CHAPTER 1

Basins in their plate tectonic environment

Assumptions, hasty, crude and vain,
Full oft to use will Science deign;
The corks the novice plies today
The swimmer soon shall cast away

(A.H. Clough, Poem (1840))

SUMMARY

Sedimentary basins are regions of prolonged subsidence of the Earth's surface. The driving mechanisms of subsidence are principally related to processes within the relatively rigid, cooled thermal boundary layer of the Earth known as the lithosphere. The lithosphere is composed of a number of plates which are in motion with respect to each other. Sedimentary basins therefore exist in a background environment of plate motion.

The Earth's interior is composed of a number of compositional and rheological zones. The main compositional zones are between crust, mantle and core, the crust containing relatively low density rocks overlain by a sedimentary cover. The mechanical and rheological divisions do not necessarily match the compositional zones. A fundamental rheological boundary is between the lithosphere and the underlying asthenosphere. The lithosphere is sufficiently rigid to comprise a number of relatively coherent plates. Its base is marked by a characteristic isotherm (*c.* 1330 °C) and is commonly termed the *thermal lithosphere*, or alternatively *mechanical lithosphere.* The upper portion of the thermal lithosphere is able to store elastic stresses over long time scales and is referred to as the *elastic lithosphere.* The continental lithosphere has a strength profile with depth which suggests that a weak, ductile zone exists in the lower crust, separating a brittle upper crust and upper mantle, giving a jam-sandwich type structure. The oceanic lithosphere, however, lacks this low strength layer, its strength increasing with depth to the brittle–ductile transition in the upper mantle.

The relative motion of plates produces deformation, vulcanicity, and seismicity concentrated along their boundaries, which are classified as *divergent boundaries,* such as the mid-ocean ridge spreading centres of the ocean basins, *convergent boundaries* associated with large amounts of shortening, such as continental collision zones, and *conservative boundaries* characterized by strike–slip deformation. Although the theory of plate tectonics has the premise that deformation is concentrated along plate boundaries, the continental lithosphere deforms far from plate boundaries, and appears to behave at geological time scales more like viscous sheets than as rigid stress guides.

Sedimentary basins have been classified principally in terms of the type of lithospheric substratum (i.e., continental, oceanic, transitional), their position with respect to the plate boundary (intracratonic, plate margin), and type of plate motion nearest to the basin (divergent, convergent, transform). The formative mechanisms of sedimentary basins fall into a small number of categories, although all mechanisms may operate during the evolution of a basin:

- *Isostatic consequences of changes in crustal/lithospheric thickness,* such as caused mechanically by lithospheric stretching, or purely thermally, as in the cooling and subsidence of the oceanic lithosphere as it moves away from oceanic spreading centres;
- *loading (and unloading)* of the lithosphere causes a deflection or flexural deformation and therefore subsidence (and uplift), as in foreland basins;
- viscous flow of the mantle causes nonpermanent subsidence/uplift known as *dynamic topography.*

From the point of view of lithospheric processes there are two major groups of basins: (i) Basins due to lithospheric stretching, belonging to the rift-drift suite, and (ii) basins formed primarily by flexure of continental and oceanic lithosphere.

Inspection of any map showing hydrocarbon occurrences (e.g., St John et al. 1984) reveals their clustered pattern. In general, provinces of hydrocarbon occurrence correspond to the locations of sedimentary accumulations greater than about 1 km thick. These accumulations include sedimentary basins in the strict sense, implying zones of pronounced subsidence (Bally and Snelson 1980) but also carbonate bank build-ups on elevated oceanic crust, cratonic arches, and so on which become fossilized in the geological record. However, sedimentary basins located at sites of prolonged and substantial subsidence are of overriding importance.

Historically, basin studies have developed from a number of distinct viewpoints such as that of stratigraphic sequences and their relation to sea-level fluctuations (Sloss 1950, 1963), the geosyncline (Kay 1947, 1951; Aubouin 1965), and, more recently, the concept of plate tectonics (Dickinson 1974; Ingersoll 1988). The location of sedimentary basins and their driving mechanisms are intimately associated with the motion of discrete, relatively rigid slabs, which together represent the cooled thermal boundary layer of the Earth, and with the convective flow of the underlying mantle. The outer shell of the Earth comprises a relatively small number of these thin, relatively rigid plates, and they are in a state of motion with respect to each other. Such motions set up plate boundary forces that may be transferred considerable distances into the interior of the plates, so that sedimentary basins exist in a background environment of stress set up by plate motion. The lithospheric plates are the surface manifestation of a slow thermal convection in the mantle, and are subject to differential thermal stresses along their bases. The mantle and lithosphere therefore do not operate as independent systems. We see spectacular evidence for the interaction of mantle processes and the lithosphere in the volcanic and topographic expression above plumes that have risen from the core–mantle boundary. We also discern, though less spectacularly, the effects on mantle flow of the subduction of cold slabs of oceanic crust at ocean–continent boundaries.

Some basic ideas on plate tectonics and Earth structure are introduced in this chapter in so far as they help to explain the location and evolution of sedimentary basins. More exhaustive summaries can be found in Wyllie (1971), Cox (1973), Le Pichon et al. (1973), Smith (1976), Bird (1980), Cox and Hart (1986), and Kearey and Vine (1996). Ingersoll and Busby (1995) is a collection of individually written chapters of relevance to basin analysis and plate tectonics.

The Earth's interior is composed of a number of essentially concentric zones that are defined on the basis of either compositional changes, or mechanical/rheological changes.

1.1 COMPOSITIONAL ZONATION OF THE EARTH

There are three main compositional units; the crust, mantle, and core (Fig. 1.1).

1.1.1 Oceanic crust

The *crust* is an outer shell of relatively low density rocks. The oceanic crust is thin, ranging from approximately 4 to 20 km in thickness, 10 km being "normal," and with an average density of about 2900 kg m^{-3}. It comprises a number of layers that reflect its mode of creation: an upper veneer (layer 1) of unconsolidated or poorly consolidated sediments, generally up to 0.5 km thick; an intermediate layer 2 of basaltic composition, consisting of pillow lavas and associated products of submarine eruptions; and a layer 3 of gabbros and peridotites that may form the parent rocks which upon differentiation give rise to the basalts of layer 2. The oceanic crust has been thought to be distinctly layered in terms of velocity of seismic waves, but more recent views are that it possesses a more gradual and continuous increase in velocity with depth.

The lifetime of oceanic crust is short, despite the fact that it occupies about 60% of the surface of the Earth (*c.* 3.2×10^9 km^2). This is because as the oceanic crust cools during aging it becomes gravitationally unstable with respect to the mantle; as a result it is consumed. This explains why the oldest oceanic crust in today's oceans is as young as Jurassic in age (*c.* 150 Ma). Compared to the continents, the oceanic crust therefore has a very short lifespan.

1.1.2 Continental crust

The continental crust is thicker, ranging from 30 to 70 km, but with an "average" thickness of perhaps 35 km. It was originally thought to be divided into two layers, each with a distinct composition and density: (i) An upper layer with physical properties similar to those of granites, granodiorites, or diorites overlain by a thin veneer of sedimentary rocks. This so-called "*granitic layer*" has a thickness of between 20 and 25 km and a density of 2500–

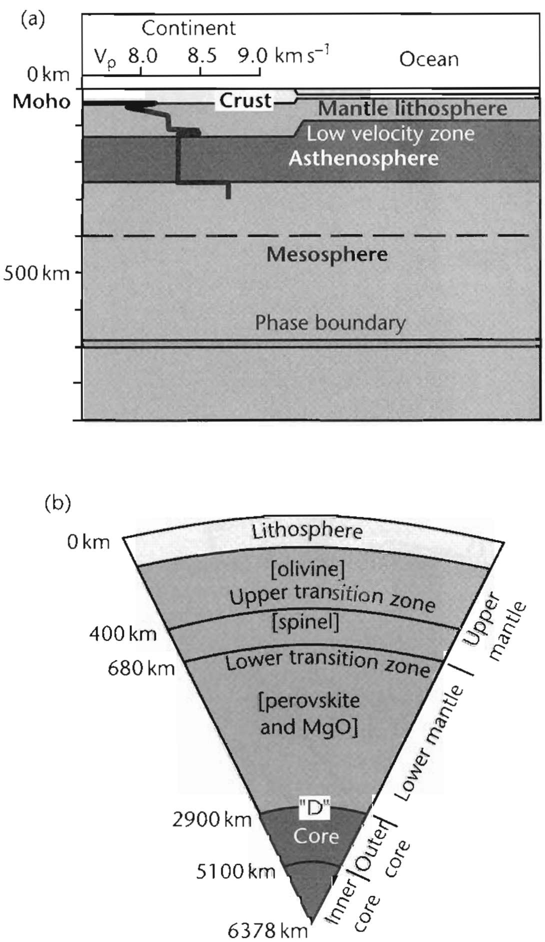

Fig. 1.1 The main compositional (a) and rheological (b) boundaries of the Earth. The most important compositional boundary is between crust, mantle, and core. There are strong compositional variations within the continental crust and compositional variations caused by phase changes in the mantle. V_p is velocity of **P** wave. The main rheological boundary is between the lithosphere and the asthenosphere. **P** wave velocities increase markedly beneath the Moho, but decrease in a low velocity zone representing the weak asthenosphere. The lithosphere is rigid enough to act as a coherent plate. **P** wave velocity in (a) from western Europe after Hirn (1976).

$2700\,kg\,m^{-3}$. The term "granitic" is, however, misleading, since average densities are greater than that of granite; (ii) a lower layer of primarily basaltic composition, but the pressure and temperature at depths in excess of 25 km imply that the rocks are granulites, or their high pressure, high temperature equivalents, eclogites or amphibolites. The density of this lower layer is $2800–3100\,kg\,m^{-3}$. These layers may not in reality be well defined, and instead a more continuous variation of composition with depth may exist.

Information on the density of crustal rocks has been obtained largely by observations on seismograms of the speed of seismic waves passing through the various layers, coupled with laboratory experiments on rock materials. The existence of a low velocity crust was discovered by the geophysicist Mohorovicic shortly after the turn of the century. At the crust–mantle boundary, seismic **P** (longitudinal) wave velocities increase markedly; this abrupt increase in velocity may reflect a corresponding increase in rock density (Fig. 1.1). This horizon is known as the Mohorovicic discontinuity or Moho. The Moho varies in depth considerably. The continental crust thickens under orogenic belts, thins under zones of rifting, and attenuates completely at continental margins (Fig. 1.2).

In some regions, particularly the attenuated margins of continents, the crust is intermediate in character and thickness between typical oceanic and continental varieties. This may be due to the injection of dense intrusions, to metamorphism, or to other processes accompanying stretching. In particular, the depth to the Moho may be abnormally great due to the igneous underplating of the crust during plume activity. Several kilometers of underplate emplaced in the Early Tertiary over the head of the Iceland plume have been interpreted from seismic experiments carried out on the northwest European continental margin.

1.1.3 Mantle

The *mantle* is divided into two layers: the upper and lower mantle. The upper mantle extends to about 680 km ± 20 km and is punctuated by phase transitions. The inner mantle extends to the outer limit of the core at 2900 km, with an increasing density with depth.

Although there are of course no *in situ* measurements of the composition of the mantle, it can be estimated from the chemistry of volcanic and intrusive rocks derived by melting of the mantle, from tectonically emplaced slivers of mantle rock preserved in orogenic belts known as ophiolites, from nodules preserved in volcanic rocks, from minerals brought to the surface explosively in kimberlites, and importantly, from the remote

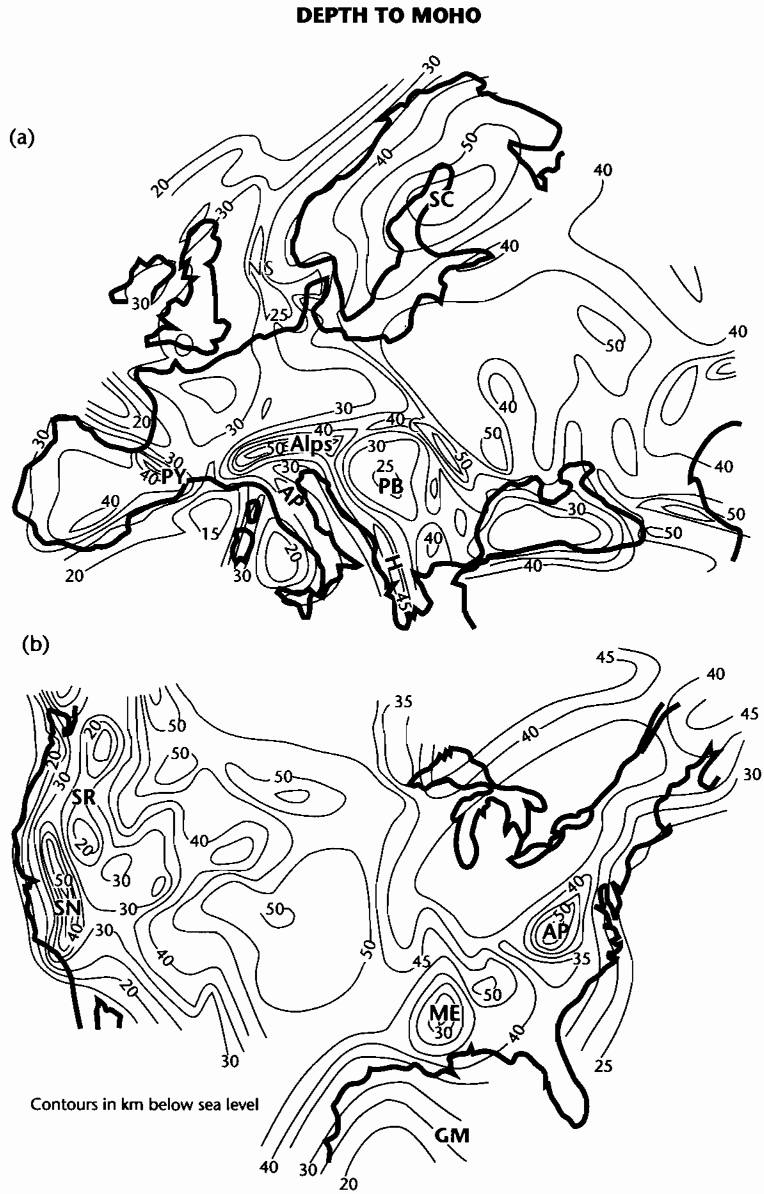

Fig. 1.2 Depth to the Moho below sea-level from (a) Europe and (b) North America, after Allenby and Schnetzler (1983) and Meissner (1986). In Europe, thick continental crust occurs in the Pyrenean–Alpine–Carpathian orogenic belt and in the Scandinavian Shield. Thin continental crust occurs along the Atlantic margin and in regions of continental stretching, such as the North Sea, western Mediterranean, Pannonian Basin, and Black Sea. In North America, thick continental crust is associated with the batholiths of the Sierra Nevada region, the Appalachian orogenic belt, the American midcontinent and the Canadian–NE USA shield. Thin crust is associated with the plateau basalts of the Snake River area, sites of rifting such as the Mississippi Embayment, and along the western active continental margin, eastern passive continental margin and the Gulf of Mexico. NS, North Sea; SC, Scandinavian Shield; PY, Pyrenees; AP, Apennines; PB, Pannonian Basin; H, Hellenides; SR, Snake River; SN, Sierra Nevada; ME, Mississippi Embayment; AP, Appalachians; GM, Gulf of Mexico. Reproduced courtesy of Elsevier.

but sophisticated measurement of the mantle using seismic waves. The main constituent of the mantle is thought to be olivine, mostly the Mg-rich variety forsterite.

Olivine is known to undergo phase changes to denser structures at pressures equivalent to depths of 390–450 km and *c.* 700 km in the Earth. At 390–450 km olivine is thought to change to spinel *via* an exothermic reaction involving a 10% increase in density. At *c.* 700 km spinel changes to perovskite and magnesium oxide in an endothermic reaction. These phase changes can be recognized by changes in the velocity of **S** waves (McKenzie 1983) and may determine the scale of convection in the mantle (Silver et al. 1988).

1.2 RHEOLOGICAL ZONATION OF THE EARTH

The mechanical or rheological divisions of the interior of the Earth do not necessarily match the compositional zones. One of the rheological zonations of primary interest to students of basin analysis is the differentiation between the lithosphere and the asthenosphere. This is because the vertical motions (subsidence, uplift) in sedimentary basins are principally a response to the deformation of this uppermost rheological zone of the Earth.

1.2.1 Lithosphere

The *lithosphere* is the rigid outer shell of the Earth, comprising the crust and the upper part of the mantle. It is of particular importance to note the difference between the *thermal* and *elastic* thicknesses of the lithosphere. It is generally believed (e.g., Parsons and Sclater 1977; Pollack and Chapman 1977) that the base of the lithosphere is represented by a characteristic isotherm (1100–1330°C) at which mantle rocks approach their solidus temperature. This defines the *thermal lithosphere.* Typical thicknesses of lithosphere under the oceans varies from *c.* 5 km at mid-ocean ridges to *c.* 100 km in the coolest parts of the oceans. The lower boundary of the lithosphere is poorly defined under continents, depths of 100 to 250 km being typical. The stepwise increases in velocities of **S** and **P** waves with depth through the lithosphere suggest that it contains compositional boundaries within it.

The rigidity of the lithosphere allows it to behave as a coherent plate but only the upper part of the lithosphere is sufficiently rigid to retain elastic stresses over geological time scales (say 10^7 years). Below this upper *elastic lithosphere* creep processes efficiently relax elastic stresses, so that there is a physical and conceptual difference between the elastic lithosphere and the thermal lithosphere. The lithosphere below the upper elastic portion must therefore be sufficiently soft to relax elastic stresses but sufficiently rigid to remain a coherent part of the surface plate.

The oceanic and continental lithosphere differ in their strength (Fig. 1.3). The strongest part of the oceanic lithosphere occurs in the mantle between 20 and 60 km depth, below which it becomes increasingly ductile. The continental lithosphere, however, appears to be markedly zoned rheologically. In particular, the upper seismically active brittle zone overlies a generally aseismic zone that may deform by ductile processes. This mid-lower crustal ductile zone has been invoked as a level of detachment of major upper crustal faults (e.g., Kusznir and Park 1987) (§3.5.2 and §3.5.3). There is a second, deeper strong layer in the mantle part of the continental lithosphere where earthquakes occasionally occur (Chen and Molnar 1983).

There are also heterogeneities in the mantle part of the lithosphere, although they are small compared with the crust. Seismological studies of western Europe (Hirn 1976) suggest a highly stratified lithosphere beneath the Moho (Fig. 1.1). In particular, a "channel" of reduced **P** wave velocities has been interpreted between 10 and 20 km below the Moho. This 10 km thick layer cannot be explained in terms of partial melting since the solidus temperature is far in excess of the actual temperature – the hydration (serpentinization) of peridotites has been postulated as a possible mechanism. Whatever the cause, this upper low velocity channel may serve as a zone of decoupling of the upper lithosphere from the lower portion of the lithosphere when acted upon by tangential tectonic forces. There are few examples, however, where a process of decoupling can be unambiguously demonstrated at these levels.

1.2.2 Sublithospheric mantle

The underlying region, the *asthenosphere* is weaker than the lithosphere and is able to undergo deformation relatively easily by flow. The upper part of the asthenosphere is known as the low velocity zone where **P** and **S** wave

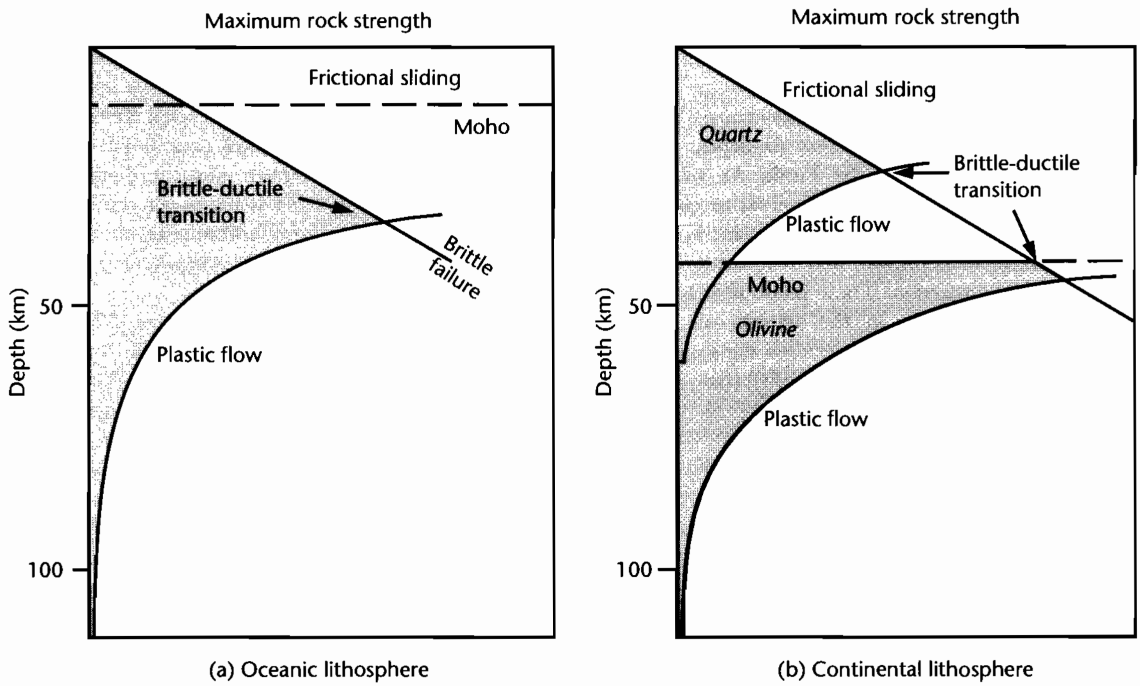

Fig. 1.3 Strength profiles for the oceanic (a) and continental (b) lithosphere, based on Molnar (1988) and Sammonds (1999). The yield strength of the continental and oceanic lithosphere is plotted as a function of depth. The olivine rheology of the oceanic lithosphere provides a strong elastic core extending to depths of over 50 km. The quartz or quartz-felspar rheology of the continental lithosphere causes a weak, ductile layer at equivalent depths. A second brittle–ductile transition occurs in the mantle lithosphere because of the compositional change to an olivine rheology. The elastic lithosphere is the upper portion that is able to store elastic stresses over long time periods. The base of the thermal lithosphere is a mechanical boundary separating the relatively strong outer shell of the lithosphere from the very weak asthenosphere.

transmission speeds drop markedly, presumably due to partial melting (Fig. 1.4).

Studies of minute variations in the transmission speeds of seismic waves has allowed the structure of the deep mantle to be visualized and mapped (Dziewonski and Woodhouse 1987), a topic known as *seismic tomography*. Zones of faster than average seismic velocity are attributed to propagation through denser rock which in turn is most likely due to a cooler temperature. Zones with slower than average seismic velocity are likewise thought to be due to warmer temperatures. Variations in temperature are probably caused by large-scale convection. Far from being inert, the mantle represents a vast volume of rock which dynamically interacts with the lithosphere. Instabilities rise from the core–mantle boundary as *plumes* of hot material which impinge on the base of the overlying lithosphere and may have a major role in continental break-up (Burke and Dewey 1973; White and McKenzie 1988). The thermal effects of the subduction of cold oceanic lithosphere and the insulating effects of supercontinental assemblies are also thought to be recognizable in the thermal structure of the mantle (Gurnis et al. 1996).

1.3 PLATE MOTION

Plate tectonics can operate because the lithosphere is composed of a number of coherent rheological "plates" (Fig. 1.5). The underlying concepts of relative plate motion come from studies of focal mechanism solutions of large earthquakes and observations of the distribution of earthquake epicenters, and from studies of magnetic lineations in the ocean basins. The nature and rates of

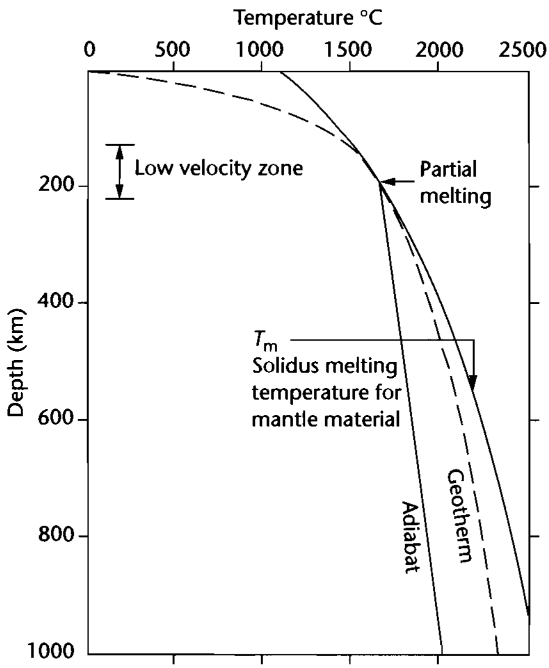

Fig. 1.4 Variation of temperature with depth, or geotherm, and the solidus temperature for mantle material (peridotite). Where the solidus curve (T_m) and the geotherm become tangential, partial melting in the mantle is likely to take place, resulting in a zone of low seismic wave velocities (low velocity zone).

relative plate motion (Minster and Jordan 1978) govern many aspects of the geodynamic environment of basins. Increasingly, reliable information on motion within plates is provided by measurement of *geodetic* strains involving the long-term occupation of triangulation networks and the use of ground position satellite (GPS) technology (Clarke et al. 1998).

The global pattern of seismic activity is of continuous and narrow belts of high frequency of earthquakes, bounding extensive regions of relative stability (Barazangi and Dorman 1969). The narrow zones of earthquake activity define plate margins. Oceanic plate boundaries are very sharply defined whilst continental boundaries are rather more diffuse.

The fact that earthquake epicenters occur at depths as great at 650 to 700 km along some plate boundaries suggests that a process exists that is capable of transferring brittle material to depths normally associated with deformation by flow. This process of plate subduction is responsible for both the relative youth of the oceanic crust and the distribution of earthquake epicenters. The fact that the interiors of plates experience only infrequent earthquake activity reflects the concentration of large relative motions of plates along their boundaries. However, this does not mean that the interiors of plates do not experience significant deformation. Continental plates clearly undergo extensive deformation a long way from plate boundaries (England 1987). In such cases, the driving force is believed to be the excess potential energy of elevated continental crust.

The lithospheric plates can be easily deformed by bending about a horizontal axis, but are highly resistant to torsion about steeply inclined axes. This latter property of strength allows the motion of plates over the Earth's surface to be modeled assuming no internal deformation, except at plate boundaries. But how do the oceanic and continental lithosphere compare in terms of flexural strength? Different views exist on this problem. On the one hand, oceanic plates are stronger because they consist of more mafic mineral assemblages, whereas the continents contain quartz which shows ductile flow at lower temperatures than olivine (see Fig. 1.3). They also contain fewer intrinsic weaknesses such as old fundamental fault systems. On the other hand, the oceanic plates are thinner and hotter, and therefore bend more easily under an applied force system. Whether continental or oceanic lithosphere is stronger in terms of their resistance to bending, or flexure, is therefore controversial and the strength must at least in part depend on parameters such as geothermal gradient and strain rate. Continental crust, however, appears to be weaker than oceanic crust when subjected to extensional stresses (Steckler and ten Brink 1986).

Three classes of plate boundary exist: divergent, convergent, and conservative (Fig. 1.6):

Divergent boundaries are typified by the mid-ocean ridge spreading centres of the ocean basins. Here, the recognition of magnetic bands correlated with a magnetic reversal chronology (Vine and Matthews 1963; Cox 1973) allows the rate of divergent plate motion to be estimated. Transform faults with strike–slip displacement offset the divergent boundaries, producing a highly segmented pattern.

Convergent boundaries are of two classes:

- Subduction boundaries where oceanic lithosphere constitutes the downgoing plate. Ocean–ocean

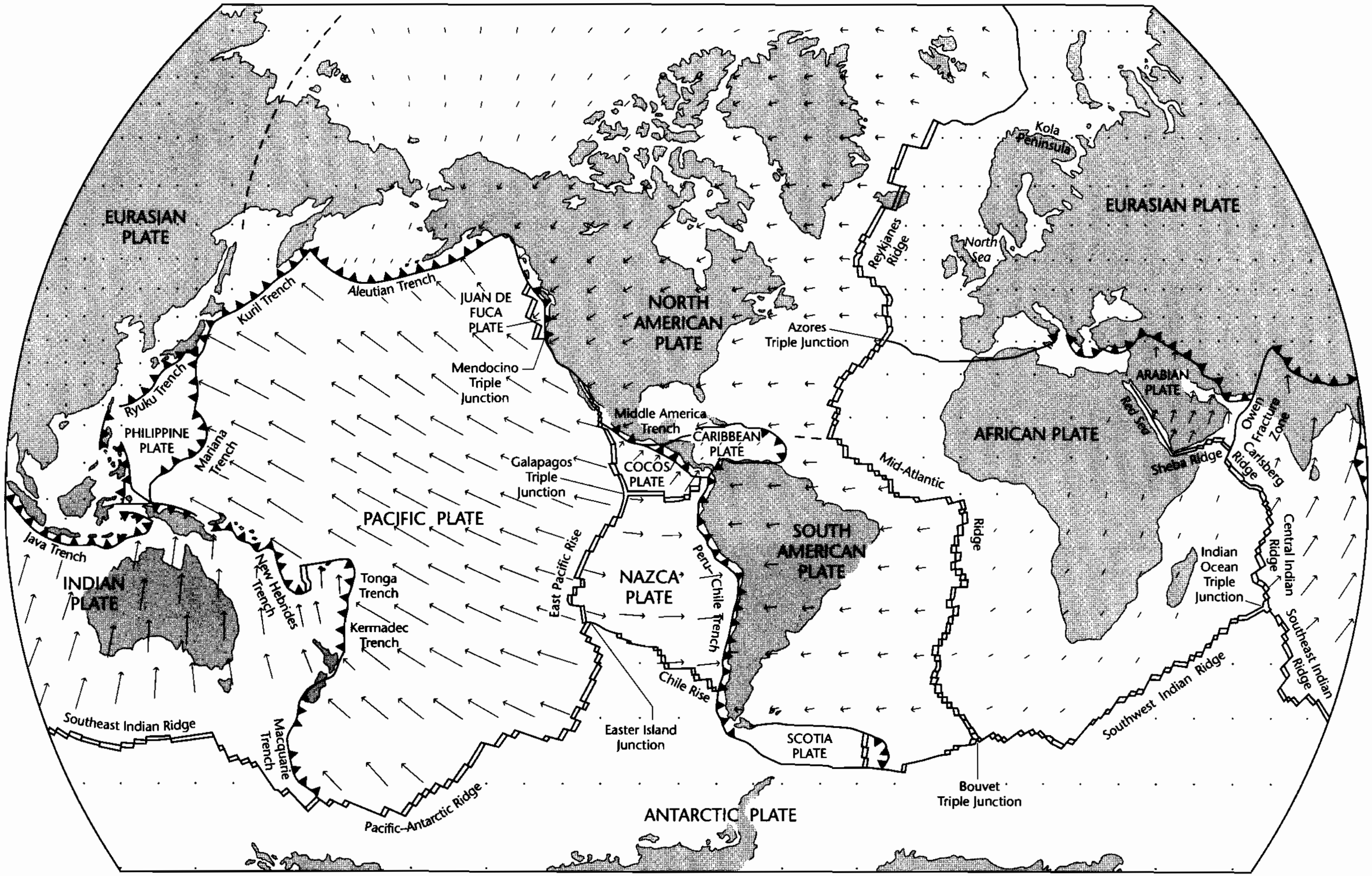

Fig. 1.5 The lithospheric plates, showing mid-ocean ridges, trenches, and transform boundaries (Le Pichon et al. 1973) and absolute motion vectors from Minster and Jordan (1978). Length of arrows is proportional to the plate speed. The fastest plate motion is in the western Pacific and Indian Oceans, whereas Africa, Antarctica, and Eurasia are almost stationary with respect to the mantle reference frame. Reproduced courtesy of American Geophysical Union.

boundaries, as for example, in the Mariana Islands, are characterized by a well-developed ocean trench and volcanic island arcs, whereas ocean–continent boundaries such as along the west of the Andes consist of an ocean trench with an associated continental magmatic arc with intense plutonic activity.

- Collisional boundaries where continental lithosphere constitutes the downgoing plate. Where both plates are continental, as in the Alps or Himalayan zones, the buoyancy of the downgoing plate resists subduction, leading to intense and widespread deformation. Less commonly, oceanic lithosphere may override continental lithosphere attached to subducting oceanic lithosphere, as in Taiwan.

Conservative boundaries occur where the adjoining plates are moving parallel to each other and are therefore dominated by strike–slip or transform faults.

The relative movement between plates causes earthquakes, a fact demonstrated by the concentration of seismic activity along plate boundaries. Earthquakes occur along trenches, ridges, and transforms, but they are distinctly different along the three types of boundary:

- Ridges are characterized by small to moderate earthquakes generated at shallow depths of <10 km;
- transforms experience larger earthquakes originating from depths of <20 km;
- subduction zones are sites of very large and deep earthquakes, with foci occurring as deep as 700 km.

The disappearance of earthquakes at relatively shallow levels along transforms and ridges is thought to be due to the change in rheology from brittle (capable of storing elastic stresses before rupture) to ductile (flowing by creep). This transition takes place in the range 600–900 °C, which corresponds to a depth of 20–30 km at transforms, but at shallower levels of about 10 km at ridges where temperatures are elevated. In contrast, at subduction zones, if the plate is descending quickly it remains cool relative to its surroundings and is capable of brittle deformation to large depths. Hence, earthquake foci along subduction zones (Fig. 1.7) may be very deep. The rate of slab subduction may affect the depth of earthquakes, however. If the slab is only slowly subducted, it may heat up sufficiently to prevent the occurrence of earthquakes at great depths.

It is possible to study the type of motion that occurred during a particular earthquake and to work out the stresses released at the time of the earthquake, and thereby the direction of plate motion that gave rise to the stresses. These methods are called *first motion* or *focal mechanism* studies, since it is the first motion of the ground surface, whether it is up/away from the source of the earthquake or down/towards the focus. Distinct radiation patterns result from first motion on strike–slip, normal, and reverse faults.

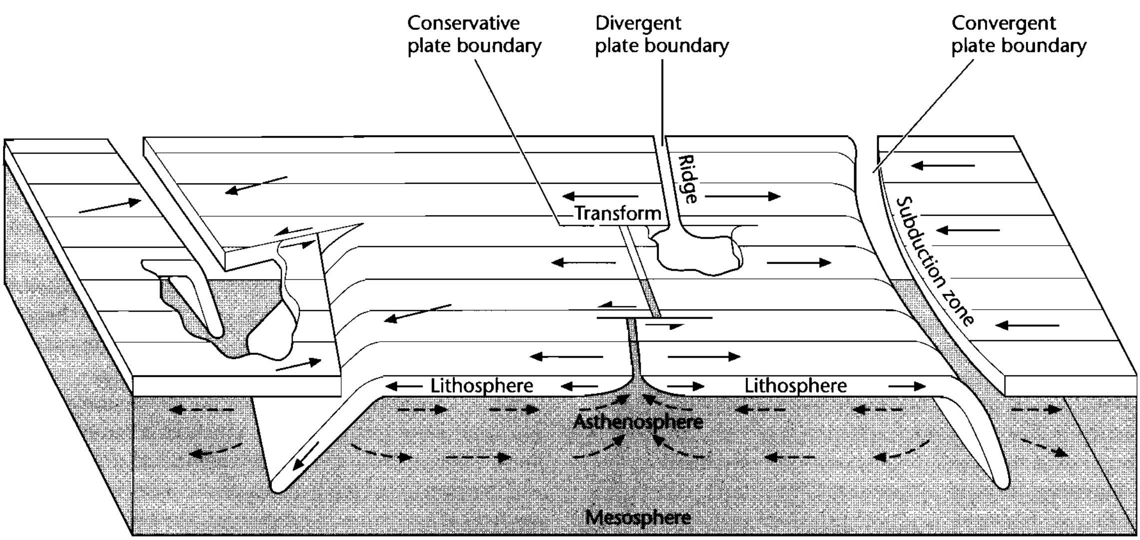

Fig. 1.6 The three types of plate boundary: convergent, divergent, and conservative (after Kearey and Vine 1996). Reproduced courtesy of Blackwell Publishing Ltd.

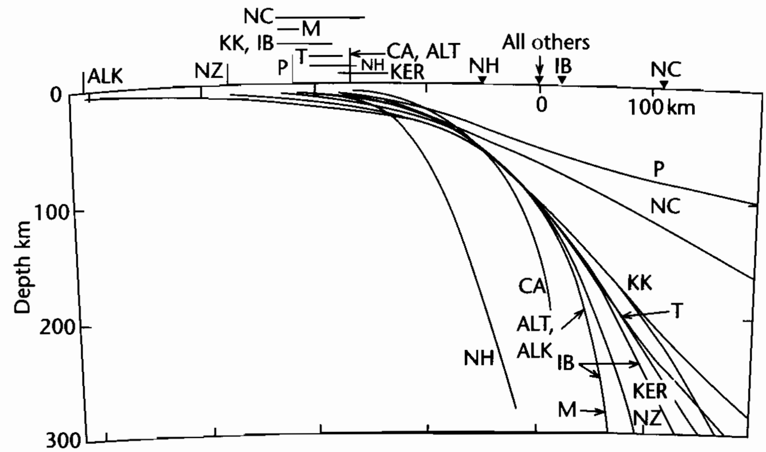

Fig. 1.7 Distribution of earthquake foci along Benioff zones, after Isacks and Barazangi (1977). NH, New Hebrides; CA, Central America; ALT, Aleutians; ALK, Alaska; M, Marianas; IB, Izu-Bonin; KER, Kermadec; NZ, New Zealand; T, Tonga; KK, Kuril-Kamchatka; NC, North Chile; P, Peru.

Sykes (1967) studied the first motion of earthquakes along mid-Atlantic ridge transforms and suggested that the motion was strike–slip, to the east on the northern blocks and to the west on the southern blocks. This is right lateral motion – opposite to that indicated by the offsets of the ridge – but the correct relative movement to support the plate tectonic interpretations of transforms as actively shearing only between the ridge segments and not beyond. The earthquakes originating from spreading centres are quite different from those being produced at transforms. First motion studies suggest that faults in the mid-ocean ridge are dip–slip and extensional. The situation at trenches is more complex. First motion studies of earthquakes along the *Benioff Zone* show that faulting takes place at roughly 45° to the inclined surface of the downgoing slab. At depths of greater than about 300 km, the focal mechanisms are compressional, but at shallower depths they are tensional. This pattern supports the view that in the lowermost part of the subducting slab, the plate experiences compression as it is forced into a zone of greater viscosity or strength at depth. The upper part, however, is in a state of tension because of the gravitational body forces on the cool plate "hanging" from its upper edge. This force constitutes slab-pull. Isacks et al. (1969) provide a detailed analysis of the use of seismological studies at sites of subduction.

The relative motion of plates with constructive, conservative, and destructive plate margins creates a continually changing picture of continental splitting, ocean basin creation, ocean closure, and continental collision. This cycle of plate motion involving the birth and closure of oceans is termed the Wilson cycle since it is based on early ideas of the opening and closing of the Atlantic ocean by John Tuzo Wilson (Wilson 1966) (Fig. 1.8). Many sedimentary basins can be fitted into a particular phase of the Wilson cycle.

1.4 CLASSIFICATION SCHEMES OF SEDIMENTARY BASINS

Ideally, classifications are theories about the basis of natural order rather than dull catalogues compiled only to avoid chaos (Gould 1989, p.98, quoted in Ingersoll and Busby 1995, p.2). In this sense, classification schemes for sedimentary basins should both reveal something of the underlying mechanisms for basin development and reflect the natural variability of the real world.

Recent classification schemes of sedimentary basins based on plate tectonics have much in common. Their lineage derives from Dickinson's influential work in 1974 which emphasized the position of the basin in relation to the type of lithospheric substratum, the proximity of the basin to a plate margin, and the type of plate boundary nearest to the basin (divergent, convergent, transform) (Fig. 1.9). The evolution of a basin could then be

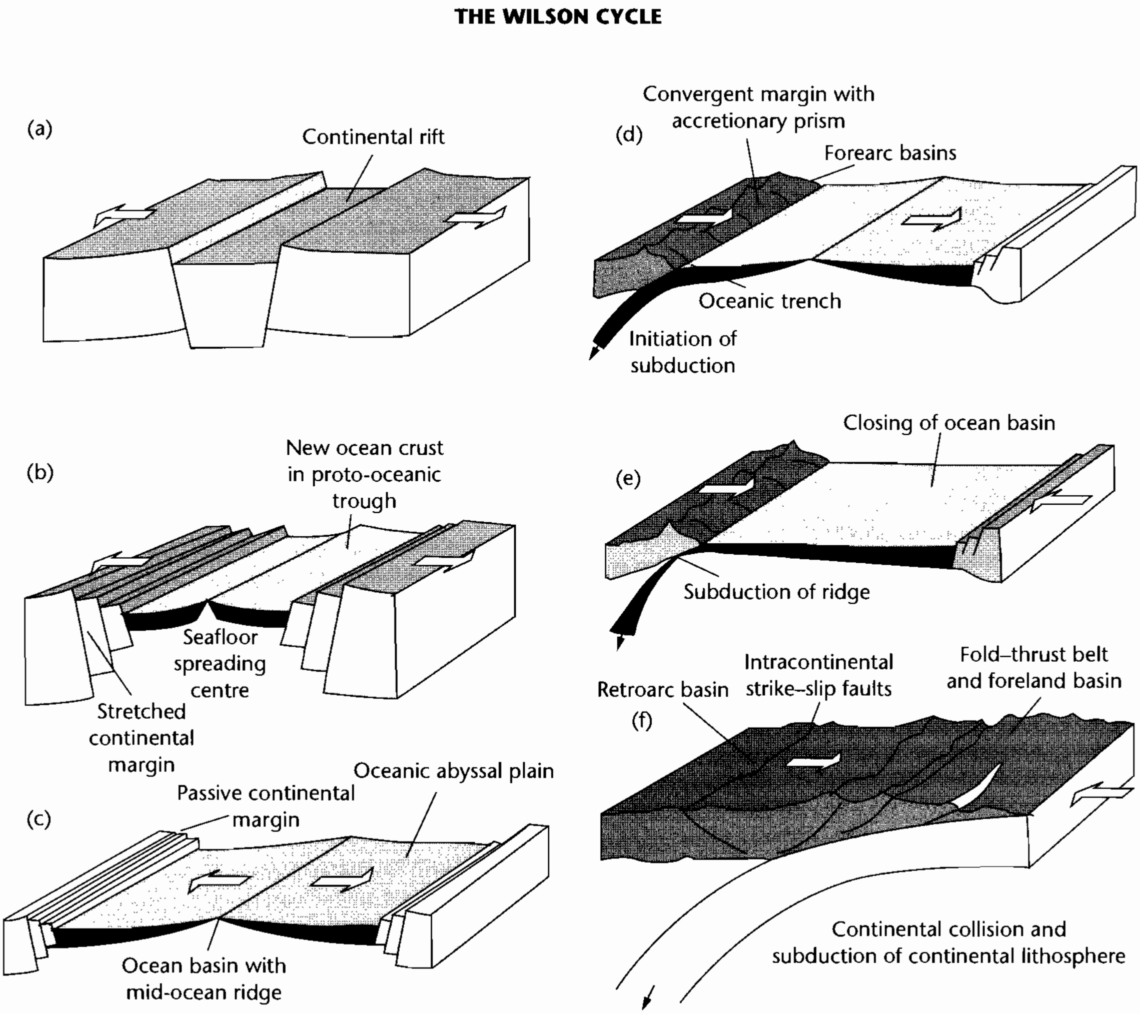

Fig. 1.8 The Wilson cycle of ocean formation and ocean closure. Continental extension (a) is followed by the creation of a new oceanic spreading centre (b) and ocean enlargement (c). Subduction of ocean floor (d) leads to closure of the ocean basin. Subduction of the oceanic ridge (e) takes place before continent–continent collision (f).

explained by changing plate settings and interactions. Dickinson (1974) recognized five major basin types on this basis: (i) Oceanic basins, (ii) rifted continental margins, (iii) arc–trench systems, (iv) suture belts, and (v) intracontinental basins.

Strike–slip or transform related basins were conspicuously missing as a distinct basin type in this classification, a deficiency corrected in Reading (1982). Bally (1975) and Bally and Snelson (1980) differentiated three different families of sedimentary basins based on their location in relation to megasutures, which in this context can be defined to include all the products of orogenic and igneous activity associated with predominantly compressional deformation. The boundaries of megasutures are often associated with subduction, whether it be of slabs of oceanic lithosphere (Benioff or B-type subduction) or of relatively buoyant continental lithosphere (Amferer or A-type subduction) and may also be the sites of impor-

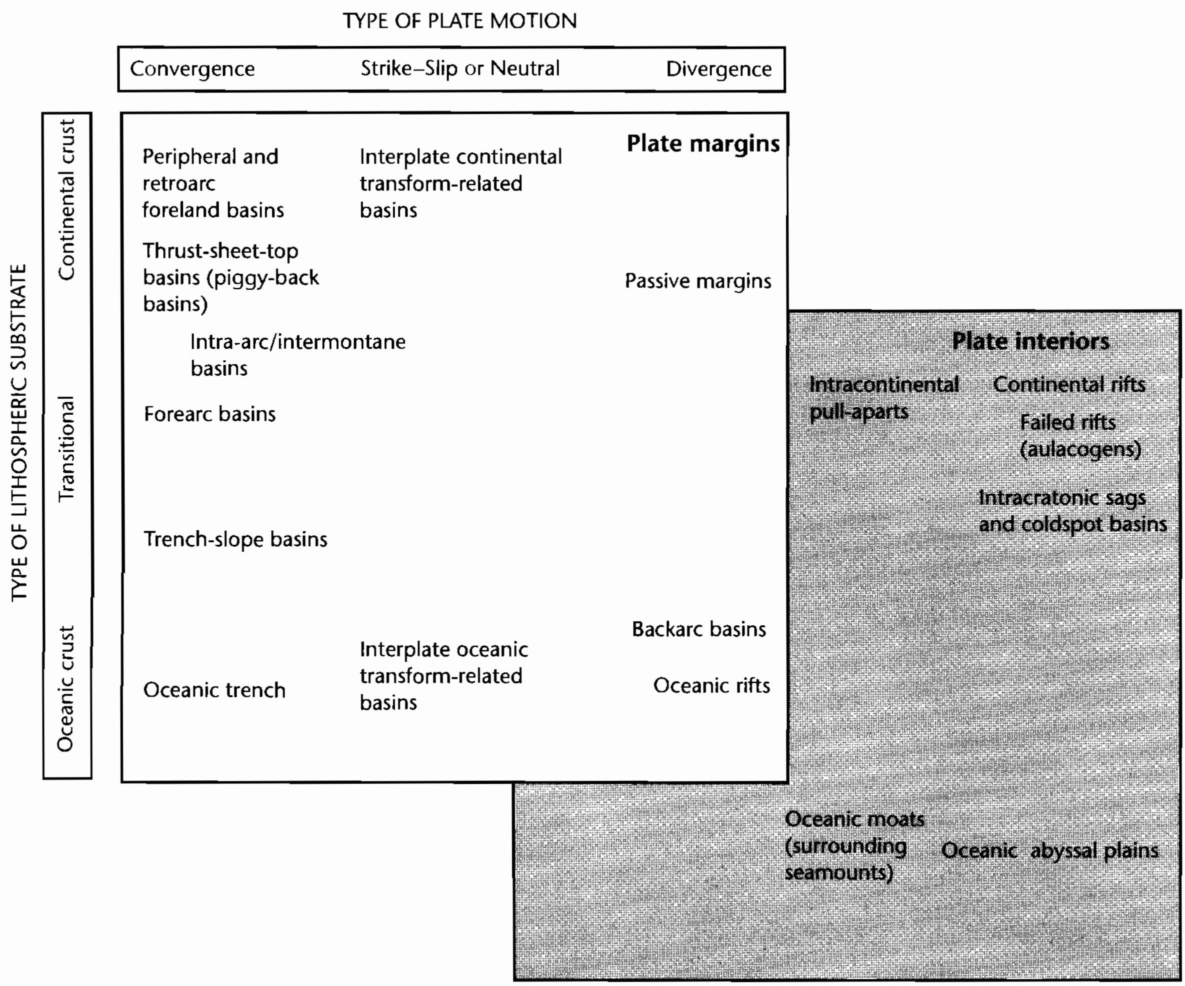

Fig. 1.9 Classification of basins using the type of lithospheric substrate, type of plate motion, and location with respect to the plate boundary.

tant wrench tectonism along transform faults. Ingersoll and Busby (1995) developed the classifications of Dickinson (1974) and Ingersoll (1988) to recognize 26 different types grouped into classes of divergent settings, intraplate settings, convergent settings, transform settings, and hybrid settings.

Industry-based classifications are typified by the scheme suggested by Halbouty et al. (1970) and later developed by Fischer (1975) and Klemme (1980). Klemme's scheme recognizes eight main types of basin based on their architectural characteristics such as linearity, asymmetry, cross-sectional geometry, which are themselves related to the tectonic setting and basin evolution. The goal of categorizing a sedimentary basin and thereby gaining some predictive insights into frontier basins is common to industry classifications such as those of Huff (1978) and Klemme (1980). It is pursued by an Exxon group (Kingston et al. 1983a, b) to the extent of devising a formula for each basin, thereby facilitating easy comparisons between basins and providing an "instant" idea of hydrocarbon potential. This classification system (Fig. 1.10) once again places basins primarily in their plate tectonic setting (lithospheric substrate, type of plate motion, and location on plate), reminiscent of Dickinson's analysis over a decade earlier, and categorizes a basin according to three critical factors:

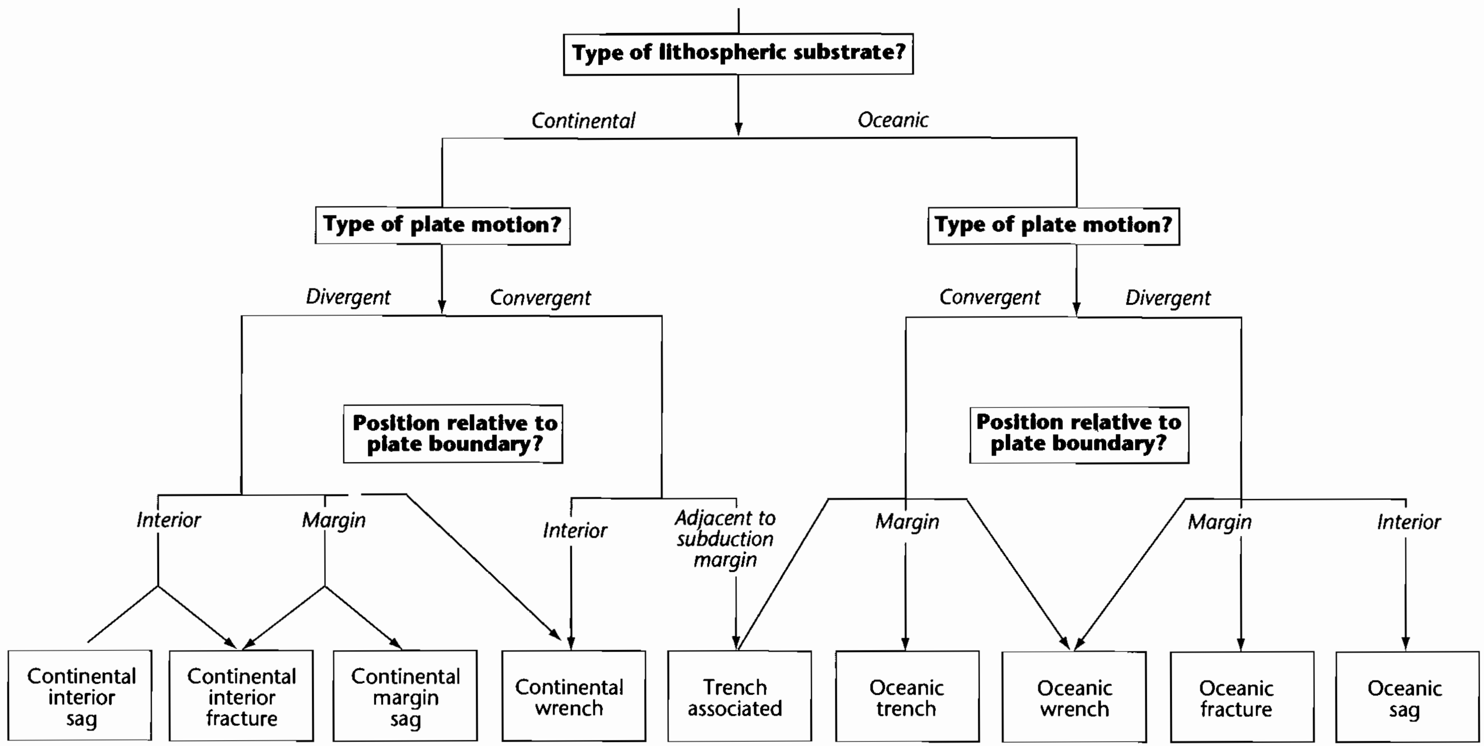

Fig. 1.10 Basin classification scheme based on Kingston et al. (1983a, b). Not all basin types appear in this scheme. Most notably, foreland basins are missing.

the basin-forming tectonics, the depositional sequences filling the basin, and the basin-modifying tectonics.

1.4.1 Basin-forming mechanisms

Although the classifications outlined above undoubtedly have their uses, particularly in predicting source presence, reservoir quality, availability of traps, etc., they have the effect of scrambling some of the essential differences and similarities between basins from the point of view of lithospheric mechanisms. Ingersoll and Busby (1995) recognized six subsidence mechanisms, operating to different degree in their 26 basin types (Table 1.1), which can be summarized as:

- Crustal thinning, such as caused primarily by stretching or surface erosion;
- lithospheric thickening, such as caused by cooling following stretching or accretion of melts derived from the asthenosphere;
- sedimentary and volcanic loading causing isostatic compensation;
- tectonic (supracrustal) loading causing isostatic compensation;
- subcrustal loading caused by subcrustal dense loads such as obducted mantle flakes or crustal densification due to phase changes;
- asthenospheric flow primarily due to the subduction of cold lithospheric slabs.

From the point of view of fundamental lithospheric processes, the major mechanisms for regional subsidence and uplift (*isostatic*, *flexural*, and *dynamic*) can be summarized as follows (Fig. 1.11):

Isostatic consequences of changes in crustal and lithospheric thickness; the thickness changes may be brought about purely thermally by *cooling* of lithosphere, for example where new oceanic lithosphere moves away from spreading centres. Thickness changes causing *thinning* may be caused by mechanical stretching, subaerial erosion or at depth by delamination or the removal (stoping) of a deep lithospheric root. Mechanical *thickening* of crust and lithosphere, as in zones of continental convergence, generally causes isostatic uplift. Thickening of the lithosphere by cooling, however, causes subsidence.

Loading and unloading at the surface and in the subsurface, including the far-field effects of in-plane stresses; *loading* of the lithosphere may take place on a small scale in the form of volcanoes or seamount chains, and on a large scale in the form of mountain belts, causing *flexure* and therefore subsidence. The sediment infilling a basin also acts as a sedimentary load, amplifying the primary driving mechanism.

Dynamic effects of asthenospheric flow, mantle convection, and plumes; subsidence or uplift are caused by the buoyancy effects of changes in temperature in the mantle. Since these temperature changes are transmitted by viscous flow, the surface elevation changes may be termed *dynamic*.

For a given basin type, some or all of the mechanisms given above may have a major or minor role. Consequently, rather than attempting an encyclopedic coverage of all basin types, we focus on the main lithospheric processes in Part 2 of this book. After an introductory chapter on the fundamentals of lithospheric mechanics (Chapter 2), basin forming processes are considered by investigating basins primarily caused by lithospheric stretching (Chapter 3) and basins primarily caused by flexure (Chapter 4). Chapter 5 discusses the role of mantle–lithosphere interactions in basin development. Basins related to strike–slip deformation are considered in Chapter 6.

Basins formed by stretching or thinning of the continental lithosphere fall within an evolutionary sequence (Kinsman 1975; Veevers 1981). The early stages of the sequence correspond to the development of intracratonic rifts often associated with crustal doming. Such rifts may evolve into oceanic spreading centres or may be aborted to form failed rifts or aulacogens. With seafloor creation and drifting of the continental edge away from the spreading centre, passive margin basins develop. The sequence has been termed the *rift-drift suite* of sedimentary basins. The mechanisms of interest within this evolutionary sequence are therefore primarily the thermal and mechanical behavior of the lithosphere under tension, and the thermal contraction of the lithosphere following stretching.

Basins formed by flexure fall into two groups. Flexure of oceanic lithosphere as it approaches subduction zones is responsible for the formation of deep oceanic trenches. It was the investigation of the deflection of the oceanic lithosphere at arc–trench boundaries that provided much of the framework for the general theory of lithospheric flexure. Flexure of the continental lithosphere in continental collision zones gives rise to foreland basins. The force system that causes flexure can be varied. In the case of ocean trenches, it is probably a combination of gravitational body forces on the downgoing oceanic slab and

Table 1.1 Basin classification adapted from Ingersoll and Busby (1995), modified from Dickinson (1974, 1976) and Ingersoll (1988), with modern and ancient examples.

Relative plate movement	Basin type	Basin description	Modern example	Ancient example
Divergent settings	Terrestrial rift valleys	Rifts in continental crust, commonly with bimodal volcanism	Rio Grande Baikal Rhine-Bresse Graben	Keeweenawan (Precambrian) Karoo (Jurassic) Viking and Central Grabens (Mesozoic)
	Proto-oceanic rift troughs	Incipient ocean basins floored by new oceanic crust, flanked by young rifted continental margins	Red Sea Gulf of California	East Greenland (Jurassic)
Intraplate settings	Continental rises and terraces	Mature rifted intraplate continental margins at continental–oceanic boundary	East coast, USA	Early Paleozoic of USA and Canadian Cordillera
	Continental embankments	Progradation of sedimentary wedges at edge of rifted continental margins	Mississippi, Gulf Coast, USA	Early Paleozoic Meguma terrane, Canadian Appalachians (?)
	Intracratonic basins	Broad cratonic basins, commonly with underlying rifts	Chad Basin (Cenozoic) Congo Basin	Paleozoic Michigan Basin Illinois Basin Williston Basin (USA)
	Continental platforms	Stable cratons with thin, extensive sedimentary cover	Barents Sea	Middle Paleozoic, North American Midcontinent
	Active ocean basins	Basins floored by oceanic crust at active divergent plate boundaries	Pacific Ocean	Various ophiolite-bearing complexes (Semail, Oman), Neoproterozoic Arabian Shield
	Oceanic islands, aseismic ridges and plateaus	Sedimentary aprons and platforms in intra-oceanic settings	Emperor–Hawaii seamounts	Mesozoic Snow Mountain Volcanic Complex (Franciscan, California)
	Dormant ocean basins	Basins floored by oceanic crust, neither spreading nor subducting	Gulf of Mexico	Phanerozoic Tarim Basin (China)
Convergent settings	Trenches	Deep troughs formed by subduction of oceanic lithosphere	Chile Trench	Cretaceous, Shumagin Island (Alaska)
	Trench–slope basins	Structurally confined basins on subduction complexes	Central America Trench	Cretaceous Cambria slab (California)
	Forearc basins	Basins within arc–trench gaps	Sumatra	Cretaceous Great Valley (California)
	Intra-arc basins	Basins along arc platform, including superimposed and overlapping volcanoes	Lago de Nicaragua	Early Jurassic, Sierra Nevada (California)

Table 1.1 *Continued*

Relative plate movement	Basin type	Basin description	Modern example	Ancient example
	Backarc basins	Oceanic basins behind intra-oceanic magmatic arcs, and continental basins behind continental margin magmatic arcs without foreland fold–thrust belts	Marianas	Jurassic Josephine ophiolite (California)
	Retroarc foreland basins	Foreland basins on continental sides of continental margin arc–trench systems	Andes foothills	Cretaceous Sevier foreland (Wyoming-Utah)
	Remnant ocean basins	Shrinking ocean basins between colliding continental margins and/or arc–trench systems (eventually subducted or deformed)	Bay of Bengal	Pennsylvanian-Permian Ouachita Basin
	Peripheral foreland basins	Foreland basins superimposed on rifted continental margins during continental collision	Persian Gulf Indo-Gangetic Plain Po Basin (Italy)	Tertiary North Alpine Foreland Basin (Switzerland)
	Piggy-back (thrust sheet top) basins	Basins carried above moving thrust sheets	Peshawar Basin (Pakistan)	Neogene, Apennines (Italy) Meso-Hellenic Trough (Greece)
	Foreland intermontane basins (broken forelands)	Basins formed among basement cored uplifts in foreland settings	Sierras Pampeanas (Argentina)	Laramide basins (USA)
Transform settings	Transtensional basins	Basins formed by local extension along strike–slip fault systems	Salton Sea, California	Carboniferous Magdalen Basin (Gulf of St Lawrence)
	Transpressional basins	Basins formed by local compression along strike–slip fault systems	Santa Barbara Basin (California)	Miocene Ridge Basin (California)
	Transrotational basins	Basins formed by rotation of crustal blocks about vertical axes within strike–slip fault systems	Western Aleutian forearc	Miocene Los Angeles Basin, California
Hybrid settings	Intracontinental wrench basins	Basins on continental crust associated with strike–slip tectonics caused by distant collisional processes	Qaidam Basin (China)	Pennsylvanian-Permian Taos Trough (New Mexico)
	Aulacogens	Former failed rifts reactivated during convergent tectonics	Mississippi Embayment	Paleozoic Anadarko Basin (Oklahoma)
	Impactogens	Rift basins caused by stresses transmitted from convergent plate margin	Baikal Rift (Siberia)	Rhine Graben (Europe)
	Successor basins	Basins in intermontane settings following cessation of orogenic activity	Southern Basin and Range (Arizona)	Paleogene Sustut Basin (British Columbia)

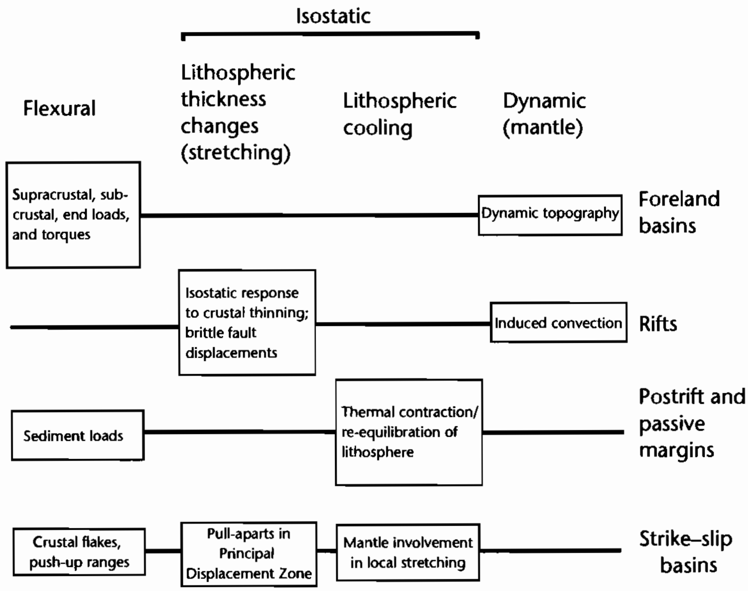

Fig. 1.11 Fundamental mechanisms of basin formation: flexural, isostatic, and dynamic. The importance of these mechanisms in foreland, rift, postrift and passive margin, and strike–slip basins is indicated by the size of the boxes.

the excess in mass of the magmatic arc. In the case of foreland basins it is a combination of topographic loads represented by the mountain belt, lateral density variations of lithospheric material (caused, for example, by prior stretching of the continental lithosphere in the overridden plate or by obduction of dense mantle flakes) and horizontal "in-plane" forces set up by the shallow "end-on" collision of buoyant continental lithosphere.

Processes within the mantle have an important role in basin development. Mantle plumes originating at the core–mantle boundary impinge on the base of the lithosphere and spread out laterally over a length scale of 10^3 km. They are instrumental in continental splitting and the formation of new ocean basins, and have a major role in igneous underplating and isostatic regional uplift. Present day *hotspots* over plumes or ascending limbs of mantle convection systems are characterized by topographic doming and commonly rifting. Basins located over downward limbs of convection systems, *cold-spot basins*, appear to be broad, gentle sags. The onset of subduction of cold oceanic slabs at ocean–continent boundaries causes far-field tilting of the continental plate towards its margin and therefore has a potentially major effect in forearc, intra-arc, and retroarc settings. In addition, it is believed that supercontinental assemblies in the geological past have experienced very long wavelength topographic doming caused by elevated sublithospheric temperatures generated by an overlying insulating lid.