lenses in sandstones of Mid. Siwalik Dhok Pathan Fm, in Baghal Chur syncline. Presently being mined. Reserves approx. 100 t Uranium.	Low grade (0.03-0.1% U_3O_8).	Moghal 1974a,b.
	Qubul Khel (Bannu Basin)	Pitchblende, coffinite.	Placer-type deposit in Dhok Pathan Fm.	Low grade	Baig 1990.
Kohistan magmatic arc.	Bunji 34°40':74°39'	Uraninite.	In pegmatites intruding meta-sediments and meta-volcanics (Hanuchal Fm) diorites (Shuta Gabbro).	Trivial	Shams 1995. Madin et al. 1989.

NON-METALLIC MINERAL DEPOSITS

Anhydrite

Anhydrite occurs with gypsum at many localities in the Salt Range. According to Ahmad 1969, it is not possible to quarry gypsum and anhydrite separately. Drilling at Dhariala (NW of Khewra), however, revealed thick, pure anhydrite beds near the top of the Precambrian Salt Range Formation. These anhydrites are of a blue grey colour, similar to dolomite and according to Heron et al. (1954) some of the dolomite associated with salt marl in the Salt Range may prove to be anhydrite.

Asbestos

Small deposits and showings of chrysotile and tremolite are found in serpentinites of the ophiolitic complexes near Wad, Muslimbagh, Naweoba (Zhob), Boya, Kaniguram (Waziristan), and Sakhakot-Qila area (Malakand). Only the Malakand deposits have been mined. Presently annual production from 15 mines is estimated at 50,000 tonnes (Jehan et al. 1997).

Barite

Barite deposits are largely confined to the Foreland sedimentary fold-and-thrust belt and the Bela ophiolitic belt, though small occurrences have been reported from the Karakoram block and the Chagai magmatic arc also. Four main types of deposits have been identified–(a) stratiform, (b) stratabound-replacement type, (c) veins and cavity fillings and (d) residual (Ahsan 1994). Though the vein and cavity-filling type mineralisation is wide-spread in sedimentary, igneous and metamorphic rocks of variable ages, the stratiform deposits are of greater economic value due to their larger reserves.

Small vein-type deposits in Precambrian Tanawal Formation, with estimated reserves of up to 1,000 tonnes, were reported from Kag–Alui and Kachi area, 10 km NW and 20 km NE of Haripur respectively.

Another vein-type deposit in Precambrian Hazara Formation, 5 km SW of Kohala (34° 06':73° 27') with estimated reserves of about 30,000 tonnes, has been mined and probably exhausted (Ahmad and Siddiqi 1992).

Large, stratabound barite deposits hosted in the Jurassic Shirinab Formation occur in the Khuzdar–Las Bela region. Some of these are associated with lead-zinc deposits. Relevant information on these has been summarised in Table 9.7.

Bauxite

Bauxite and laterite with relatively high alumina content occur at several localities in the Salt Range and the ranges of the Lesser Himalayas. They are mainly located along major unconformities and described in many papers under various lithologic terms e.g., clayey bauxite, bauxitic clay, bauxitic material, lateritic bauxite, lateritic clay, pisolitic clayey laterite, high-alumina clay etc. (Ahmad 1969). High-alumina residual deposits which may be categorised as 'commercial bauxite' mainly occur in the Katha–Pail area of the Salt Range, the Chhoi–Akhori area of the Kalachitta Range and the Muzaffarabad–Kotli region of the Hazara–Kashmir Syntaxis in Azad Kashmir.

Katha–Pail deposits

A clay-bauxite-laterite bed, 1 to 7 m thick, occurs extensively at the unconformity

Table 9.7. Barite deposits of Khuzdar–Bela region (Bela ophiolite belt).

Locality	Geological setting	Size, estimated reserves, production
Shekran 27°88':66°38'	Stratabound-replacement type deposit with fracture filling, hosted in Shirinab Fm. Barite is associated with galena, siderite and fluorite.	Not estimated
Gunga 27°44':66°32'	Stratiform and stratabound deposits in Anjira Mem. of Shirinab Fm. Main barite bed underlain by siliceous gossan containing Sedex-type Pb-Zn-barite deposit. Ore contains 40–95% $BaSO_4$.	9.5 m.t (measured). Annual production 10-20 thousand tonnes.
Kudni–Dham 26°35':66°31'	Fracture fillings and replacement deposits along crest of an anticline in Spingwar Mem. of Shirinab Fm. Ore contains 40–95% $BaSO_4$.	Reserves >70,000 t. (approx).
Araro (170 km NW of Karachi)	Mineralisation along bedding plane and fault zone in carbonate and clastic rocks (Shirinab Fm.).	21,000 t. (estimated).
Bankhari (160 km NW of Karachi)	Vein-type deposit along shear zone in Spingwar Mem. (Shirinab Fm.) Ore contains 80–95% $BaSO_4$.	64, 000 t (estimated).
Kinar Dhoro 26°15':66°39'	Stratiform deposit in marl and mudstone of Anjira Mem. (Shirinab Fm.).	85,000 t. (estimated).
Kanraj (6 km W of Kanraj 26°16':66°45')	Stratabound replacement and fracture filling-type deposit in Spingwar Mem. (Shirinab Fm).	92,000 t.
Zibro Dhoro (5 km SW of Kanraj)	Similar to Kanraj.	33,000 t.
Duddar 26°05':66°56'	Pb-Zn-barite stratiform Sedex-type deposit in Anjira Mem. (Shirinab Fm.).	1.86 m.t. (proved).
Bambh (3 km S of Duddar)	Similar to Duddar.	9,500 t. (estimated).

(Compiled from Mohsin et al. 1983, Jankovic 1984, Ahsan 1994).

between the Permian sequence (Wargal and Amb Formations) and the Paleocene Hangu Formation. It contains 35.5 to 72.5% alumina, 8.68 to 50% silica and 10 to 20% iron (Ashraf et al. 1972). There is considerable vertical and lateral variation in the composition of the deposit. Although it may be categorised as a low grade bauxite with high silica, moderate to low alumina and iron, it contains several large lenticular bodies of high-alumina bauxite (Al_2O_3 63%, SiO_2 9.5%, Fe_2O_3 4.18%). The ore contains boehmite and kaolinite, with minor amounts of diaspore and gibbsite. The reserves have been estimated at >100 million tonnes (Khan and Hussain 1970, Cheema 1974, Cruijs 1975).

Chhoi–Akhori deposits

These deposits occur in a 20–25 km long belt in the highly tectonised and imbricated sedimentary sequence of the Kalachitta Hills. The deposits occur at the unconformity between the Paleocene Hangu and Cretaceous Lumshiwal Formations and lower down at the unconformity between the Jurassic limestone and the Triassic Kingriali Formation. The upper horizon comprises an upper unit of up to 3 km thick oolitic or pisolitic laterite and a lower unit of bauxite material up to 4 m thick. In this zone aluminous clay, laterite, and bauxite occur as lenticular bodies associated with ironstone, ferruginous sandstone, claystone and quartzose sandstone. The bauxite contains approximately 32–76% Al_2O_3, 2.5–43% SiO_2, 0.25–12.0% Fe_2O_3 and 2.2–4.2% TiO_2 (Cruijs 1975). The lower lateritic horizon at the base of the Datta Formation, is more widely distributed and contains aluminous clays, claystones and laterites. According to Cruijs (1975) most of this could be used for recovery of alumina. The alumina clay samples from Chhoi indicate 74.24–86.84% Al_2O_3, 0.64–0.74% Fe_2O_3, 6.0–7.0% SiO_2 and 3.65–4.28% TiO_2 (Hussain and Naqvi 1973). Some of this material is being mined and used as fireclay. Estimated reserves (down to 33 m depth) are shown in Table 9.8.

Table 9.8. High alumina clay and bauxite deposits of Kalachitta Range

Locality	Alumina content(%)	Reserves (m.t.)
Bagh Nilab	40–50	2.0
Chhoi	74–86	11.0
Surg	45	13.0
Buta	–	17.0
Akhori	55	3.5

(From Hussain and Naqvi 1973).

Muzaffarabad–Kotli deposits

Small, scattered bauxite deposits are centred around the two outcrops of the Cambrian Sirban Formation of Abbottabad Group near Muzaffarabad (34° 21':73° 30') and Kotli (33° 31':73° 54'). They occur along the unconformity between the Sirban Limestone and the overlying shales of the Eocene Patala Formation. The bauxite is pisolitic, embedded in a clayey matrix, varies in thickness from 0.5 to 1.2 m and contains 51–89% clayey matter, 0–12% boehmite, 0–14% gibbsite, 1.5–10% quartz, 0–35% chalcedony, 2–8% iron oxides (Ahmad and Siddiqi 1992). The chemical contents and reserves of the ore at various localities are shown in Table 9.9. These deposits have been sporadically mined.

Table 9.9. Grade and reserves of Muzaffarabad–Kotli bauxite deposits

Locality	Alumina (%)	Silica (%)	Iron (%)	Reserves (m.t.)
Dhanwan	41–60	18–40	1–8	4.9
Kanroti (33°30':74°02')	50–70	9–28	1–2.5	1.36
Sawar	52–56	25	5	0.93
Dandili (33°22':73°58')	34–46	36–44	–	1.18
Nikial (33°29':74°04')	41–46	13–35	2–27	0.424
Goi	47	35	–	1.103
Shisetar (33°28':74°03')	NA	NA	NA	0.656
Bermoach	51	23	–	0.20
Balmi	46	31	–	0.209
Khandar Karela (33°26':74°06')	NA	NA	NA	0.209
Palan	NA	NA	NA	0.283
			Total	11.454

NA= Data not available.(From ECL 1979).

Celestite

Small to medium-size celestite deposits are being mined in Pakistan at two localities. The Daud Khel deposit (32°35':71°35') is located in the Salt Range near Jabba and Khairabad. Celestite occurs in irregular veins and as nodules in a 12 m wide mineralised zone, associated with a massive gypsum bed at the top of the Eocene Sakesar Limestone. Here the Eocene sequence is overlain by the Murree Formation. The celestite reserves were estimated at about 10,000 tonnes down to 6.5 m and the ore contained 82.7% $SrSO_4$, 9% SiO_2 and 3.5% CaO (Heron et al. 1954).

The Thano Bula Khan deposit (25°22':67°48') occurs in the Eocene Laki Formation in the southern part of the Kirthar fold belt. The mineralised vein follows a major strike-fault for over 25 km. The vein contains celestite, calcite and small amounts of hematite, limonite and pyrite. The ore contains over 98.57% $SrSO_4$ and the reserves were estimated at 320,000 tonnes (Moosvi 1973). The annual production of celestite in Pakistan varies from about 3,000 to 1,000 tonnes (Butt and Latif 1992).

Clays

A variety of clay deposits, ranging from kaolin, Fuller's Earth, fireclay, ball clay to pottery clay, occur in the sedimentary sequence of the Foreland fold-and-thrust belt and in the Indus platform zone (Kazmi and Safdar 1963). Most of these deposits occur in terrigenous sequences or near major unconformities, e.g. in the Permian Tobra Formation, in the basal part of Early Jurassic Datta Formation, in the Paleocene Hangu and Patala Formations and the Sonahri Member of the Laki Formation.

Kaolin deposits occur at Ahl (34°34':73°09'), Shah Dheri (34°53':72°54') and near Nagar Parkar (24°16':70°44'). The Ahl clay resulted from in situ weathering of granite. It is of low grade and the reserves were estimated at about 65,000 tonnes (Mumtazuddin 1951). The Shah Dheri (Swat) deposits are the weathering product of plagioclase-rich quartz diorite and kaolin occurs as patches, pods and streaks in unaltered rocks. The reserves were estimated at about 2.8 million tonnes (Moosvi et al. 1974). The best and the largest kaolin deposits are situated in the vicinity of Nagar Parkar and comprise several large lenticular pockets within the residual lateritic clay and laterites derived from the weathering of the Proterozoic Nagar Parkar granite (Kazmi and Khan 1973). The proven reserves are over 3.6 million tonnes (Kella 1983). The Shah Dheri and Nagar Parkar deposits are being mined.

Bentonite occurrences are confined to the sedimentary fold-belt of the Lesser Himalayan Ranges, and the Kohat-Potwar Plateau. One of these, the Karak deposit, is of Eocene age. It overlies the Bahadur Khel Salt Formation. It is overlain by the Jatta Gypsum Formation. It is 56 km NW of Bannu and covers an area of about 18 sq.km. On an average it contains 52.27% silica, 18.71% alumina, 5.98% ferric oxide, 0.59% titania, 3.22% lime, 4.68% magnesia, 0.79% soda and 1.33% potash. The clay is of smectite type. Its bleachability is 86–89% though its swelling index is only 300%. This is the largest bentonite deposit in Pakistan and the reserves have been estimated at about 36 million tonnes (Ahmad and Siddiqi 1992).

Other bentonite deposits occur in the Miocene-Pliocene sequence of the Rawalpindi and Siwalik Groups. The Chasma Wali deposit near Jabli village (33°13':71°36') is in the basal part of Chinji Formation. Most of the other deposits are in the Dhok Pathan Formation. They include the Rohtas (32°58':73°35'), Padhrar (32°40':72°30') and Ganda (33°51':73°31') bentonites of the Potwar Plateau, and the Bhimbar (32°58':74°05') and Chitta Dheri (33°01':74°03') deposits in the Sub-Himalayan region of Kashmir. The Bhimbar deposits are extensive and may be traced for about 48 km. They are of good quality and have been mined (Heron et al. 1954, Ahmad and Siddiqi 1992).

Fuller's Earth (montmorillonitic clay with strong absorption properties and a high colloidal content) occurs in the Eocene sequence of Sulaiman–Kirthar fold belt and also in the Indus platform zone. These deposits form a part of the same general fluvial system which generated the coal deposits. The main deposits are located near D. G. Khan (in Ghazij Group and in Kirthar Formation), at Thano Bula Khan, and at Shadi Shahid between Sukkur and Kot Diji (in Laki Formation). The deposits are being mined and the reserves run into several hundred million tonnes. Annual production is more than 17,000 tonnes (Raza and Iqbal 1977, Kazmi and Abbas 1991).

Fireclay is associated with the bauxite and high alumina clays in the Kalachitta Hills (described earlier). The main deposits are at Bagh, Chhoi and Surge. Large deposits are located in the eastern part of the Salt Range at Manhiala (32°43':73°00'), Wehali (32°45':73°03'), Nali (32°42':73°02') and Dalwal (32°42':72°52'), where the clay occurs at the base of the Paleocene Patala Formation near the unconformity above the Permian Warcha Formation. At Ara (32°45':73°13') near Khewra, it occurs at top of the Tobra Formation. At Karauli and Ratucha (32°42':72°59'), it is in the form of underclay associated with Dandot coal seam (Ahmad 1969). A number of these deposits are being mined.

In the western part of the Salt Range, fire clay is found in the Jurassic Datta Formation and the main deposits are at Dhak Pass (32°40':71°47'), Manza Bazar (32°38':71°48'), Chabil (32°40':71°47'), Dama (32°39':72°48') and Gole Wali (32°30':71°50'). In the Trans-Indus Salt Range, fire clay occurs at the same stratigraphic horizon and the main deposits are at Chapri near Isa Khel and Paniala (Dera Ismail Khan), and at Kot Kasarain (D. G. Khan) farther to the south. Small deposits of fire clay are located in the Eocene Sonahri Member of the Laki Formation, at Sonhari Dhand (near Jhimpir) and Kero (near Jungshahi).

The total fireclay resources in Pakistan are estimated at over 100 million tonnes and annual production is over 100,000 tonnes (Gauhar 1969, Shams and Khan 1987, Butt and Latif 1992).

Ball clay deposits are found in the Datta Formation near Sarai (32°30':71°53'). The deposits are apparently small (Kazmi and Safdar 1963).

Numerous small deposits of alluvial clays occur on the Indus Plain, largely as clay plugs in silted up channels and ox-bow lakes. Some of these deposits (e.g., Pasrur and Hala clays) are of good quality and have been used for glazed ware, pottery, glazed tiles and for mud in drilling of shallow wells (Kazmi and Safdar 1963).

Decorative stones

A wide variety of decorative stones is found in Pakistan. Large deposits of white, finesaccharoidal marble occur in the Paleozoic sequence at Shahid Mena (34°09':71°17'), Maneri (34°08':72°28'), Ghundai Tarako Hill (34°13':72°28'), Saidu Sharif (34°53':72°54'), Nowshera and at Karora. In this region some of the marbles are coloured in shades of brown, pink, red, grey and black. Very large deposits of white crystalline marbles are found in Chitral and Hunza areas.

High quality "onyx marble" (aragonite and travertine) in shades of green, brown and red occurs in large quantities in the Chagai District and have been mined extensively. Beautiful, brown- and white banded aragonite occurs in the Thano Bula Khan area. The multicoloured porcelainous Parh limestone has been used widely as a decorative stone. The pink and grey granites of Nagar Parkar are being quarried and are greatly fancied for their beauty and colour. A wide range of exquisite fossiliferous limestones and brecciated rocks from various melanges are also being marketed.

Dolomite

Extensive deposits of dolomite occur in the Foreland fold-and-thrust belt and are found in sedimentary sequences ranging from Precambrian to Eocene (see Chapter 5). The main lithostratigraphic units which contain large deposits of dolomite are the Precambrian Salt Range Formation, the Cambrian Jutana and Khisor Formations, the Devonian Shagai Limestone, the Triassic Kingriali Formation, the Jurassic Samana Suk and Takatu Limestone and the Eocene Chorgali Formation. In the Paleozoic sequence of the Karakoram block, dolomite is found at several localities, particularly in the Devonian Kuragh dolomite sequence. The major dolomite deposits have been listed in Table 9.10.

Table 9.10. Dolomite deposits.

Deposits	Formation	MgO%	CaO%	Reserves
Kachhi (Haripur)	Abbottabad Fm.	NA	NA	Large
Sherwan	Abbottabad Fm.	NA	NA	Large
Wagh	Precambrian	21.02	31.43	Large
Nilawahan	Salt Range Fm.	19.72	28.71	Large
Saidu Wali (Khisor Range)	Salt Range Fm.	20.11	31.05	Large
Ghundai Tarako (Mardan)	Paleozoic	19–21	31–33	Not known
Khyber Agency	Khyber Fm. Shagai Lst.	20.1	30.1	Large
Kuch (Kalabagh)	Kingriali Fm.	21.8	30.8	0.5 m.t.
Makarwal (Datta, Doya–Lunda, Narmai–Punnu)	Kingriali Fm.	20.2 to 21.6	28.9 to 31.3	900 m.t.
Burikhel (Mianwali)	Kingriali Fm.	16.8	31.05	Large
Kalachitta	Kingriali Fm.	NA	NA	Very large
Chiltan Range	Takatu.Lst.	20	32	250 m.t.
Kohat (NW of Pail)	Chorgali Fm.	17.95	26.32	Large

NA= data not available. (From Khan and Gauhar 1966 and Raza and Iqbal 1977).

Feldspar

Large deposits of sodic and potassic feldspar are found in pegmatites associated with the Swat Granite Gneiss (Mingora deposits), the Mansehra Granite (deposits near Khaki–Susa gali and Rajdhwari), Nanga Parbat Gneisses (Khaltoro, Shengus, Bulechi and Stak deposits), Karakoram Batholith (Dassu, Niyet Bruk, Hunza deposits), and Tirich Mir, Hot Spring and Kafiristan plutons. Deposits of feldspar occur in pegmatites of the Nagar

Parkar Granite. The annual feldspar production is about 10,000 tonnes and most of it is used in ceramic and glass industry (Heron et al. 1954, Khan 1987, Kazmi et al. 1990, Kazmi and Abbas 1991).

Fluorite

Fluorite has been reported near Mirgasht and Yarkhun in Chitral, in the northern sedimentary fold-and-thrust belt of the Karakoram block (Mining World 1959). In the Himalayan crystalline zone, two small deposits near Chakdarra (Dir) comprise fluorite-quartz veins cutting two-mica Swat Granite Gneiss (Sillitoe 1979). In the Khyber–Hazara metasedimentary fold-and-thrust belt a small showing of fluorite has been reported from Bichoha Kurd (34°11':73°03') near Sherwan, where it occurs as disseminations in silicic dykes cutting schists of the Hazara Formation (Nagell 1969). Trivial showings occur in the Zhob ophiolite belt at Brunj Kili and Khojakzai Kalai (31°33':69°31') where fluorite-calcite veins cut the Triassic Wulgai Formation (Heron et al. 1954, Nagell 1969).

The largest fluorite deposits occur in Jurassic Chiltan (Takhatu) limestone of the Kirthar fold belt. The Phade Maran (29°25':66°50') deposit in southern part of the Koh-i-Maran Range is characterised by fluorite, galena, sphalerite and calcite. It forms concordant, replacement bodies along sheared or unconformable contact of the Chiltan Limestone and the overlying Sembar Formation. Fluorite also forms fissure-fillings along minor faults and fractures. Reserves are estimated at about 100,000 tonnes. In the northern part of Koh-i-Maran Range there is a small deposit of fluorite (with calcite and barite) at Chah Bali (Abbas et al. 1980, Ahmad 1975).

The other major fluorite deposit is at Dilband (29°32':66°55') in the Koh-i-Dilband Range which runs parallel to the Koh-i-Maran. The deposit contains fluorite and calcite and occurs under geological setting similar to the Koh-i-Maran deposits. The reserves are estimated at about 93,000 tonnes (Mohsin and Sarwar 1980). In recent years the annual production of fluorite from the Koh-i-Maran and Dilband deposits has ranged from 1,500 to 5,000 tonnes (Butt and Latif 1992).

Gemstones

A large variety of gemstones is found in Pakistan (Fig. 9.6). Some of these are being mined whereas the mere occurrence of the others has been reported. From the geological standpoint, the gemstones of Pakistan may be placed under the following four main categories (Kazmi and O'Donoghue 1990).

1. *Suture-associated gemstones:* Emerald, peridote, epidote, actinolite, vesuvianite, ruby, sapphire, spinel and pargasite.
2. *Pegmatite-associated gemstones:* Aquamarine, emerald, feldspar (moonstone), garnet, quartz, topaz and tourmaline.
3. *Gemstone in hydrothermal veins:* Azurite, pink topaz, rutile and zircon.
4. *Miscellaneous gemstones:* Agate, green grossular garnet (tsavolite), hessonite garnet,

Some of the significant gemstone occurrences are briefly reviewed in this section.

Suture associated gemstones

These deposits occur along or in close proximity to the suture zones, that is the Indus suture zone and the Karakoram suture zone. The deposits are largely pneumatolytic, hydro

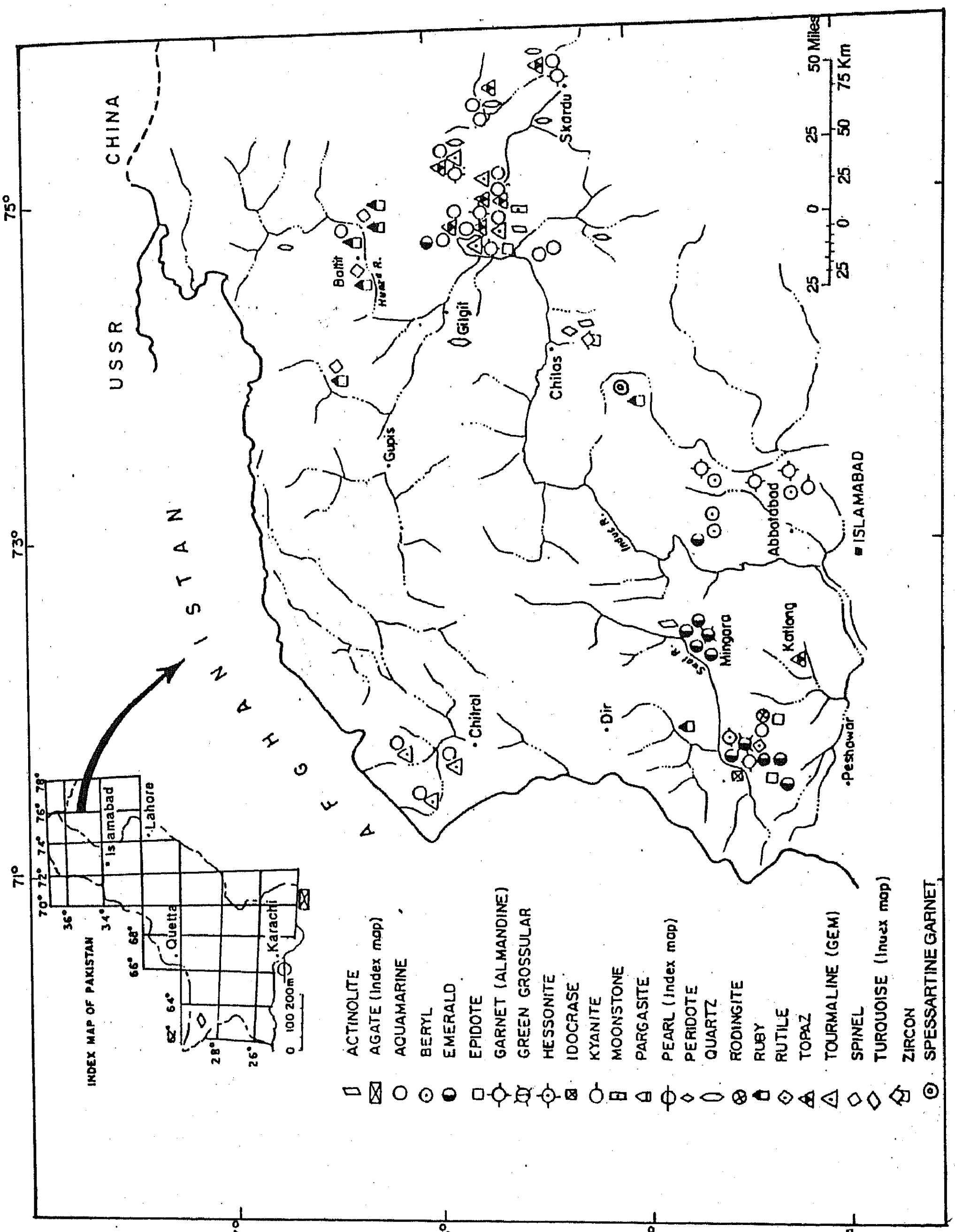

Fig. 9.6. Map showing gemstone occurrences in Pakistan (from Kazmi and O'Donoghue 1990).

thermal or metamorphic-cummetasomatic. They are the ultimate products of continental collision.

***Emerald*:** Deposits and showings of emeralds have been reported from Dandao Kandao, Bucha, Nawe Dand, Gandao, Kot, Arang Barang, Mingora, Charbagh, Makhad, Gujarkili, Shamozai and Khaltaro (Fig. 9.6). The Mingora and Gujarkili deposits have been mined regularly and the annual production has varied from about 78,000 to 102,000 carats. Others have been mined only sporadically.

The Mingora emeralds are the best and world famous. The deposit is spread over an area of 180 acres and emeralds occur in Mingora ophiolitic melange. This melange is composed of tectonised blocks and clasts of serpentinite, talc-dolomite schist, greenstone, metabasalt, green-schist and meta-chert (Kazmi et al. 1984). The talc-chlorite-dolomite schist is the main host rock for the emeralds. Thrust faults have formed several tectonic slices. Two parallel north-trending normal faults run through the deposit and are believed to be the principal avenues for mineralising fluids (Fig. 9.7). Except for Khaltaro, the geological setting of all other emerald deposits in Pakistan is similar to the Mingora deposit. The Khaltaro emerald occurs in pegmatites (Kazmi et al. 1989).

Epidote: In Pakistan most of the gem grade deposits of epidote occur in hydrothermal veins and as alteration products in gabbro and greenstone of the melange complex. These deposits are located in the Indus suture zone close to Bunji in Gilgit area; near Kot, Pranghar and Bucha in Malakand and Mohmand Agencies and at several localities in the Swat District (Fig. 9.6).

Peridote: An important deposit of peridot has recently been found near the Kohistan–Kaghan watershed to the NE of Naran. The gemstones occurs along shear zones and in pockets in dunitic host rocks, and are associated with clinochlore, magnetite and local magnesite. The host rocks occur in the immediate hanging wall of the Indus suture. They represent the basal cumulates of the Sapat mafic-ultramafic complex and presumably belong to Kohistan magmatic arc thrust onto India (Jan et al. 1993, Khan and Kausar 1996).

The peridot is transparent to translucent and pale- to dark yellowish green or, rarely, greenish yellow. A good proportion of the specimens is of brilliant gem quality and adequate size, and is thus suitable for faceting. It is medium- to coarse-grained with largest crystals reportedly measuring more than 10 cm in length and up to 2 kg in weight. Koivula et al. (1994) and Jan et al. (1995) have presented physical data and microprobe analses of the minerals. The peridot is mostly Fo_{91} in composition but ranges from Fo_{90} to Fo_{94}. Jan et al. (1995) suggest that the mineralisation may be related to hydrothermal solutions possibly derived from post-tectonic Himalayan leucogranites of Eocene age.

Ruby and sapphire: Ruby, and sapphire, spinel and pargasite are found along the Karakoram suture zone. These gemstones commonly occur side by side in metamorphosed, recrystallised marble beds which form part of the thick sequence of metasediments of Dumordo Group. The gem-bearing marbles extend for over 100 km, between the Hunza Valley (in the east) and the Ishkuman Valley (to the west). However, at present mining of the gemstones is confined to the Hunza Valley.

In this region the suture zone is characterised by a thin belt of greenschist, serpentinite, black slates and mylonite (Fig. 9.8). These rocks have been thrust southward over the Kohistan island arc sequence along the northern slopes of the Rakaposhi mountains (Fig. 9.9). The Dumordo Group of rocks comprising the gem bearing marble, garnet schists and

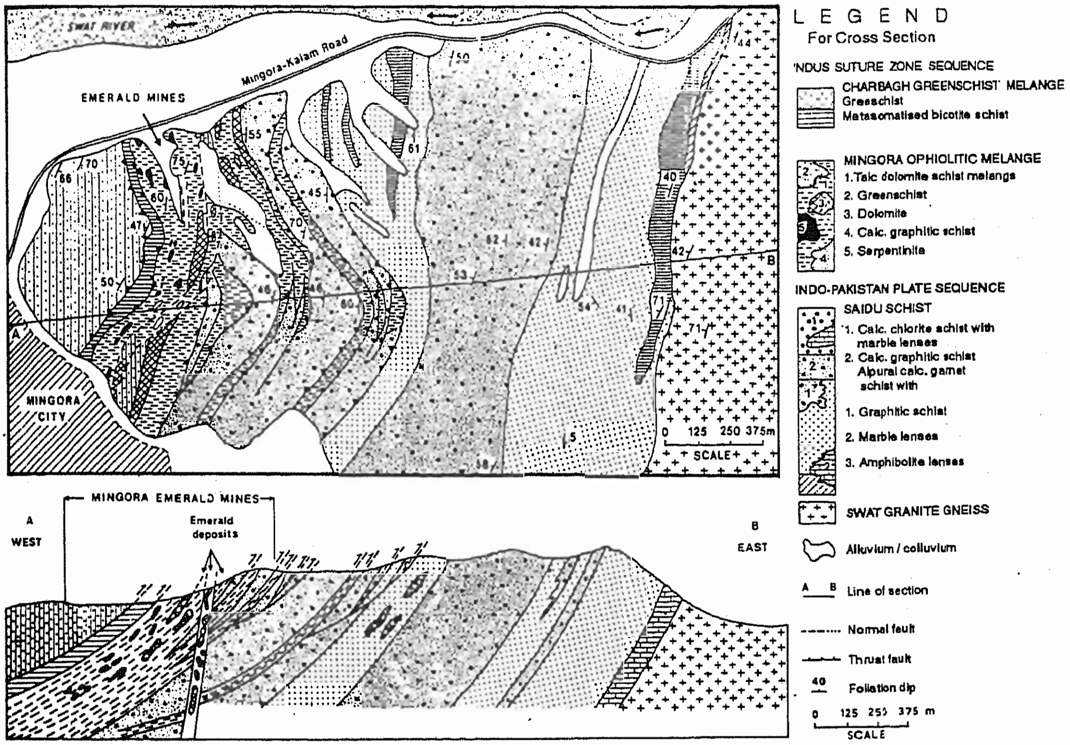

Fig. 9.7. Geological map and cross-section of the Mingora emerald-mine area (from Kazmi et al. 1984).

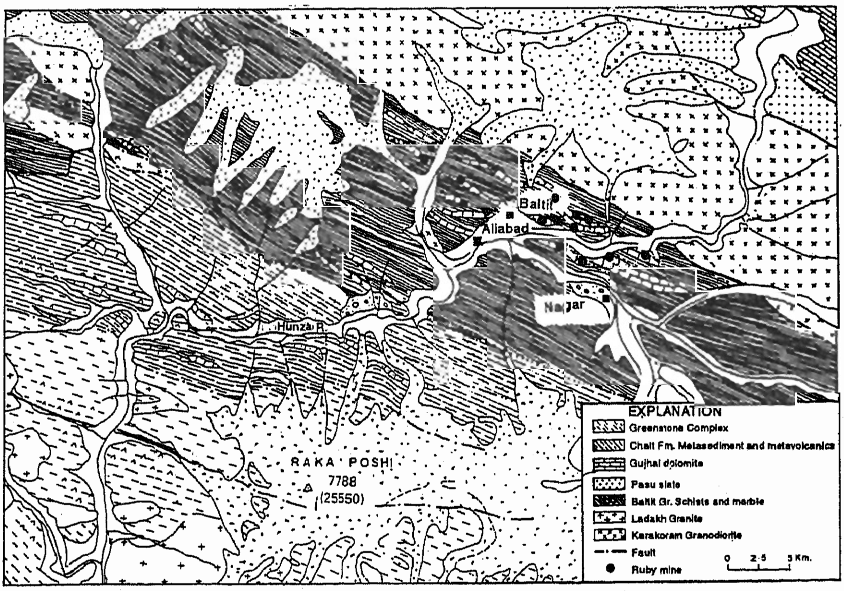

Fig. 9.8. Geological sketch map of Hunza area showing ruby mines (from Kazmi and O'Donoghue 1990, Kazmi 1995).

quartz-mica schists has been thrust southward over the rocks of the suture zone described earlier. It is noteworthy that the gem bearing marbles crop out adjacent and parallel to the suture zone. Northward, the Dumordo Group has been intruded by the granites of the Karakoram Batholith. Extensive apophyses of this granite form pegmatitic protrusions and aplite veins in the Dumordo Group. They are also intruded by mafic rocks.

Presently most of the mining has been confined to about 13 mining centres spread over a length of 15 km in the Hunza Valley. Some of the ruby crystals found in Hunza are of the classical "pigeon's blood" colour, resembling the Burmese ruby. Cabochon grade violet or indigo-coloured sapphires occur with the rubies. In addition to the rubies and sapphires, the marble host-rock contains the following mineral association: spinel, pargasite, margarite, phlogopite, chlorite, graphite, pyrite, rutile, dolomite, sphene, apatite, tourmaline, plagioclase, pyrrhotite, quartz, calcite and goethite (Okrusch et al. 1976, Gubelin 1982, Kazmi et al. 1990).

Good quality ruby is also found in the Shonther–Kalejander (34°59':74°23') region of the Neelam Valley in Azad Kashmir. The main deposits are at Nangimali, Chitta Katha, Naril, Kalejander and Khandi Gali–Maidanwali. In this region there are two main lithostratigraphic units – a basement sequence of granitic rocks, migmatites and metasediments (Migmatite Complex and Bhurjanwali Formation of Malik 1995) and a cover sequence (Nangimali Formation of Malik 1995) which is largely comprised of quartzites, amphibolites, limestone, and schist (Fig. 9.10). The ruby deposits are hosted in the crystalline limestone and occur in calcite veins along bedding planes. The rubies are transparent to translucent and brownish pink to pinkish red or deep red. The deposits are being mined and the reserves are estimated at about 24.9 million grams (Malik 1995).

A low quality ruby-sapphire is being exploited from the amphibolites of southern Kohistan near Timargarha, Dir.

Spinel: The Hunza spinels occur in a wide variety of colours ranging from brown, red, plum red, violet to blue. These colour variations are mainly a result of slight changes in chemical composition. The Hunza spinels are larger than those customarily found in Burma and are far attractive. They are commonly euhedral, showing recognisable crystal forms.

Pargasite: Exquisite deep pistachio-green crystals of pargasite are found along with ruby and spinel in the Hunza Valley. The stones are translucent to opaque and are commonly fashioned into beautiful cabochons. Locally they are sold as "Hunza emeralds".

Pegmatite-associated gemstones

These gemstones are found in the Northern areas where extensive pegmatite fields have formed consequent to the emplacement of granitic batholiths (Table 9.11). The pegmatites are largely concentrated in the Gilgit–Skardu region (Nanga Parbat, Haramosh and Karakoram Ranges), in the Indus Kohistan area, and in Chitral (Hindukush Range). Some of these pegmatites contain gemstones and mineral specimens ranging from beryl, aquamarine, topaz, green and multicoloured tourmaline, garnet, fluorite, microcline, orthoclase, montmorillonite, muscovite, and less commonly, cassiterite, fluorapatite, hambergite, hydroxyl-herderite, manganotantalite, and zircon (Kazmi et al. 1985). The gemstones that are more significant from the point of exploitation are briefly described below.

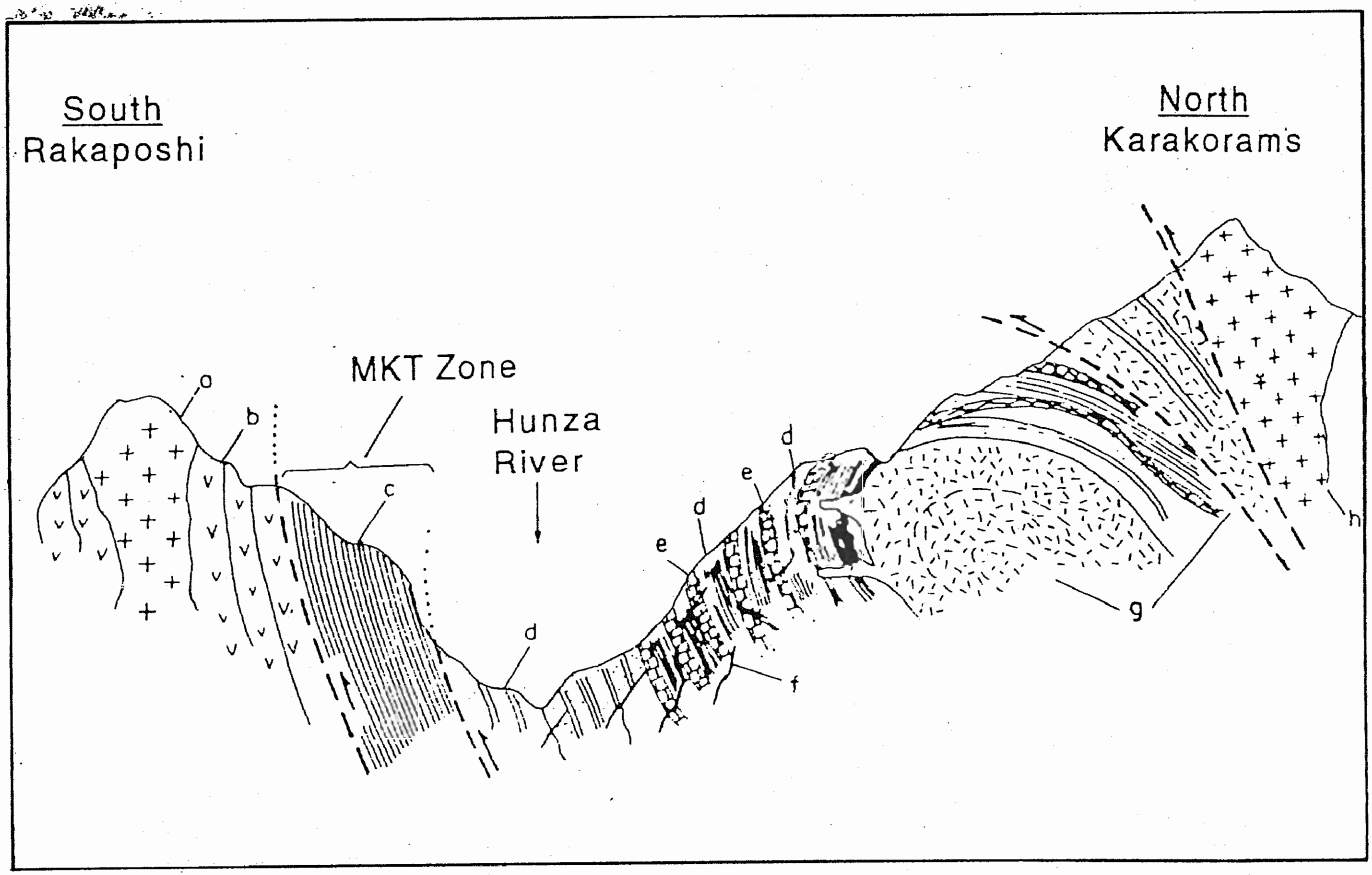

Fig. 9.9. Diagrammatic geological cross-section across Hunza valley near Aliabad a=granodiorite; b=metavolcanics, calcschists with marble bands; c=mylonite, greenschist, serpentinite lenses, black slates; d=garnet schist, biotite schists; e=ruby bearing white marble; f=aplite, diorite dykes; g=biotite granite gneiss; h=biotite hornblend granite (from Kazmi and O'Donoghue 1990).

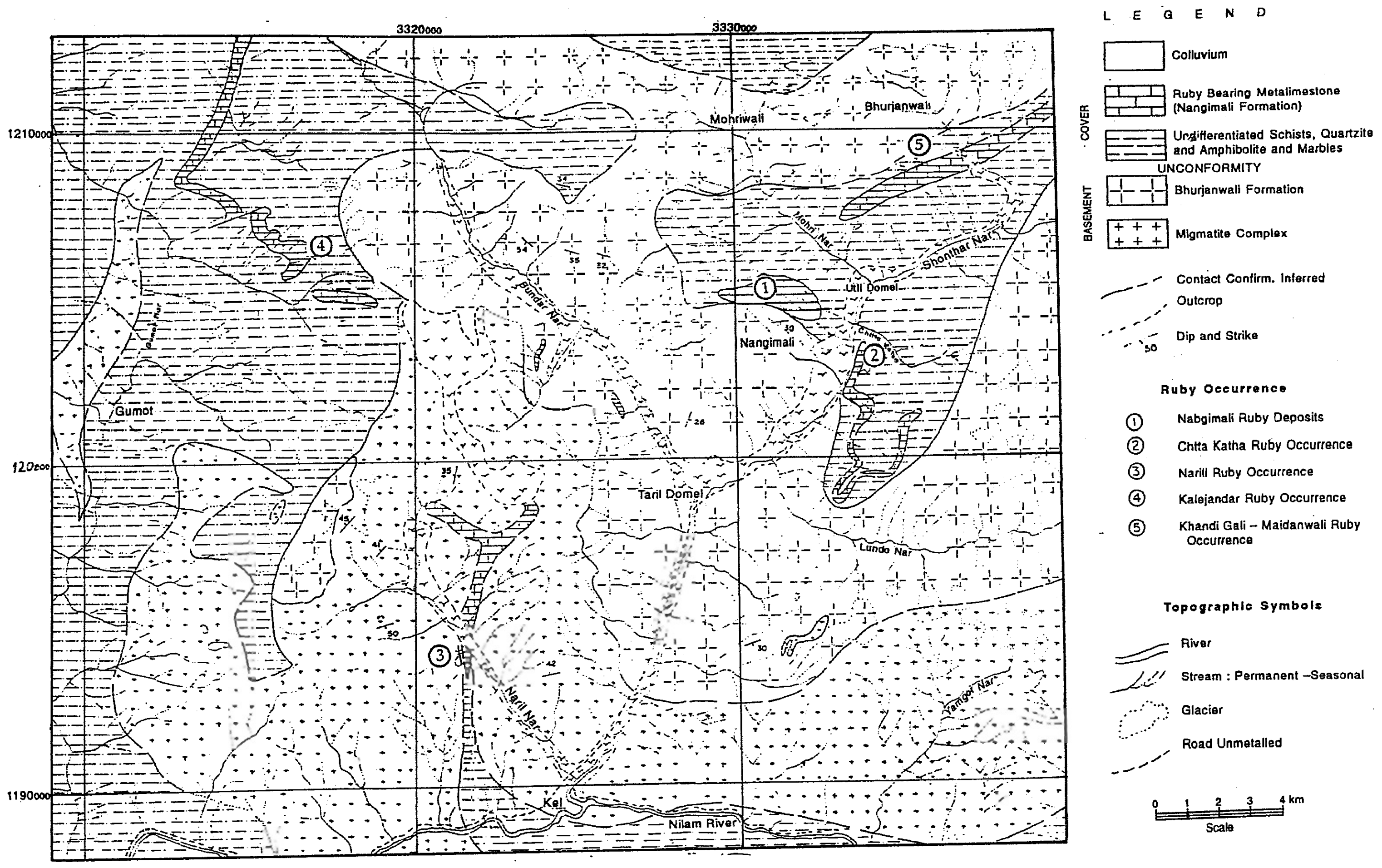

Fig. 9.10. Geological map of Shonter alley, showing Nangimali ruby deposits (from Malik 1955).

Table 9.11. Gem pegmatites of northern areas and their mineral content (from Kazmi and O'Donoghue 1990).

Location	Alttitude (metre)	Gemstones	Other minerals.
KARAKORAM			
Dusso	2,800-3,000	Aquamarine Rose quartz	Plagioclase feldspar, quartz, muscovite, biotite, schorl, beryl.
Nyet Bruk	4,600-5,000	Brown topaz	Plagioclase feldspar, quartz, tourmaline, muscovite, beryl.
Gone schorl.	2,600-3,000	Aquamarine Brown topaz	Plagioclase, orthoclase, quartz, muscoite, beryl, biotite,
Tigstun	2,600-3,000	Aquamarine Garnet	Plagioclase, quartz, muscovite, biotite, schorl, beryl.
HARAMOSH			
Stak Nala	3,300-4000	Bicolour tourmaline Aquamarine	Feldspar, quartz, mica, fluorite, apatite, schorl, beryl.
Skere Shah Batot Shingus	2,300-4,000	Aquamarine Topaz Goshenite Morganite	Feldspar, quartz, mica, garnet, schorl, tourmaline beryl.
Khaltaro	4,600-5,000	Emerald Aquamarine	Feldspar, quartz, mica, schorl, beryl.
DEOSAI MOUNTAINS			
Bulechi	2,600-4,000	Topaz Aquamarine Green tourmaline Bicolour tourmaline Moonstone	Feldspar, quartz, mica, schorl, garnet, beryl.
NANGA PARBAT			
Astore	2,500-4,000	Smoky quartz Green quartz (large books of mica)	Quartz, feldspar, schorl, garnet, beryl.
Buldargah	2,600-3,300	Aquamarine	Quartz, feldspar, mica, schorl, garnet.
HINDU KUSH			
Gobar-o-bakh (Luthko).	5,000-6,000	Aquamarine Tourmaline. Garnet Quartz	Quartz, feldspar, mica, tourmaline, beryl.
Dao Ghari	4,000-5,6000	Aquamarine Garnet	Quartz, feldspar, mica, tourmaline.

In addition to above listed minerals, stray specimens of fluor-apatite, manganotantalite, hydroxyl herderite and hambergite also have been reported from Karakoram and Haramosh pegmatites (Kazmi et al. 1985).

Aquamarine: It is found as euhedral crystals in the gem pegmatites of Dusso (35°42':75°29'), Tisgtung (35°43':75°28'), Bulechi (35°41':74°48'), Shingus (35°43':74°48'), Iskere (35°54':74°45') and Khaltaro (35°57':74°41') in the Gilgit–Skardu region, and in the Gabor-o-Bakh (36°5':71°21') region of Chitral. The Gilgit–Skardu aquamarines are rather lighter in colour, whereas the more attractive inky blue aquamarines are found in Chitral. Goshenite and morganite (varieties of beryl) are also commonly found in these pegmatites.

Tourmaline: Deposits of gem quality tourmaline (pink, blue and green) are found in the pegmatites in the Haramosh Range near Gilgit. The best known deposits are in the Stak Nala (between Gilgit and Skardu). The locality is now famous for bi-colour and tri-colour tourmalines. The colours are commonly dark green or black at the base with the terminations being grass-green, pink or blue. The crystals are euhedral with prismatic terminations and are commonly 4 to 6 cm in length.

Gem grade tourmalines also occur at Bulechi and Shingus (Gilgit Division). Indicolite (blue tourmaline) is found in the gem pegmatites of Garam Chashma (Chitral). Recently a new tourmaline locality has been discovered in the Neelum Valley of Azad Kashmir from where some of the best and fairly large specimens of bi-colour and tri-colour tourmalines have been found.

Topaz: Topaz found in the gem pegmatites ranges from colourless to slightly yellow, brown to deeper sherry-colour. Large, perfectly euhedral crystals are quite common. They mostly appear in a microcline-quartz-muscovite matrix. Topaz-bearing pegmatites are largely found at Bulechi (35°41':74°48') and Shingus (35°43':74°48') in Gilgit area, and at Niyit Bruk (35°43':75°31') and Gone (35°41':75°31') near Dusso in the Skardu area.

Quartz: Clear and well formed crystals of quartz occur in gem pegmatites in Skardu, Gilgit and Chitral areas, and in Azad Kashmir. Smoky quartz is also commonly found in these areas. Rose quartz is abundant in Dusso pegmatites near Skardu. Agate and chalcedony are found near Nagar Parkar (24°10':70°23') in Sindh, while jasper occurs in Lasbela area (Balochistan).

Gemstones in hydrothermal veins

These gemstones comprise pink topaz, zircon, rutile and azurite. The last three were observed only as scattered showings. Only the pink topaz, which is perhaps the foremost gemstone of Pakistan, is being mined at Katlang (34°21':72°04').

Pink topaz: The Katlang topaz is the only known naturally deep-red or deep-pink coloured topaz in the world. The colour of these Katlang stones ranges from colourless to very pale-beige to light-brown, from very pale to deep-pink to bright-red. Even violet-coloured crystals have been found (Jan 1978, Gubelin et al. 1986, Kazmi and O'Donoghue 1990). The Katlang topaz deposit is located in the Ghundao Hill, 4 km north of the town of Katlang, about 20 km north of the city of Mardan. The Ghundao Hill is a small anticlinorium with fold axes trending east-west. It consists of grey, thin- to thick-bedded Paleozoic (?) crystalline limestones which have been drag-folded and extensively faulted. The limestones contain stockworks of calcite and quartz veins which host the topaz mineralisation. The gem-bearing veins are mostly along a series of parallel, normal and reverse faults, near the crest of the tightly-folded anticlines.

Zircon: Beautiful euhedral crystals of brownish-red zircon have been found in quartz-calcite veins in schists and gneisses, about 20 km south of Chilas, in Gilgit Agency.

Rutile: Showings of brown to reddish-brown rutile have been observed in Silai Patti–Kolangi granite gneiss and talc-chlorite schist south of Kote. At both these localities rutile is opaque and the mineralisation is associated with hydrothermal quartz veins (Hussain et al. 1984).

Azurite: Beautiful crystals of azurite have been found associated with chalcopyrite, malachite and bornite in metavolcanics near Henzel village, 21 km northwest of Gilgit.

Miscellaneous gemstones

Agate: Among the miscellaneous gemstones, agate is one of the cryptocrystalline varieties of quartz; other varieties such as jasper, chalcedony, carnelian, and onyx are based on colours and patterns. Agate is the variety which usually contains coloured layers or bands, flat or concentric. Nodules of agate are found in cavity fillings in basaltic and andesitic lavas and are widely distributed in the Nagar Parkar area (Sindh) and in Dir Kohistan (NWFP).

Garnet: Gem quality almandine (red) garnet is found in Chitral District, red spessartine garnet in the gem pegmatites of Dusso and Shingus, in the Northern Areas. Tsavolite or green grossular appears in graphitic schists in the Jambil area of Swat, near Kot in Malakand District and Targhao in Bajaur Agency. The Jambil garnets have provided exquisite facetted gems, whereas the Targhao garnets make good cabochons. Near Targhao (Bajaur Agency) beautiful honey-yellow euhedral crystals of hessonite garnet are found in mineable quantities in quartz-mica schist.

A high quality, orange-red spessartine garnet has been recently discovered in pegmatites in Neelam Valley of Azad Kashmir. Large transparent crystals are common. Our unpublished data show that the garnet is rich in spessartine component (79–85 mole%) and has a density of 4.19.

Mineral Specimens: Mineral specimens are also of growing interest in Pakistan. The attraction of mineral specimens, as distinct from facetted stones, lies in their form and colour. Mineral specimens do not have to be of gem quality, though the few gem crystals that escape cutting are admittedly most beautiful. In recent years a large and a flourishing market for good mineral specimens as collector's items has developed world wide. The pegmatites of the Northern areas have yielded excellent light pink crystals of fluorapatite, green fluorite, spessartine, hambergite, green microcline, aquamarine, tourmaline, topaz, and garnet. Exquisite mineral specimens of ruby, spinel and pargasite are found in the Hunza Valley. Beautiful pyrite, malachite and azurite specimens can be collected near Gilgit. Attractive violet coloured fluorite crystals occur in the Koh-e-Dilband fluorite mines (29°30':66°55') in Kalat District. The Indus suture zone, between Chilas and Mohmand Agency, abounds with epidote, actinolite and rodingite; less commonly are zircon and rutile. Some of the rarest and best pink topaz specimens are found in the Katlang topaz mine.

Glass sand

Large deposits of glass sand occur in the sedimentary sequence of the Foreland fold-and-thrust belt. At Munda Guchha (34°40':73°16') near Mansehra, 150 million tonnes of silica sand have been reported from the Precambrian Hazara Formation (Heron et al. 1954, Raza and Iqbal 1977). Very large deposits of glass sand are found in the Jurassic Datta Formation in the Khisor and Marwat Ranges (between Paniala and Pezu), in the Surghar Range and the Salt Range (near Mianwali). In the Trans-Indus Salt Range, large deposits are located in the Cretaceous Lumshiwal Formation between Kalabagh and Makarwal and at Mallakhel (Gee undated, Heron et al. 1954). In the central and eastern Salt Range, the Paleocene Patala Formation contains extensive deposits of silica sand (Shah 1980).

In the southern part of the Kirthar fold belt between Meting and Jhimpir, large deposits of glass sand are found in the Eocene Sonahri Member of the Laki Formation. Near Jungshahi (24°52':67°10') the deposits are in the Oligocene Nari Formation and in the Unt Palan Range (25°13':67°31') they are in the Miocene Gaj Formation. Near Kach (48 km E of Quetta) glass sand deposits are located in the Miocene Nagri Formation (Heron et al. 1954, HSC 1960, Raza and Iqbal 1977). Annual production of glass sand is about 165,000 tonnes (Butt and Latif 1992).

Graphite

Small occurrences of graphite have been reported from several localities (Heron 1954, Ali 1959, Ahmad 1969). Graphite is found in the metasedimentary sequence of the Karakoram block, e.g., in Sewakht Formation near Shah Salim (36°20':71°09') in Lutkho Formation near Momi (36°03':71°43'), in Darkot Formation near Chhelish and Bola Das (36°36':73°17') and in Dumordo Formation near Chalt (36°15':74°17') and SE of Baltit (36°20':74°35'). In the Himalayan crystalline zone, graphite occurs in the Nanga Parbat–Haramosh Massif near Sasli (35°52':74°44'), in the Abbottabad Formation near Balakot (34°34':71°21'), and in the Alpurai Schist near Loe Agra (34°33':71°04'). In the Khyber–Hazara metasedimentary thrust belt it is found in Paleozoic rocks of Khyber Agency near Spinkai (35°55'30':70°41'), Shah Salim (36°05':71°09') and Shahid Mena (34°09':71°17') and in Hazara Formation near Sherwan (34°12':73°04'). The Loe Agra deposits are reportedly large and the reserves are estimated at about 18 million tonnes containing 15–20% graphitic carbon (Working Group on Minerals 1978). High quality graphite occurs near Mohriwali (35°01':74°25') in the Neelam Valley of Azad Kashmir. Here the measured reserves are about 1.0 million tonnes, containing an average of about 12% graphitic carbon (Ali 1959, Ministry of Petroleum and Natural Resources 1994).

Gypsum

Extensive deposits of gypsum are associated with Eocene and Precambrian sequences in the Sulaiman Range, the Potwar Plateau and the Salt Range. In the Salt Range, the evaporite sequence of the Precambrian Salt Range Formation contains many beds of gypsum, particularly in the Sahwal Marl Member and the Bhandar Kas Gypsum Member (Ahmad and Siddiqi 1992). In western Salt Range large deposits of gypsum occur in the Eocene Sakesar Limestone at Daud Khel (32°53':71°43') and Saiduwali (43°45':72°10'). Near Kohat, bedded to massive gypsum is exposed in the Early Eocene Jatta Gypsum Formation. In the Sulaiman fold-belt, gypsum is found in the Eocene Ghazij Group and the main deposits are near Drazinda and Moghalkot (D. I. Khan), Dera Ghazi Khan (30°08':70°32'), Chamlang (30°12':60°25'), Spin Tangi (29°55':68°05'), Khattan and Mamand (Mari–Bugti Hills), and Barkhan (29°52':69°30'). The estimated reserves of gypsum in Pakistan are well over 350 million tonnes and the annual production is over 400,000 tonnes (Shams and Khan 1987, Butt and Latif 1992).

Limestone

Pakistan is endowed with inexhaustible resources of limestone suitable for construction as well as industrial purposes (Ahmad 1969, Bender 1995). The entire Foreland fold-and-thrust belt contains extensive, thick to massive limestone beds in sequences ranging from Permian to the Miocene (see Chapter 5). Large production of limestone comes mainly from Kohat (Kohat Plateau), Nowshera (Attock–Cherat Range), Hasanabdal (Lesser Himalayan Ranges), Wah (Margalla Hills), Beth Kas–Khairabad–Daud Khel (Salt Range), Pezu (Marwat Range), Moghal Kot and Zindapir (Sulaiman Range), Rohri–Hyderabad–Thatta (Indus platform), and Jungshahi–Murli Hill–Mangho Pir (southern Kirthar Range). The average annual limestone production is 7–8 million tonnes (Butt and Latif 1992).

Magnesite

Magnesite occurs as alteration product of serpentinised ultramafic rocks or as replacement deposits in dolomite or dolomitic limestone. There are several occurrences in the ultramafic rocks of the ophiolitic thrust belt but these comprise small deposits or trivial showings. The stratabound replacement-type deposits in dolomitic sequences, however, are fewer but larger, and of greater economic value. The largest of these is located at

Table 9.12. Magnesite deposits: location, geology and potential.

Locality	Geological Setting	MgO%	Reserves (m.t.)
Skhakot (37°24':71°56')	Veins in serpentinised ultramafic rocks of the Dargai klippe (Indus Suture zone).	–	Very small
Spin Kan (Nasai) (30°47':68°06')	In serpentinised ultramafic rocks of Zhob ophiolitic thrust-belt.	43.38 to 45.4	60,000
Shabi Ghundi (30°48':68°00')	(as above)	38.04 to 42.36	6,000
Tlerai Mohd Jan (30°53':67°42')	(as above)	N. A.	Very small
Zizha (31°30':69°37')	(as above)	N. A.	Very small
Kakru (27°43':66°09')	Vein in serpentinised ultramafics of the Bela ophiolitic thrust-belt.	32.8	Very small
Loya Na Pani (27°15':66°20')	(as above)	32.84 to 44.56	Very small
Baran Lak (26°59':66°18')	(as above)	18.08	20,000
Sinchi Bent (26°30':66°21')	(as above)	N. A.	Very small
Sra Salawat (30°40':67°53')	In Eocene dolomite unconformably overlying the Zhob ophiolites.	46.49	16,000
Nal (27°41':66°11')	Replacement veins in limestone of Shirinab Formation, Bela ophiolitic thrust-belt.	N. A.	Very small
Sherwan (Kumhar)	Replacement deposits in dolomitic limestone of Cambrian Abbottabad Formation.	44.9 to 46.7	11,268,000

N.A.= Data not available. (From Gauhar 1966, Nagell 1969, Chemical Consultant 1970).

Sherwan near Abbottabad. The location, geological setting and potential of the magnesite deposits is summarised in Table 9.12.

Mica

Muscovite sheet-mica occurs in pegmatites at several localities (Heron 1954, Ahmad 1969, Kazmi et al. 1990). The mica sheets are rather small, largely less than 0.3 m in size. In the Karakoram block limited and sporadic mining has been done near Baltit (36°15':74°25'), Dassu (35°20':73°20'), Mogh (32 km NE of Chitral), and Kasu (35°34':72°52') NE of Drosh. In the Kohistan island arc, mica has been worked at Khadong Banda (near Dir). In the Himalayan crystalline zone, mica deposits have been reported from Astor, Bagarian (34°33':73°10'), Hawa Gali (34°29':73°06'), Rajdhawari (34°00':73°06') and in the Neelam Valley of Azad Kashmir. Amongst these the better deposits are the ones in Neelam Valley and near Astor.

Nepheline Syenite

Nepheline syenite crops out extensively south of Koga village, 56 km NE of Mardan and forms a part of the Ambela alkaline magmatic complex (Chapter 6). The alkaline composition and significant alumina content makes this rock a good substitute for feldspar in the glass and ceramic industry (Mikrch 1976). In Russia such rocks have been used for extraction of alumina, manufacture of portland cement and soda ash. Pilot plant tests by the Sarhad Development Authority (SDA) and industrial-scale tests by the industry have established the feasibility of its large scale use in glass industry and ceramics. The Koga nepheline syenite contains SiO_2 59.92–61.65%, Al_2O_3 20.35–23.46%, Fe_2O_3 1.7–3.13%, Na_2O 6.12–10.68%, and K_2O 4.70–6.93%. Mechanical treatment reduces the iron content to less than 0.1%. The total reserves are estimated at about 6,000 million tonnes. Ore reserves of the material suitable for glass and ceramics is estimated at 82.78 million tonnes of the ore (Khan and Ahmad undated).

Ochre

Yellow and brown to reddish coloured ferruginous clays are of widespread occurrence along lateritic horizons in the sedimentary sequences of the Foreland fold-and-thrust belt (Heron et al. 1954, Ahmad 1969, Bender 1995). Deposits of economic value, some of which are being mined, occur in the Reshian region of Azad Kashmir and at several localities in the Salt Range e.g., Uchhali (32°32':72°02'), Kutki (32°59':71°21') and Pirkahar (32°39':72°43'). Small deposits occur in the southern part of the Kirthar Range near Jhal Dhand (24°52':67°56') and Sonhari Dhand (25°00':68°04').

Rock Phosphate

Phosphorite occurrences have been reported from the Cretaceous sequence near Kohat (Chichali Formation), near Chhoi in Kalachitta Range (Kawagarh Formation); from the Paleocene and Eocene rocks in the Salt Range (Patala Formation), in the Sulaiman Range (Khadro Formation and Ghazij Group), and in the Oligocene sequence in southern Kirthar Range (Nari Formation). These are of little or no economic value due to low P_2O_5 content (Ahmad 1969, Raza and Iqbal 1977). The main economic deposits of rock phosphate occur in the Cambrian Abbottabad Formation and the Precambrian Hazara Formation, northeast of Abbottabad, along the western flank of the Hazara–Kashmir Syntaxis (Hasan and Ghaznavi 1980, Khan and Ahmad 1991). The mineral constituents of the ore comprise collophane, minor francolite, glauconite, dolomite, iron oxide and pyrite. The deposits are of pelletal-type and spread over an area of 155 sq km. Location, chemical content and reserves of the main deposits in this region are given in Table 9.13. Some of these deposits are being mined and the reserves have been greatly depleted.

Table 9.13. Hazara Phosphate deposits.

Locality	P_2O_5%	Reserves (m.t.) (all categories)
Kakul–Mirpur (9.5 km from Abbottabad)	24–32	1.8
Kalue–de–Bandi and Lagarban (40 km NE of Abbottabad)	10–35	10.80
Dolola (6 km from Garhi Habibullah)	19.2	9.20
Sirban Hill (34°06':73°11')	10.15	1.90

(From Hasan and Ghaznavi 1980).

Rock Salt

Rock salt occurs extensively in the Precambrian and Eocene evaporite sequences in the Salt Range and the Kohat Plateau. Estimated reserves are well over 2 billion tonnes (Working Group on Minerals 1978), and the annual production is between 500,000 and 600,000 tonnes (Butt and Latif 1992). In the Salt Range, the salt-bearing evaporites occur in the Precambrian Salt Range Formation (Billianwala Salt Marl Member) and this unit crops out extensively along the southern faulted escarpment of the Salt Range (Asrarullah 1967, Alam et al. 1975). Several small salt mines and quarries are located along this escarpment, though most of the production comes from the large mining centres at Khewra (32°39':71°45'), Warcha (32°29':71°59') and Kalabagh (32°55':71°32').

In the Kohat Plateau, rock salt occurs in the Eocene Bahadur Khel Salt Formation which is exposed over an area of about 5,000 sq km in the Jatta–Bahadur Khel–Karak region. To evaluate the deposit, a 1,500 m deep exploratory hole was drilled at the Bahadur Khel Salt anticline (33°09':70°59') in 1963 without reaching the base of the salt (Rashid and Husain 1967).

Potassium salts are present in the rock salt and salt marl of the Billianwala Salt Member of the Salt Range Formation and occur as layers and lenses up to 4 m thick and 240–400 m wide. These lenticular bodies contain 7.83–9.4% K and the reserves at Khewra are estimated at 124,350 tonnes (Alam and Asrarullah 1973). Similar occurrence of potash salt is reported at Warcha and Kalabagh (Alam et al. 1975). Selective mining of these deposits is uneconomic owing to the low potash content and lenticular nature of the potash -bearing beds.

Soapstone and talc

Soapstone is found at several localities associated with the ultramafic complexes in the ophiolitic thrust-belts, the Indus suture zone (including the Dargai klippe) and the Shyok suture zone. It also occurs as veins and replacement bodies along bedding and shear zones in dolomitic rocks. Ophiolite-associated deposits have been reported from Parachinar area and the Zhob Valley. Deposits associated with dolomitic sequences occur at Derai (SE of Alpurai), Sherwan (40 km west of Abbottabad), near Landi Kotal and Jamrud, west of Peshawar (Ahmad 1969). The Derai deposit is in the Paleozoic metasedimentary sequence along the western margin of the Besham crystalline nappe and the proved, indicated and inferred reserves are 100,000; 390,000; and 1,000,000 tonnes respectively (Working Group on Minerals 1978). The Sherwan deposit in the Cambrian Abbottabad Formation has estimated reserves of 200,000 tonnes (Raza and Iqbal 1977). The total reserves of soapstone are estimated at over 1.6 million tonnes. The annual production is over 38,000 tonnes (Shams and Khan 1987).

Sulphur

The Oligocene Nari Formation contains a small sulphur deposit at Sanni (29°02':67°29') along the foot-hills of the Kirthar Range. Sulphur is in the form of veins and as replacement of the matrix of sandstone. The reserves are estimated at about 58,000 tonnes, containing 45% sulphur. The deposit has been mined intermittently (Muslim 1973a).

A much larger, solfataric-type sulphur deposit occurs at the extinct Koh-i-Sultan Volcano (Fig. 9.11). Large lenses and layers of massive sulphur are interbedded with Pleistocene tuffs. The reserves are estimated at about 738,000 tonnes of ore containing 50% sulphur (Muslim 1973b). The ore has been mined and the annual production varies between 300 and 1,300 tonnes (Butt and Latif 1992).

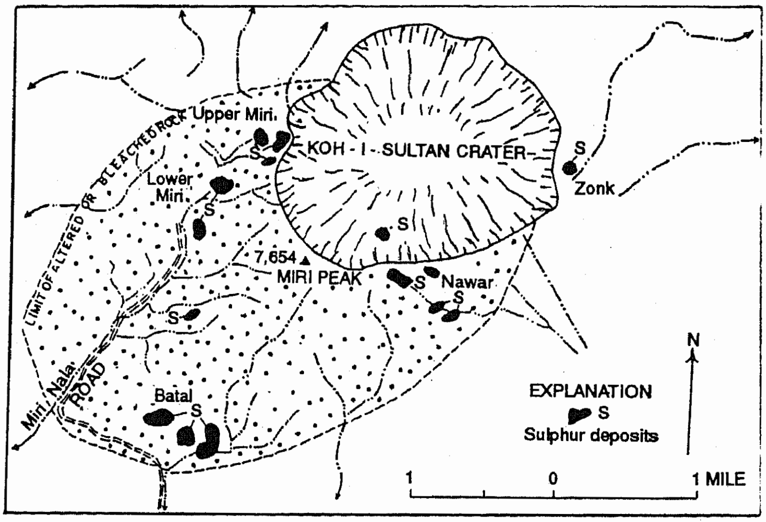

Fig. 9.11. Sketch map of Koh-i-Sultan sulphur deposits (from HSC 1960).

Vermiculite

Vermiculite is associated with ultramafic rocks and has been reported from the Indus suture zone, at Shinai Ghundai in Mohmand Agency NW of Peshawar and in Arin Valley, 56 km north of Saidu. The main deposits, however, are located in the western Ras Koh Range, in the Doki River area, 21 km south of Dalbandin. Hydrothermal alteration of ultramafic rocks has formed large deposits of vermiculite schist. Here the reserves of coarse-grained vermiculite with good bloating capability are estimated at about 11 million tonnes (Ahmad 1969).

10

Mineral Fuels

COAL

Coal has been mined for several decades in the territory that comprises Pakistan. After the discovery of natural gas in Bugti area (Balochistan), the main use of coal at present is in the brick industry and its role in the energy scenario has decreased from about 60% in late forties to about 5% at present. Coal mainly occurs in Paleocene and Eocene rocks, though one occurrence in the Late Permian rocks of the Salt Range is also known and graphitised coal deposits have been reported from the Permian sequence near Reshit (Chapursan Valley) north of Hunza (Faruqi 1997). The Indus Basin contains three main coal-bearing regions.* From north to south these are as follows:

-The Kohat-Potwar coal region which lies in the northern part of the Indus Basin.
-The Quetta–Harnai–Duki coal region which covers northeast Balochistan and forms a part of the Sulaiman stratigraphic province of the Indus Basin.
-The southern Sindh coal region which comprises the southern part of the Kirthar stratigraphic province of the Indus Basin.

Kohat – Potwar Coal Region

In the Kohat–Potwar region there are two main coal-fields, one located in the Surghar Range, known as the Makarwal coal-field and the other located in the central and eastern Salt Range (Fig. 10.1). Smaller deposits occur near Hangu, Cherat and Kotli.

The **Makarwal coal-field** covers an area of about 75 km^2 and is located 45 km south-west of Kalabagh (32°58':71°34'). Coal crops out in an elongate anticline with steep over-turned eastern limb and 15° to 50° dip on its western limb. A number of faults affect the coal-field and cause difficulties in mining. The main coal seam occurs at the base of Hangu Formation (Fig. 10.2) and varies from 0.3 to 3.0 m in thickness, with an average thickness of about 1.1 m (Warwick and Hussain 1990). The annual coal production from Makarwal has been about 225,000 t. (Kazmi and Abbas 1991). The coal resources of this field are estimated at 22 m.t. (Warwick and Hussain 1990). The Makarwal coal is of high-volatile B and C bituminous rank (Landis et al. 1971, Ahmed et al. 1986). According to Hasan (1990) the maceral contents of this coal comprise vitrinite (71%), inertinite (18.9%) and liptinite (9.8%). Proximate analyses of Makarwal coal are given in Table 10.1.

The **Salt Range coal-field** covers an area of approximately 1,500 km^2 in the central and eastern Salt Range. The northern limit of the coal-field is marked by the approximate areal extent of the Eocene limestones (Warwick and Hussain 1990). The coal occurs in the Patala Formation in a synclinal plateau bounded by eroded anticlines. A number of normal faults traverse the area forming small escarpments and cause repetition of the exposures of the Patala Formation which could be mined for coal. There are more than two coal-seams in the Salt Range but in most cases only one is mineable. The maximum

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* A coal region may be defined as a geologically and/or geographically distinct area in which coal bearing rocks are known or believed to be present. It includes two or more smaller entities that might be called "areas" or "fields."

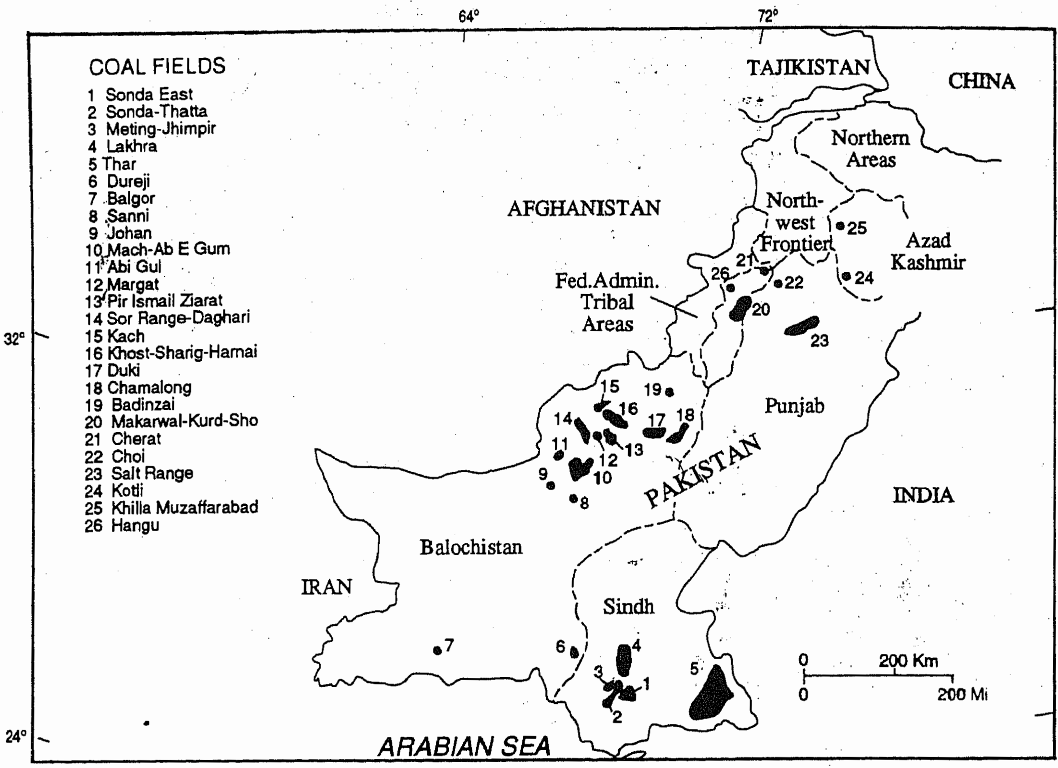

Fig. 10.1. Map showing coal-fields and coal occurrences in Pakistan.

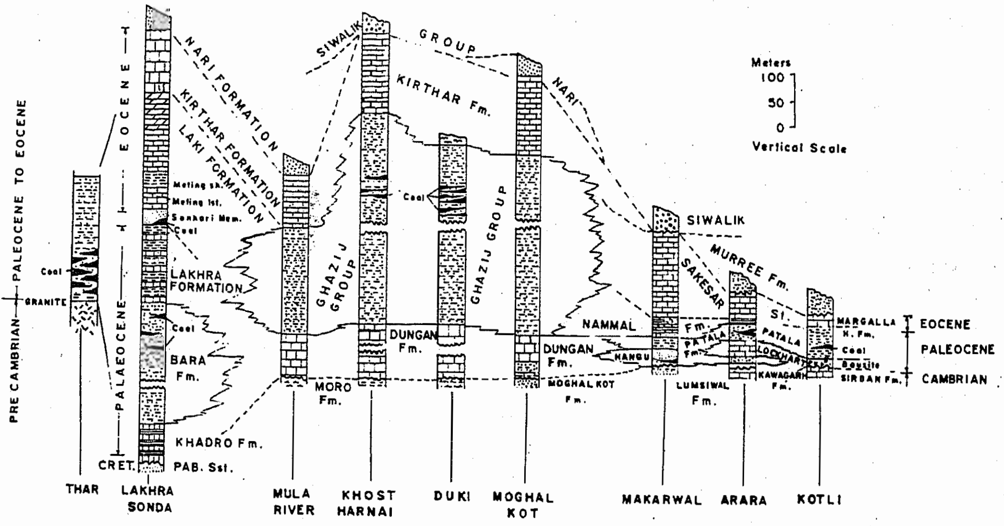

Fig. 10.2. Stratigraphic sections showing coal bearing formations in Pakistan.

thickness of the coal seam is about 2.13 m.

The rank of the Salt Range coal ranges from high-volatile C to B bituminous (Landis et al. 1971); its calorific value ranges from 4,182 to 5,267 K Cal/kg. According to Hasan (1990) the maceral contents of Salt Range coal comprise vitrinite (79.1%), inertinite (11.2%) and liptinite (9.7%). Proximate analyses of the coal are given in Table 10.1. The annual production of coal is about 225,000 m.t. and the coal resources are estimated to be about 235 m.t.

Minor coal occurrences are found in the **Hangu–Karak** area in a series of steeply dipping east-west trending parallel folds which plunge to the east. On the western side the open ends of the folds form escarpments with outcrop of the Hangu Formation which contains coal. This is likely to be a continuation of the Makarwal coal-field. Coal is being mined on a limited scale. Annual coal production from this area is about 26,500 tonnes (Gauhar 1988).

In the **Cherat** area, coal occurs in the folded Patala Formation. Exposed coal beds are 1.0 m in thickness. The coal has some lateral extent as indicated by a few bore holes. Limited data indicate that the calorific value of the coal is 13,000 BTU; it contains 3.0% ash and 0.5% total sulphur. The annual production from Cherat is about 500 tonnes (Gauhar 1988).

In the **Kotli** area the Patala Formation contains two coal beds. The lower bed commonly overlies bauxite or fire clay along the unconformity between Cambrian dolomite and the Palcocene sediments. The thickness of these beds ranges from 0.02 to 2.2 m (average 0.6). The coal laterally grades into carbonaceous shale. On an average the coal contains 27.94% ash, 48.4% fixed carbon and 2.14% sulphur. Recoverable reserves are estimated at 61,000 tonnes (Ahmad 1981, Shah and Bhutta 1988).

Quetta–Harnai–Duki Coal Region

This coal-bearing region covers an area of approximately 12,500 km^2 and extends for about 240 km westward from near Bahlol (30°01':69°28') to Quetta and then another 160 km southwards up to Johan (29°20':66°52'). In the Quetta–Harnai–Duki region, coal occurs in the Ghazij Group of Lower Eocene age. According to Kazmi (1962) and Shah (1990), the lower part of the Ghazij Group was deposited in a sub-littoral off-delta or pro-delta environment whereas the upper part (Toi Formation) formed largely in deltaic, lagoonal and fluviatile environment. Coal is mainly mined from Sor Range–Deghari, Pir Ismail Ziarat, Mach, Khost–Harnai and Duki areas. Coal deposits also occur in the Chamalong–Bahlol area, east of Duki and near Johan.

The **Sor Range–Deghari coal-field** is located about 10 km east and southeast of Quetta and comprises a northwest-southeast trending syncline. Along its eastern limb a large number of coal mines are clustered and are located in the upper part of the Ghazij Group. A prominent 12–21 m thick conglomerate bed occurs 30–50 m beneath the Kirthar limestone and serves as a marker horizon. The coal commonly occurs below this conglomerate bed. The rank of the coal varies from sub-bituminous B to sub-bituminous A (Landis et al. 1971). According to Ghaznavi (1988) this coal contains 83–89% vitrinite, 4.3–5.7% inertinite and 6.6–10.8% liptinite. The calorific value ranges from 4,831 to 6,060 K.Cal/kg with an average value of 5,644 K.Cal/kg (Warwick and Javed 1990). Proximate analyses of the coal are given in Table 10.1. Annual production from this field is about 460,000 t. About 50 m.t. of coal resources have been estimated by Shah (1990).

Table 10.1 Proximate analyses of Pakistan coals (from Kazmiand Abbas 1991, Fasset and Durrani 1994).

Coal field	Moisture %	Volatile %	Fixed Carbon %	Ash %	Sulphur %	Calorific Value Col/kg.
BALOCHISTAN						
Duki	3.7 - 4.81	32.77 - 38.37	46.49 - 50.82	9.11 - 16.03	4.31 - 6.15	2,826 - 6,748
Mach	7.1 - 12.0	34.5 - 39.4	32.4 - 41.5	9.6 - 20.3	3.2 - 7.4	5,104 - 5,717
Sharig- Harnai	1.53 - 7.9	30.29 - 43.69	28.49 - 53.14	6.33 - 30.34	2.8 - 12.63	4,173 - 7,531
Sor Range	5.1 - 21.2	31.0 - 43.1	36.0 - 43.0	2.7 - 14.3	0.4 - 5.6	4,822 - 6,049
PANJAB TRANS- INDUS						
Makarwal	2.8 - 5.3	42.4 - 48.1	36.7 - 44.9	6.4 - 11.5	2.8 - 6.8	6,328 - 6,769
PANJAB CIS- INDUS						
Salt Range	3.2 - 7.6	26.3 - 38.8	29.8 - 44.8	12.3 - 37.7	3.5 - 10.7	3,941 - 6,161
SINDH						
East of Indus	39.40 - 40.91	16.67 - 28.61	23.80 - 28.91	2.94 - 11.65	0.23 - 1.84	3,836 - 4,268
Jherruck (Sonda)	27.24 - 39.26	18.07 - 31.07	20.12 - 44.51	4.88 - 29.27	0.52 - 4.89	3,003 - 5,488
Jhimpir-Meting	15.4 - 29.8	29.8 - 39.9	31.0 - 36.3	8.2 - 14.6	3.4 - 7.4	4,106 - 5,439
Lakhra (north Central, South)	26.4 - 33.3	22.4 - 39.0	22.3 - 35.7	6.7 - 26.6	2.9 - 9.0	2,840 - 4,990
Sonda	7.0 - 26.98	15.94 - 48.1	12.58 - 64.9	3.96 - 39.2	0.56 - 8.25	3.646 - 6,957
Thar	44.05 - 52.55	54.66 - 63.62	19.34 - 20.38	5.53 - 13.4	0.57 - 3.6	2,677 - 3,151

The **Pir Ismail Ziarat** coal-field is relatively small (20 km^2 area) and is situated 60 km southeast of Quetta and about 15 km north of Mach (29° 52':67° 51'). It is comprised of a north-south trending anticline with resistant Jurassic to Paleocene limestone exposed in the core. The south-western part of the coal-field has been cut off by a large norhtwest-southeast trending thrust fault. There are two coal seams; the upper one is 0.6 m–0.7 m thick whereas the lower one is 0.4–0.45 m thick. The coal is high volatile bituminous C in rank and its moisture content ranges from 5.2% to 10%. Its heating value varies from 5,353 to 5,939 K.Cal/kg. The annual production of coal is about 115,000 t. The coal resources have been estimated at 11.0 m.t. (Ahmed et al. 1986).

The **Mach coal-field** is spread over an area of about 45 km^2 around the town of Mach. Here the rocks of the Ghazij Group occur in a large synclinal structure the greater part of which is covered by the alluvial valley-fill. There are several coal seams ranging in thickness from 0.3–1.5 m though only 3 or 4 seams (average thickness 0.75 m) are workable. The coal is largely blocky and ranks from sub-bituminous C to sub-bituminous B. Its heating value ranges from 5,100 to 5,730 K.Cal/kg (Ahmed et al. 1986). Proximate analysis is given in Table 10.1. The Mach coal is high in vitrinites (87.8%), low in inertinites (2.6%) and liptinites (9.6%). The average annual coal production from Mach is about 125,000 t. The resources of this coal-field are estimated at about 23 m.t. (Ahmed et al. 1986).

The **Khost–Harnai coal-field** is the largest in Balochistan. It is located about 160 km southeast of Quetta on the eastern flank of the Zarghun molasse basin. The Ghazij Group is exposed along the eastern margin of this basin and contains coal in the upper part of the Group. Coal mines are dotted along the ridge in a narrow belt, about 40 km long, and extend from northwest of Khost (29° 12':67° 05') to a few kilometres southeast of Harnai. The Khost–Harnai coal ranges in rank from lignite to bituminous B. The average rank is sub-bituminous C to A. Its heating value ranges from 4,420 to 7,000 K.Cal/kg. It contains about 83% vitrinites, 1.8% to 4.7% inertinites and 11.6% to 15% liptinites (Ghaznavi 1988). Chemical analysis of the coal is given in Table 10.1. The average annual production of coal from this field is about 100,000 t. The total coal resources have been estimated at about 76 m.t.

The **Duki coal-field** is near the town of Duki (39° 09':68° 34') in Loralai District, about 320 km east of Quetta. It comprises a moderately dipping, east-west trending, 30 km long and 5 to 10 km wide asymmetrical syncline. Rocks of the Ghazij Group crop out in the syncline, though most of the central part is covered with alluvium. Coal is being mined on both limbs of the syncline. There are 17 coal seams out of which 15 are being worked. The Duki coal is high sub-bituminous B to C, though sub-bituminous A type and even lignite are present (Khan et al. 1987). The heating value ranges from 4,610 to 6,380 K.Cal/kg (Ahmed et al. 1986). Its maceral composition is comprised of 84.4% to 87.4% vitrinites, 2.4% to 3.2% inertinites and 6.4% to 12.8% liptinites (Ghaznavi 1988). Proximate analysis of Duki coal is given in Table 10.1. The average annual production of coal from Duki is around 25,000 t. The coal resources of this field are estimated at about 50 m.t.

Lower Sindh Coal Region

A number of coal basins occur in the Kirthar Province of the Lower Indus Basin in Sind. These basins extend westward from near Chachro (25° 40':70° 15') in the Thar Desert through Badin (24° 40':68° 50') to Lakhra–Sonda–Thatta area (Fig. 10.1). The western part of this zone (west of the Indus) falls in the fold belt zone, whereas most of the eastern part covers the platform slope. Platform shelf and carbonate deposits ranging in age from Triassic to Recent overlie the basement slope. The rock formations dip gently eastward though disrupted by a series of deep-seated normal faults. In the fold belt region, these rocks form gently dipping folds.

Presently four coal-fields are known which have been explored, namely Lakhra, Sonda–Thatta, Meting and Thar coal-fields. The extent of these coal-fields and the coal resources have been delineated through a large number of exploratory holes (Kazmi et al. 1990, Fasset et al. 1994). The coal occurs in the Middle Paleocene Bara Formation and in the Sonhari Member of the Early Eocene Laki Formation. The Sonhari coal is restricted to the Meting coal-field. The Sonhari Member is about 30 m thick and consists of sandstone and lateritic clay. Outerbridge and others (1989) suggest that the Sonhari was deposited in coastal environment; the vegetation in shallow, brackish-water swamps formed this coal.

The coal in the Lakhra and Sonda–Thatta coal-fields is in the Bara Formation which is conformably overlain by the Lakhra Formation (Fig. 10.2). The latter consists of fossiliferous limestone, sandstone, siltstone and claystone and it is 40 to 150 m thick. The Bara Formation is largely comprised of sandstone with subordinate claystone and siltstone. It varies from 60 to 600 m in thickness. It is predominantly marine but also contains lacustrine, estuarine, deltaic and lagoonal deposits with plant fossils and carbonaceous beds at several horizons (Kazmi et al. 1990). The Bara Formation contains two main coal-bearing horizons, one in the upper part and the other in the lower part. The lower one, named the Jherruck coal zone, is presently known only in the Sonda-Thatta coal-field. According to Ahmed et al. (1986) the Bara coal was formed in limnic back barrier or deltaic depositional environment, though Sanfilipo et al. (1990) suggest deposition in low relief coastal setting with eustatic sea level changes controlling peat forming coastal swamps.

The **Lakhra coal-field** is located about 20 to 25 km northwest of Hyderabad and covers an area of approximately 300 km^2. Structurally the coal-field comprises a broad gently folded anticline (dip ± 5°), which has been dissected by several normal faults (1.5 to 9 m displacement). Coal occurs in the Bara Formation 50 to 150 m below the surface. A number of coal seams have been encountered in bore holes. Nine of these are more persistent (Schweinfurth and Hussain 1988) and three of these are of greater economic significance. In descending order these are known as Dhanwari, Lalilian and Kath coal seams (Fig. 10.3). The Lakhra coal is dull black. It contains amber and resin flakes, and clay and gypsum partings. It tends to crumble on long exposure and is susceptible to spontaneous combustion. Petrographic studies (Ghaznavi 1988) show that the Lakhra coal contains 54.3 to 77.4% huminite, 12.4% to 26.7% inertinite and 0 to 27% liptinite macerals. The rank of the coal varies from lignite A to sub-bituminous C and its heating value ranges from 2,570 to 4,260 K.Cal/kg. Information concerning coal quality is given in Table 10.1.

The average annual production of coal from Lakhra is about 1.5–2.0 m.t.* Based on data from more than 244 drill holes, the coal resources of Lakhra have been estimated to be about 1,592 m.t. (Table 10.2).

The **Sonda coal-field** is located south of the Lakhra coal field, between the towns of Thatta and Sonda, east of Karachi. It covers an area of more than 1,400 km^2 and extends eastward across the Indus up to Tando Mohammad Khan and beyond. Its trans-Indus extension has been referred to by some workers as Indus East coal-field (Thomas and Khan 1990).

The Sonda coal-field comprises a nearly flat (dip 2° or less), north-south trending anticline (Schweinfurth and Husain 1988). Coal occurs in thick lenses of siltstone and mudstone of the Bara Formation. More than 29 coal seams of varying thickness and continuity have been encountered in various test holes. Coal occurs relatively more persistently at three main horizons (Ahmad et al. 1986) which have been named Dadhuri, Sonda and Jherruck coal zones (Fig. 10.3). In between these coal zones four other (but minor and less persistent) zones, referred to as "Upper Strays", Inayatabad, "W" and "Lower Strays", have been identified (Schweinfurth and Husain 1988). These minor coal zones consist of one or more recognisable coal beds separated by relatively persistent parting of sandstone or shale or both. The Sonda coal zone contains the thickest and most persistent coal bed, though at places it splits into thinner seams and becomes discontinuous or changes to carbonaceous shale. Near Jherruck, the cumulative thickness of coal reaches 9 m, and the Sonda coal seam attains a thickness of 6.2 m (Kazmi et al. 1990).

According to Ahmed et al. (1986), Sonda coal was deposited on a lower deltaic plain. Sanfilipo et al. (1990) have pointed out that the clean sandstones associated with the coal zone lack obvious scouring and contain abundant glauconite, burrows and marine

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*Verbal communication Mr.Bashir Ansari, Managing Director, Lakhra Coal Development Company.

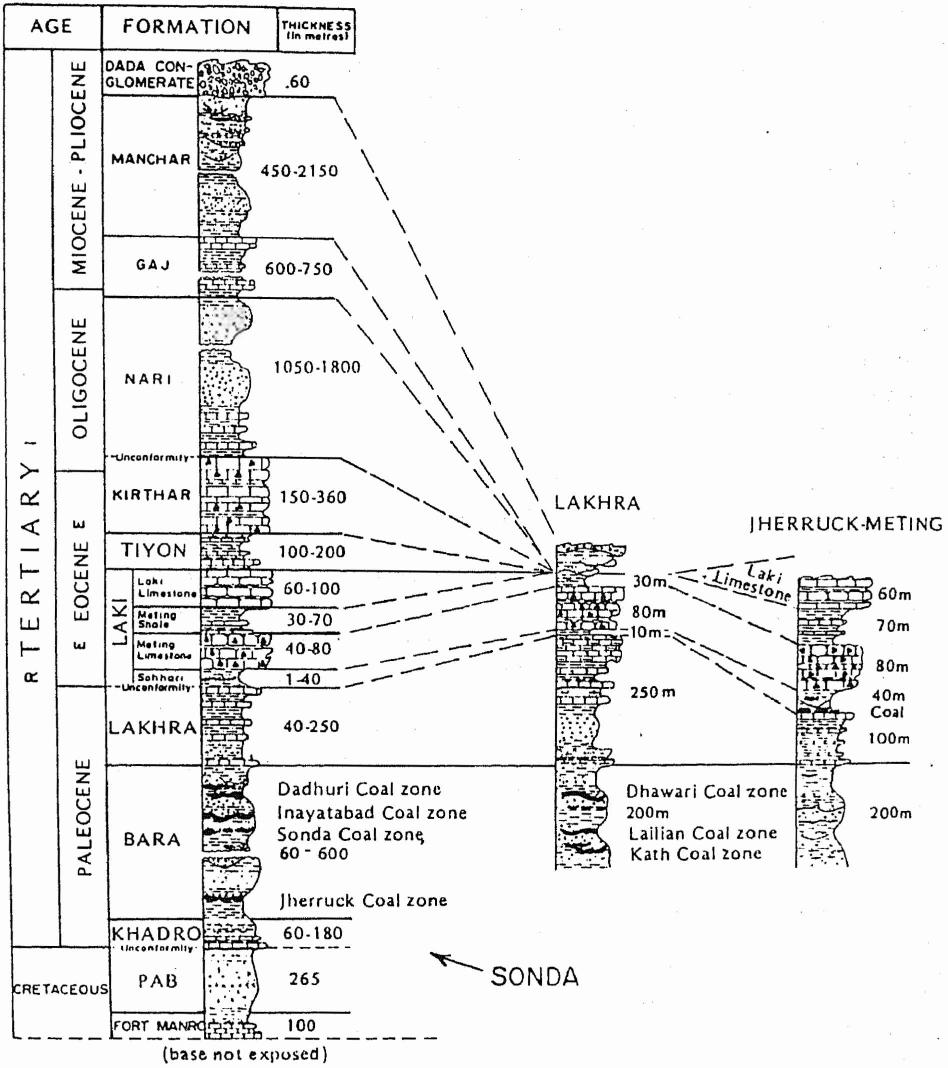


Fig. 10.3. Stratigraphic sections of Lakhra, Sonda and Meting coal-fields (from Kazmi et al. 1990).

fossils. This feature suggests marginal marine conditions for the Bara Formation. Kazmi et al. (1990) have shown that the coal was formed in near shore paleo-channels and swamps. Coal petrology data (high ulminite, textinite and well preserved homodetrinite) confirm this view (Hasan 1990).

The Sonda coal ranks from lignite A to sub-bituminous C (Schweinfurth and Husain 1988). Its heating value ranges from 3,600 to 5,700 K.Cal/kg (Ahmed et al. 1986). Proximate analysis of the coal is given in Table 10.1. Based on subsurface data from bore holes the coal resources of Sonda coal-field have been estimated at about 7,300 m.t.(Table 10.2).

The eastern extension of the coal-field has not yet been explored. The coal resources of Sonda are thus likely to be much higher.

The **Meting coal-field** is relatively small and covers an area of about 90 km^2 east of the railway line between Meting and Jhimpir (25° 01':68° 01') railway stations. The coal is in the Sonhari Member of the Laki Formation of Eocene age, near the contact with the Upper Paleocene Lakhra Formation (Fig. 10.3). There is only one workable coal bed which is commonly thin and lenticular and ranges in thickness from 0.3 m to 1.0 m with an average of 0.5 m. The Meting coal ranges from high volatile bituminous C to B. Its proximate analysis is given in Table 10.1.

The average annual production of coal is about 40,000 t. The total coal reserves of Meting based on a cut off thickness of 0.6 m and depth of 50 m is about 161 m.t. (Kazmi and Abbas 1991).

The **Thar coal-field** is the largest in Pakistan and is located on the Indus Platform, in the Thar Desert in southeastern corner of Pakistan. It covers an area of about 9,000 km^2. The field is covered by stabilised longitudinal sand dunes. Test drilling data shows that the aeoline sand is up to 250 m thick. It is underlain by a 250 to 350 m thick Paleocene to Eocene coal bearing sedimentary sequence. The Nagar Parkar Granite underlies the Paleocene-Eocene sediments at shallow depths. The coal-bearing sequence contains several coal seams of varying thickness. The coal seams occur at depths ranging from 123 to 245 m. Some of the seams are more than 10 m thick, the largest ones being more than 20 m in thickness (Figs 10.4 and 10.5). The measured coal resources of this field are about 78 billion tonnes. The coal is lignite B in rank with average heating value of 5,333 Btu, 1.57% sulphur, 8.83% ash, and 48.57% moisture. The average dry and ash-free heating value of Thar coal is 12,322 Btu (Fassett and Durrani 1994).

With the recent discovery of the vast Thar coal-field, Pakistan's measured reserves are now 87.5 billion tonnes, eleventh largest in the world (Table 10.2).

PETROLEUM AND NATURAL GAS

Petroleum has been known to occur in the Kohat area since 1833. An 1870 survey for petroleum in the Punjab reported many oil seepages (Heron et al. 1954). Since those early times several oil and gas seepages have come to light from various localities in the NWFP, Punjab, Sindh and Balochistan (Fig. 10.6). In the early days these oil shows were considered important guides to locate economic oil deposits. Nowadays, though they have lost their importance, they are still of interest to companies in search for oil.

The history of exploratory oil drilling goes back to 1866 when the first well was drilled at Kundal in Khisor Range (NWFP). In 1884 it was followed by borings near Khattan (29° 34':68° 31') in Balochistan, which went down to depths of 20 to 130 m. A small quantity of oil was produced and in 1889 it amounted to 218,490 gallons (Heron et al. 1954). The first commercial oil discovery, however, was in 1920, when the Attock Oil Company discovered the Khaur oil field in the Potwar Plateau. By the time of Independence (1947), three more oil fields (Dhulian, Joya Mair and Balkassar) were in production. In the early fifties, the average annual production of oil from these sources was about 1.158 million barrels (approximately 3,173 barrel/day). Since then many oil and gas fields have been discovered (Fig. 10.7). Until 1997 about 431 exploratory wells have been drilled resulting in 115 discoeries, 51 of oil and 64 of gas and condensate. These oil fields are, however, small and their production ranges from less than 50 to about 18,000 barrels/day.

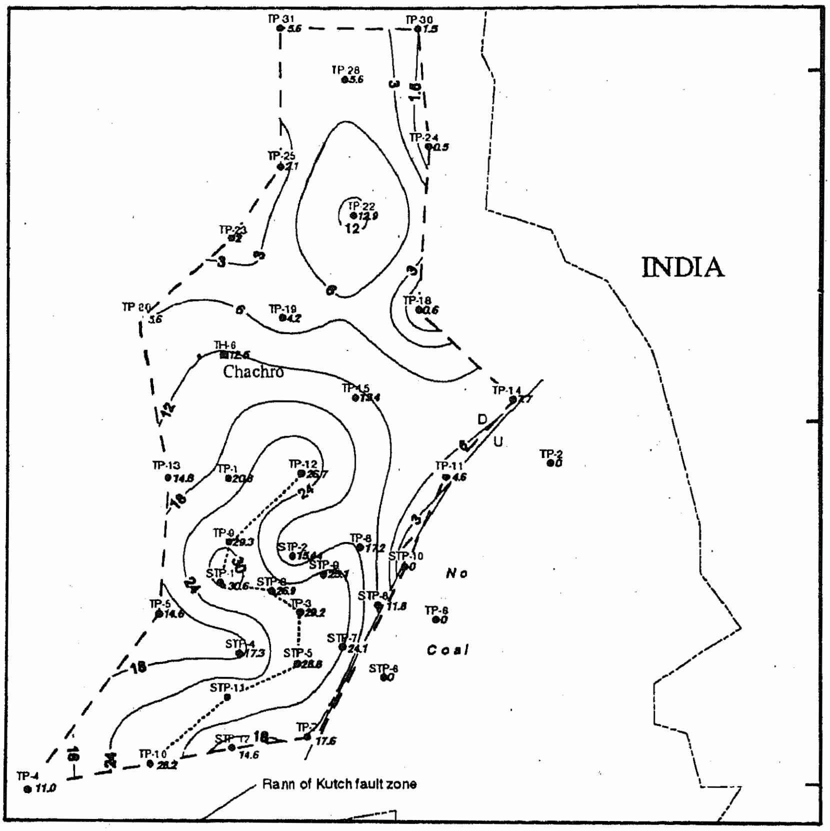


Fig. 10.4. Isopach map of total coal thickness in the Thar coal field. Contour interval in metres. Black circles show drill holes and numbers indicate cumulative thickness of coal seams (from Fasset and Durrani 1994).

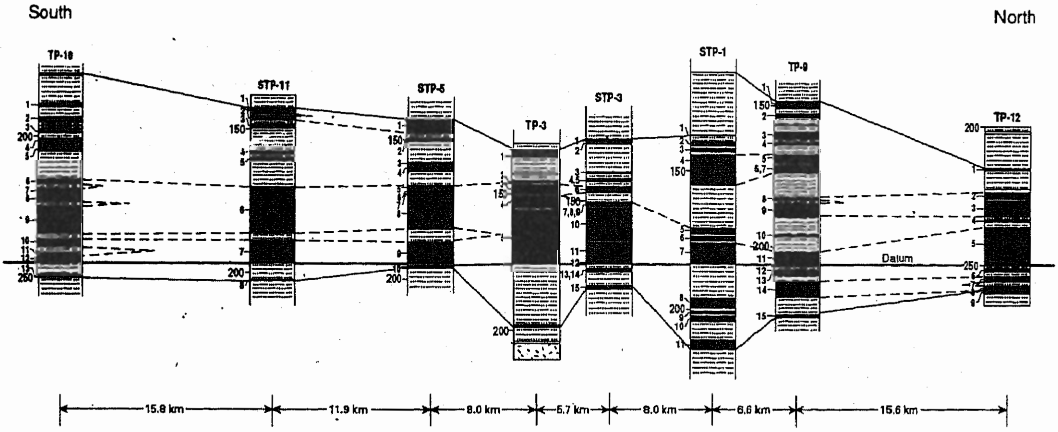


Fig. 10.5. Coal correlation section through the Thar coal field (from Fassett and Durrani 1994).

Table 10.2. Summary of coal resources of Pakistan (in million tonnes) .

Coal field	Average mineable thickness	Proved	Indi-cated	Inferred	Hypothe-ical.	Total	Coal Rank (ASTM)	Annual Production.
BALOCHISTAN								
1. Duki	0.5	14	11	25	—	50	SubC-SubA	0.250
2. Mach- Abegum	>0.75	9	—	14	—	23	SubC-SubB	0.125
3. Sor Range-Daghari	0.95	15	—	19	—	34	SubB-SubA	0.460
4. Pir Ismail Ziarat	0.5	1.5	1.5	8	—	11	hvCb	0.115
5. Khost-Sharig-Harnai	0.75	13	—	63	—	76	hvBb-hvAb	0.100
PANJAB								
6. Makarwal	0.75	5	8	9	—	22	hvCb-hvBb	0.225
7. Salt Range	0.5	43	13	—	178	234	hvCb	0.225
SINDH								
8. Lakhra	1.5	244	629	739	28	1640	LigA-SubC	2.00
9. Sonda		188	1388	5724	—	7300	LigA-SubB	—
10. Jhimpir	.5	10	43	108	—	161	LigA-SubC	0.040
11. Thar		—	—	—	—	78,196	— —	—
Total		542.5	2093.5	6709	206	87747	—	3.54

(modified from Kazmi and Abbas 1991).

Four of the gas fields (Sui. Mari, Uch and Qadirpur) are giants and two majors (Pir Koh and Khairpur). According to Raza (1997) the total petroleum and natural gas reserves in Pakistan are estimated at about 0.6 billion US Barrels and 31.0 trillion cu ft respectively. Recently a large gas field has been discovered at Bhit in the Kirthar Piedmont zone and the prospects are that it may proe to be a major (Aslam 1997).

Sedimentary basins and oil and gas fields

Inasmuch as the sedimentary formations are the natural habitat for oil and gas, the sedimentary basins are of particular significance from the standpoint of the occurrence, exploration and development of these deposits. As mentioned earlier (Chapter 5), Pakistan comprises two sedimentary basins– the Balochistan and Indus Basins. Based on tectonostratigraphic considerations these basins have been divided into various petroleum zones by Raza et al. (1989a) as shown in Figure 5.1.

Balochistan Basin

The Balochistan Basin, which comprises the Makran accretionary zone and the Makran offshore trench, contains 5,000–15,000 m thick flysch-type terrigenous slope and shelf sediments and turbidites (Chapter 5). This is the least explored region of Pakistan. Very limited seismic survey has been done and only six wells have been drilled in this vast basin of over 300,00 km^2. Despite several gas shows along the Makran Coast, no commercial hydrocarbons have been found in the basin so far.

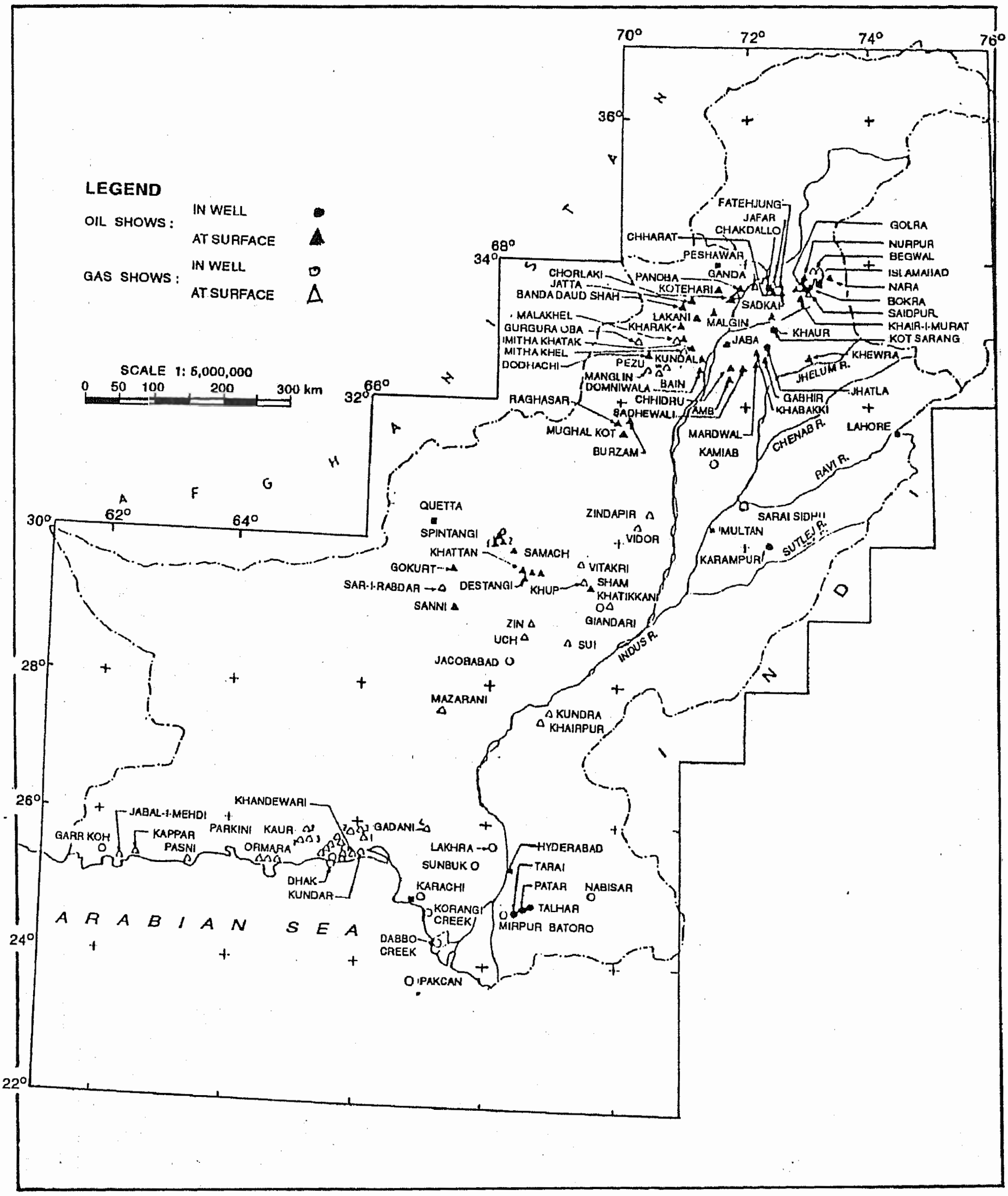


Fig. 10.6. Map showing gas and oil seepages in Pakistan (from Raza et al. 1989).

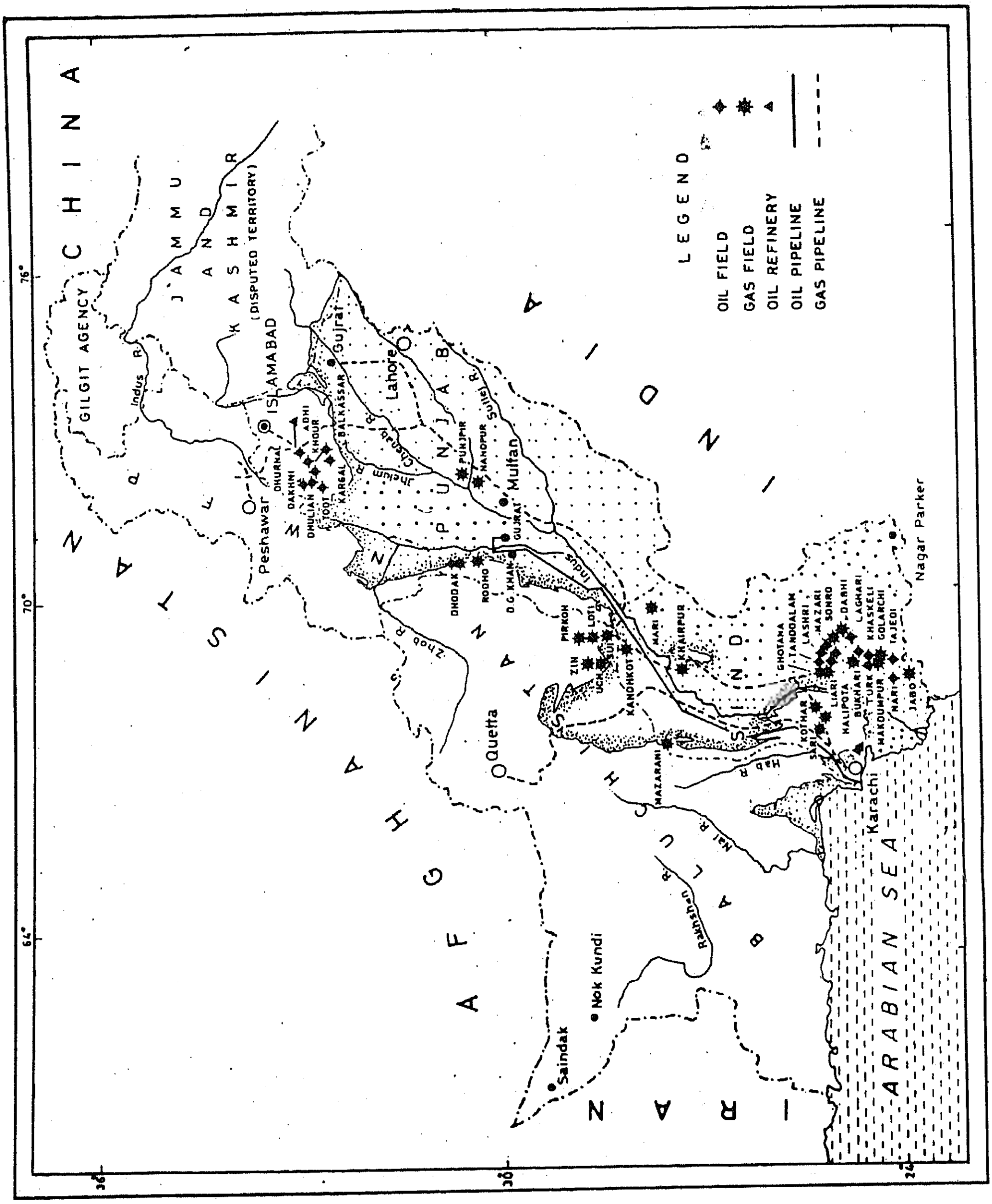


Fig. 10.7. Map showing the major oil and gas-fields of Pakistan (modified from Shams and Khan 1987, Kadri 1995).

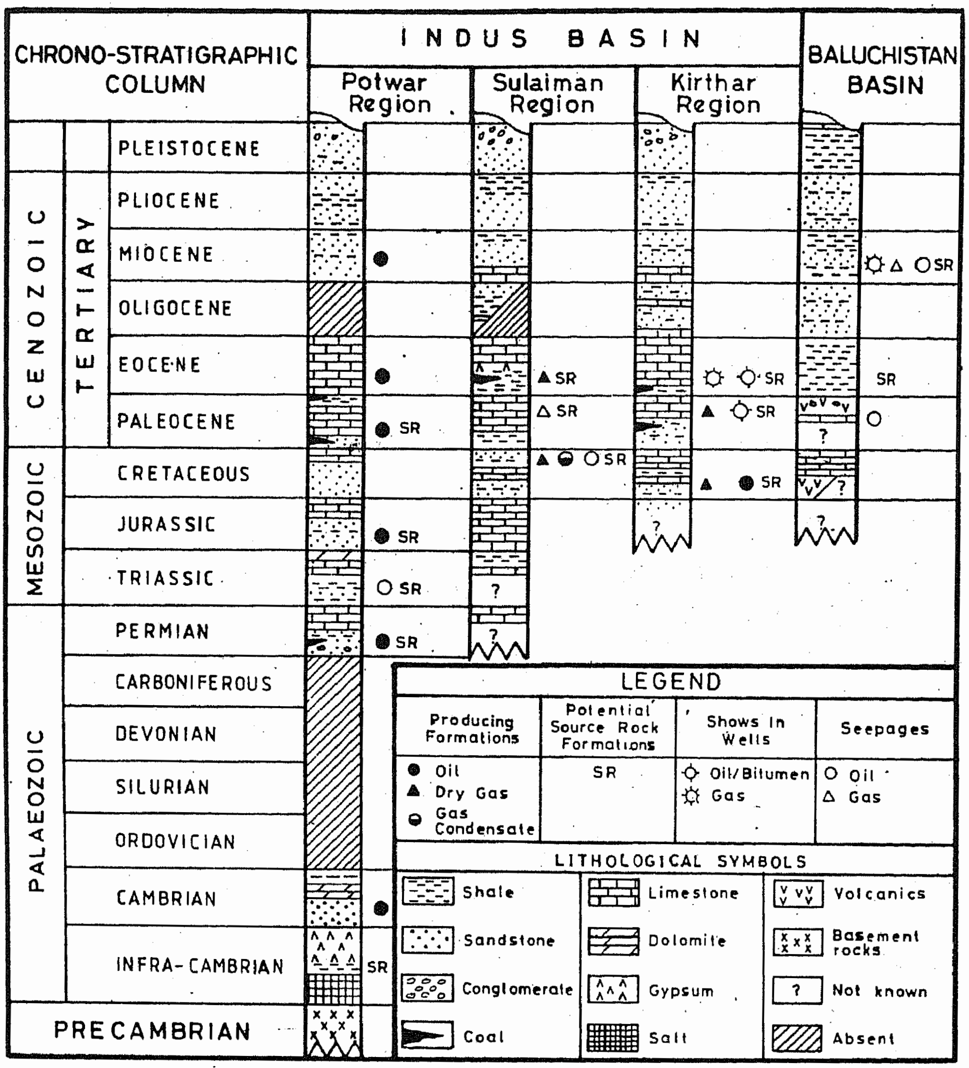


Fig. 10.8. Generalised stratigraphic columns showing occurrence of hydrocarbons in Pakistan (from Khan and Raza 1986).

Attractive structures comprising long, asymmetrical to overturned, parallel to en echelon anticlines and faulted thrust anticlines occur in the Makran coast and offshore area (Fig. 4.43). Extensive source rocks (Parkini, Panjgur and Siahan Formations), suitable reservoir rocks (Talar and Panjgur sandstones) and appropriate cap rocks have been identified in the basin (Malik et al. 1988, Raza and Ahmad 1990). Data on maturity, however, are meagre. According to Khan and Raza 1986, the geothermal gradient seems to be relatively low (1.5° to 2.5°C/100 m). Presently the hydrocarbon prospects in the Makran Basin are rated

Table 10.3. Oil and gas reseres as on June 30, 1993 (data from Director General Petroleum Concessions, Govt. of Pakistan, in Kadri 1995).

	Non-associated Gas fields	Original Recoverable Reserves	Balance Recoverable Reserves
1.	Adhi	0.11600	0.09208
2.	Kandhkot	0.78300	0.65284
3.	Khairpur	1.00000	1.00000
4.	Mazarani	0.01859	0.01859
5.	Sui	8.62400	3.17474
6.	Mari	6.30000	5.08937
7.	Bhal Syedan	0.00330	0.00278
8.	Bobi	0.04224	0.04102
9.	Buzdar	0.00810	0.00810
10.	Dakhni	0.25500	0.22863
11.	Daru	0.01301	0.01301
12.	Dhodak	0.58140	0.58140
13.	Hundi	0.05940	0.04308
14.	Jandran	0.08230	0.08230
15.	Kothar	0.01180	0.01160
16.	Loti	0.27695	0.21573
17.	Nandpur	0.29600	0.29600
18.	Nur	0.00608	0.00608
19.	Panjpir	0.03350	0.03350
20.	Pirkoh	1.80000	1.39816
21.	Qadirpur	3.97873	3.97873
22.	Rodho	0.10300	0.01300
23.	Sari	0.03900	0.02530
24.	Uch	4.05000	4.05000
25.	Zin	0.10000	0.10000
26.	Ratana	0.35000	0.35000
27.	Bhatti	0.03494	0.03487
28.	Bukhari	0.06794	0.04959
29.	Dabhi	0.01631	0.01258
30.	Dabhi South	0.00372	0.00372
31.	Golarchi	0.05536	0.01900
32.	Halipota	0.00218	0.00165
33.	Jabo	0.00280	0.00280
34.	Kato	0.00474	0.00474
35.	Khorewah	0.09975	0.09958
36.	Koli	0.01483	0.01483
37.	Mahi	0.01308	0.01308
38.	Matli	0.05555	0.01994
39.	Mukhdumpur	0.02366	0.01901
40.	Nakurji	0.02565	0.02565
41.	Nari	0.00727	0.00200
42.	Pir	0.00142	0.00142
43.	Rind	0.00150	0.00150
44.	Sonro	0.01784	0.01133
45.	Tando Ghulam Ali	0.00390	0.00390
46.	Turk	0.11356	0.03205
47.	Turk Deep	0.03171	0.03171
48.	Kadanwari	0.72800	0.72800
	ASSOCIATED GASES*	0.56153	0.18063
TOTAL: TCF		30.72864	22.81960
	Million TOE	*578.9*	*406.94*

	Field	Original Recoverable Reserves	Balance Recoverable Reserves
1.	Khaur	4.310	0.1260
2.	Dhulian	41.400	0.0400
3.	Joyamair	10.450	3.6680
4.	Meyal	42.500	7.2080
5.	Toot	15.800	4.3460
6.	Dhodak	16.200	16.2000
7.	Fimkassar	30.000	25.0880
8.	Dakhni	12.440	11.2100
9.	Tando Alam	20.160	11.1610
10.	Ghotana	0.400	0.2340
11.	Chak-Naurang	4.700	2.5250
12.	Lashari South	0.230	0.2300
13.	Thora	11.210	3.3420
14.	Sono	8.650	4.8270
15.	Lashari Centre	5.270	2.0200
16.	Bobi	9.760	9.0220
17.	Kunar	14.020	13.2620
18.	Daru	0.260	0.2600
19.	Pasakhi	9.370	4.3980
20.	Bhal Seydan	0.198	0.1290
21.	Lashari East[1]		
22.	Buzdar	0.081	0.0810
23.	Missakaswal	34.730	32.3360
24.	Dhamraki	1.370	1.3700
25.	Sadkal	0.738	0.7380
26.	Buzdar North	0.210	0.2020
27.	Meyun Ismail[1]		
28.	Khaskheli	8.196	0.1180
29.	Laghari	20.261	1.8860
30.	Dabhi	4.445	1.2840
31.	Tajedi	0.464	0.4640
32.	Golarchi	0.196	0.0600
33.	Nari	0.399	0.3670
34.	Turk & Turk deep	1.322	0.4830
35.	Mazari	14.745	1.6580
36.	South Mazari	10.902	5.5580
37.	Sonro	0.950	0.5940
38.	Halipota	0.432	0.1460
39.	Liari	5.429	1.4760
40.	Ghungro	0.787	0.7870
41.	Duphri	0.105	0.1050
42.	Matli	0.311	0.0630
43.	North Akri	1.487	1.2910
44.	Paniro	0.197	0.1970
45.	Dabhi South	0.047	0.0470
46.	Bhatti	0.617	0.5450
47.	Bukhari	1.651	1.2370
48.	Jabo	0.014	0.0140
49.	Kato	0.141	0.1410
50.	Khorewah	0.698	0.6960
51.	Kotli	0.179	0.1790
52.	Mukhdumpur	0.237	0.1770
53.	Rind	0.062	0.0620
54.	Mahi	0.209	0.2090
55.	Bari	2.531	2.4100
56.	Pir	0.007	0.0070
57.	Dhurnal	50.940	7.0840
58.	Balkassar[2]	34.015	1.5220
59.	Ratana	12.000	12.0000
60.	Bhangali	1.840	0.1200
61.	Adhi	10.221	6.6200
Total: Million Barrels		480.494	203.630
	Million TOE	*64.46*	*27.32*

as average, though the Makran Coast is likely to hold better prospects (Raza et al. 1989b).

Indus Basin

The Indus Basin covers an area of about 533,500 km^2 and contains more than 15,000 m thick sediments ranging in age from the Precambrian to Recent (Fig. 10.8). Oil and gas fields have been discovered in the inner folded zones of the Sulaiman and Kirthar Ranges, Kohat–Potwar Plateau, Sulaiman–Kirthar depression (foredeep), Karachi depression, and the Indus platform (Punjab monocline, Sukkur rift and Sindh monocline). The Jacobabad–Khairpur and Mari–Khandkot highs (Sukkur rift zone) and the Sargodha high (Sargodha–Shahpur buried ridge) divide the Indus Basin into three main tectonostratigraphic zones (Fig. 5.1).

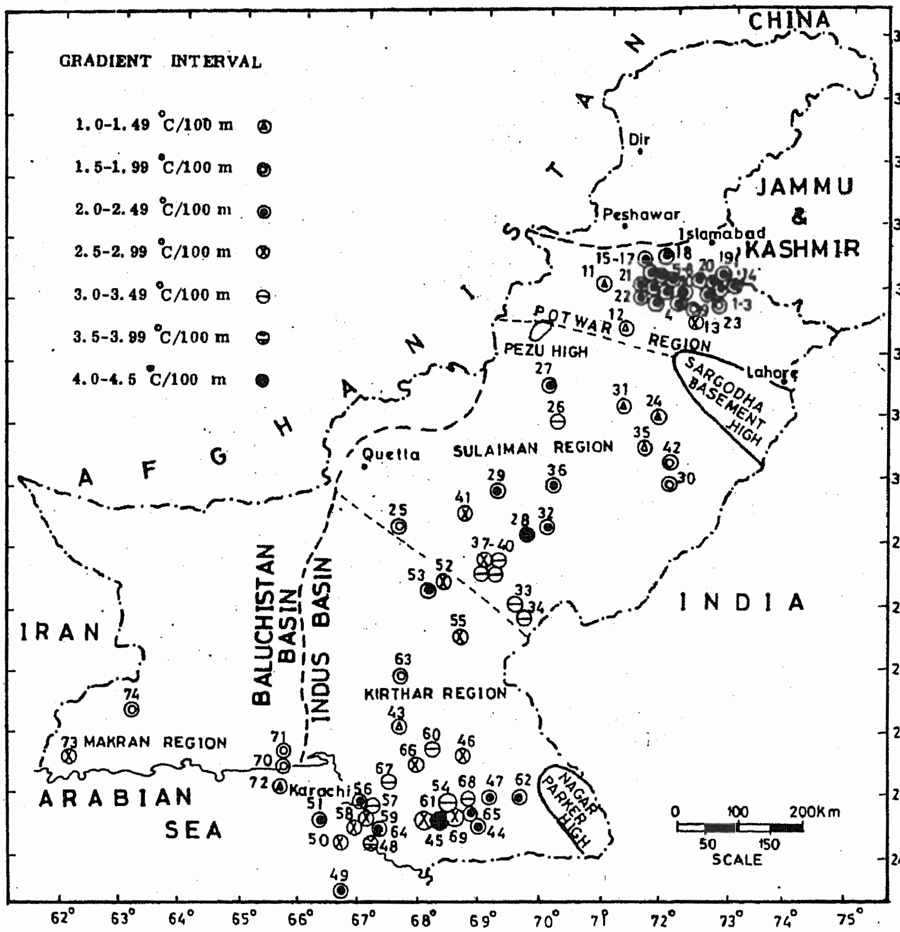
Northward, the region between the Sargodha high and the MBT forms the northern zone which includes the Kohat–Potwar Plateau, the Bannu Basin, the Cis- and Trans-Indus Salt Range, and the northern Punjab monocline. The region between the Sukkur rift zone and the Sargodha high, comprising the Sulaiman fold belt and foredeep and the southern Punjab monocline constitutes the central part of the Indus Basin. The southern zone lies to the south of the Mari–Khandkot horst and comprises the Jacobabad–Khairpur High, Kirthar fold belt and its depression and the Sindh monocline (Raza et al. 1989a, Raza and Ahmed 1990, Kadri 1995).

Northern Indus Basin: This basin is characterised by complex structural styles (Chapter 4) and a stratigraphic sequence ranging from Precambrian to Recent (Fig. 10.8). A number of oilfields occur in this zone (Fig. 10.7). The Dhurnal oilfield is the largest and has reserves of about 52 million barrels of oil and 0.13 TCF of gas. The main oil and gas producing horizons, possible source and cap rocks have been shown in Figure 10.8 and Table 10.4. Geothermal gradients in this zone vary from 1 to 2°C/100 m. The oil window occurs at depth of 2,750–5,200 m (Khan and Raza 1986) and this is reflected in the occurrence of oil at depths greater than 2,750 m.

Central Indus Basin: This basin is comprised of duplex structures characterised by large anticlines and domes in the passive roof sequence of the Sulaiman fold belt, followed eastward by gently dipping strata of the Punjab monocline which has few tectonic folds and faults (Chapter 4). The basin contains a sedimentary sequence ranging from Precambrian to Recent. It is essentially a natural gas-bearing zone and contains nine gas fields, including one giant field (Sui) with 8.6 TCF recoverable reserves and one large field (Pirkoh) with 2.6 TCF recoverable reserves. The main producing strata range in age from Cretaceous to Eocene (Fig. 10.8, Table 10.4). This basin is characterised by wide variations in geothermal gradients. Low (1.2°C/100 m) geothermal gradients occur in the eastern part (Fig. 10.9). It has been observed that there is a zone of very low geothermal gradient around the Sargodha–Shahpur buried ridge, which may be due to the high thermal conductivity of the shield (Khan and Raza 1986). However, in the central part of this basin, in the Sulaiman depression, there is a zone with the high geothermal gradient of 4.1°C/100 m. The main gasfields of Mari, Khandkot, Sui, Uch, Loti, Zin, Pirkoh and Jandran are concentrated in this region and it is likely that the heat-flow from this 'hotspot' has contributed to the development of these fields. Khan and Raza (1986) are of the view that in this region the 'oil window' may be below the gas-producing horizon with the possibility of oil occurrences in the Cretaceous sediments below the gas horizons.

Table 10.4. Potential source, reservoir and cap rocks in the Indus Basin.Producing reservoir rocks marked by asterisks. (Data from Raza et al. 1990, Kadri 1995).

Age	Source Rocks			Reservoir Rocks			Cap Rocks		
	Upper Indus	Middle Indus	Lr. Indus & Offshore	Upper Indus	Middle Indus	Lr. Indus & Offshore	Upper Indus	Middle Indus	Lr. Indus & Offshore
Pleistocene									
Pliocene									
Miocene			Gaj (shale) (offshore only)	Murree (sandstone)		Gaj (shale) (offshore only)	Murree (clays)		Gaj (shale) (offshore only)
Oligocene			Nari (shale) (offshore only)			Nari (sst/lst) (offshore only)			Nari (shale) (offshore only)
Eocene	Jatta Gypsum (shale) Nammal (shale)	Kirthar (limestone) Laki/Ghazij (lst / sh)	Kirthar (sh & lst) Laki/Ghazij (lst / sh)	Chorgali* Sakesar* Shekhan (limestone)	Kirthar * (limestone) Laki/Ghazij (limestone) Sui Main (lst)*	Kirthar (limestone) Laki/Ghazij (limestone)	Kohat Kuldana Nammal (shales)	Kirthar Laki Ghazij (shales)	Kirthar Laki Ghazij (shales)
Paleocene	Patala (shale) Lockhart (limestone)	Dunghan/ Ranikot (shale)	Lakhra (shale) Bara (shale)	Patala (limestone) Lockhart* (limestone)	Dunghan Ranikot* (sandstone)	Lakhra* (limestone) Bara (sst)	Patala Hangu (shales)	Dunghan Ranikot (shales)	Lakhra Bara Khadro (shales)
Cretaceous	Chichali (shale)	Moghal Kot (lst/marl) Chichali/ Sembar (shale)	Moghal Kot (Limestone) Goru (shale) Sembar(shale)	Lumshiwal (sandstone)	Pab (sst)* Chichali Goru (sst)* Sembar (sst/lst)	Pab (sst)* Goru (sst)* Sembar (sst)	Datta (shale)	Moghal kot Chichali Sembar (shales)	Moghal Kot Goru (shales)
Jurassic	Datta (shale)			Samana Suk* (limestone) Datta* (Shale)					
Triassic									
Permian				Nilawahan-Zaluch Gr.*(sst/lst) Tobra * (conglomerate)			Dandot (shale)		
Cambrian	Salt Range Formation	Salt Range Formation		Khewra * (sandstone)	Khewra (sandstone)	Khewra (sandstone)	Kussak (shale)	Kussak (shale)	



WELLS SHOWN IN THE ABOVE MAP

1. *Adhi-3*
2. *Adhi-5*
3. *Adhi-6*
4. *Balkassar-1*
5. *Dhulian-2*
6. *Dhulian-3*
7. *Dhulian-42*
8. *Dhulian-43*
9. *Kallar Kahar-1*
10. *Karsal-4*
11. *Karak-1*
12. *Kundian-1*
13. *Lilla-1*
14. *Mahesian-1*
15. *Meyal-4*
16. *Meyal-5*
17. *Meyal-6*
18. *Mianwala-1*
19. *Qazian-1*
20. *Tanwin-1*
21. *Toot-5*
22. *Toot-9*
23. *Warnali-1*
24. *Budhuana-1*
25. *Bannh-1*
26. *Dhodak-2*
27. *Domanda-1*
28. *Giandari-1*
29. *Jandran-1*
30. *Karampur-1*
31. *Kamiab-1*
32. *Kot-Rum-1*
33. *Mari-2*
34. *Mari-3*
35. *Sarai Sidhu-1*
36. *Sakhi Sarwar-1*
37. *Sui-20*
38. *Sui-22*
39. *Sui-23*
40. *Sui-25*
41. *Tadri Main-1*
42. *Tola-1*
43. *Badhra-1*
44. *Badin-1*
45. *Damiri-1*
46. *Dasori-1*
47. *Digh-1*
48. *Dabbo Creek-1*
49. *Indus Marine-A1*
50. *Indus Marine-B1*
51. *Indus Marine-C1*
52. *Jacobabad-2*
53. *Jhatpat*
54. *Khaskheli-1*
55. *Khairpur-2*
56. *Karachi-1*
57. *Karachi-2*
58. *Karachi South-A1*
59. *Korangi Creek-1*
60. *Lakhra-1*
61. *Mirpur Batoro-1*
62. *Nabisar-1*
63. *Phulji-2*
64. *Paitiani Creek-1*
65. *Patar-1*
66. *Sunbak-1*
67. *Sari Singh-1*
68. *Talhar-1*
69. *Tarai-1*
70. *Dhak-1*
71. *Dhak-2*
72. *Jal Pari-1A*
73. *Gar Koh-1*
74. *Kech Band-1*

Fig.10. 9. Geothermal gradients in selected oil and gas wells in the Indus Basin (from Khan and Raza 1986).

Southern Indus Basin: This basin is characterised by passive-roof duplex-type structure and a passive backthrust along the Kirthar fold belt, a passive roof thrust forming a frontal culmination wall along the margin of the fold belt, and the Kirthar depression, and out-of-syncline intra-molasse detachments in the Kirthar depression sequence (Chapter 4). The Kirthar and Karachi depressions contain several large anticlines and domes and some of these contain small gas fields (Mazarani, Sari, Hundi and Kothar). The eastern part of the basin comprising the Sindh monocline (Indus Platform) is largely comprised of faulted and tilted blocks of Mesozoic rocks which form structural traps and contain small oil and gas fields (Fig. 10.7). The fault blocks are unconformably overlain by Deccan Trap Basalts and Tertiary sedimentary rocks (Chapter 4). The northern margin of the Southern Indus Basin comprises the Sukkur Rift zone which bears large anticlinal structures and contains the Khandkot and Mari gas fields. The latter is a giant field with 6.3 TCF of reserves.

The main reservoir rocks in the Sindh monocline are the Cretaceous Lower Goru sandstone. In the Karachi depression production is from Paleocene Ranikot limestone and sandstone, in Kirthar depression and Sukkur Rift zone it is from Eocene Sui Main/Habib Rahi Limestone. The potential source, reservoir and cap rocks in this basin are listed in Table 10.4.

The Southern Indus Basin is also characterised by high geothermal gradients which range from 2 to over 4°C/100 m. The highest gradient has been recorded from the Damiri I well (Fig. 10.9).

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