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Part I

Types of Sedimentary Basins

1 Basin Classification and Depositional Environments (Overview)

1.1 Introduction

Sedimentary basins are, in a very broad sense, all those areas in which sediments can accumulate to considerable thickness and be preserved for long geological time periods. In addition, there also exist areas of long-persisting denudation, as well as regions where erosional and depositional processes more or less neutralize each other (creating what is known as non-deposition or omission).

In plan view sedimentary basins can have numerous different shapes; they may be approximately circular or, more frequently, elongate depressions, troughs, or embayments, but often they have quite irregular boundaries. As will be shown later, even areas without any topographic depression, such as alluvial plains, may act as sediment traps. The size of sedimentary basins is highly variable, though they are usually at least 100 km long and tens of km wide.

We can distinguish between (1) active sedimentary basins still accumulating sediments, (2) inactive, but little deformed sedimentary basins showing more or less their original shape and sedimentary fill, and (3) strongly deformed and incomplete former sedimentary basins, where the original fill has been partly lost to erosion, for example in a mountain belt.

As many workers have pointed out, the regional deposition of sediments, non-deposition, or denudation of older rocks are controlled mainly by tectonic movements. Hence, most of the recent attempts to classify sedimentary basins have been based on concepts of global and regional tectonics which will be briefly discussed below. In spite of obvious advantages, however, this approach has some serious shortcomings if it is not supplemented by additional criteria. One ought always bear in mind that the characteristics of sediments filling a basin of a certain tectonic type are pre-dominantly controlled by other factors and can be extremely variable. With few exceptions (also discussed later), there is hardly such a phenomenon as a "tectonic sedimentary facies". For example, the broad concept of "geosynclinal sediments", often postulated in the past, was more misleading than helpful.

In addition to tectonic movements in the basinal area itself, sedimentary processes and facies are controlled by the paleogeography of the regions around the basin (peri-basin morphology and climate, rock types and tectonic activity in the source area), the depositional environment, the evolution of sediment-producing organisms, etc. Many sedimentologists therefore prefer a classification scheme based mainly on criteria which can be recognized in the field, i.e., the facies concept and the definition of the depositional environment (fluvial sediments, shelf deposits etc.). A further approach is the subdivision of sediments into important lithologic groups, such as siliciclastic sediments of various granulometries and composition, carbonate rocks, evaporites, etc. Having established the facies, succession, and geometries of such lithologic groups, one can proceed to define the tectonic nature of the basin investigated.

In this book an attempt is made to combine some principal points of these different classification systems and to show the interaction between tectonic and environmental characteristics of depositional areas.

1.2 Tectonic Basin Classification

Basin-generating tectonics is the most important prerequisite for the accumulation of sediments. Therefore, a tectonic basin classification system is briefly introduced at the beginning of this chapter. Such a classification must be in accordance with the modern concept of global plate tectonics and hence will differ from older classifications and terminology.

In recent years, several authors have summarized our current knowledge on the interaction of plate tectonics and sedimentation (e.g., Dickinson in Dickinson and Yarborough 1976; Kingston et al. 1983; Miall 1984; Mitchell and Reading 1986; Foster and Beaumont 1987; Klein 1987; Perrodon 1988) and proposed basin classification systems. Although basically identical, these systems differ somewhat and do not use exactly the same terms. In this text we essentially use the system described by Mitchell and Reading, but add some minor modifications.

The different types of sedimentary basins can be grouped into seven categories, which in turn may be subdivided into two to four special basin types (Table 1.1 and Figs. 1.1 through 1.4):

Intracratonic or interior sag basins (Fig. 1.1a). Basins on continental crust are mainly generated by divergent plate motions and resulting extensional structures and thermal effects (cf. Sect. 8.1). In the case of large interior sag basins, however, major fault systems forming the boundaries of the depositional area or a central rift zone may be absent. Subsidence occurs predominantly in response to moderate crustal thinning or to a slightly higher density of the underlying crust in comparison to neighboring areas. In addition, slow thermal decay after a heating event and sedimentary loading can promote and maintain further subsidence for a long time (Sect. 8.1). Alternatively, it was recently suggested that long-term subsidence of intracratonic basins may be related to a decrease of the mantle heat flow (abnormal cooling) above a "cold spot" (Ziegler 1989). In general, rates of subsidence are low in this geodynamic setting (cf. Sect. 12.3).

Continental graben structures and rift zones form narrow elongate basins bounded by large faults (Fig. 1.1b and c). Their cross sections may be symmetric or asymmetric (e.g., halfgrabens, see Sects. 11.4 and

Table 1.1 Tectonic basin classification. (After Kingston et al. 1983; Mitchell and Reading 1986)

Basin category	Special basin type or synonym(s)	Underlying crust	Style of tectonics	Basin characteristics
Continental or interior sag basins	Epicontinental basins, intra-cratonic basins	Continental	Divergence	Large areas, slow subsidence
Continental or interior fracture basins	Graben structures, rift valleys and rift zones, aulacogens	Continental	Divergence	Relatively narrow basins, fault-bounded, rapid subsidence during early rifting
Basins on passive continental margins, margin sag basins	Tensional-rifted basins, tension-sheared basins, sunk margin basins	Transitional	Divergence + shear	Asymmetric basins partly outbuilding of sediment, moderate to low subsidence during later stages
Oceanic sag basins	Nascent ocean basin (growing oceanic basin)	Oceanic	Divergence	Large, asymmetric, slow subsidence
Basins related to subduction	Deep-sea trenches	Oceanic	Convergence	Partly asymmetric, greatly varying depth and subsidence
	Forearc basins, backarc basins, interarc basins	Transitional, oceanic	Dominantly divergence	
Basins related to collision	Remnant basins	Oceanic	Convergence	Activated subsidence due to rapid sedimentary loading
	Foreland basins (Peripheral), retroarc basins (intramontane), broken foreland basins,	Continental	Crustal flexuring, local convergence or transform motions	Asymmetric basins, trend to increasing subsidence, uplift and subsidence
	Terrane-related basins	Oceanic		Similar to backarc basins
Strike-slip/ wrench basins	Pull-apart basins (transtensional) and transpressional basins	Continental and/or oceanic	Transform motion, ± divergence or convergence	Relatively small, elongate, rapid subsidence

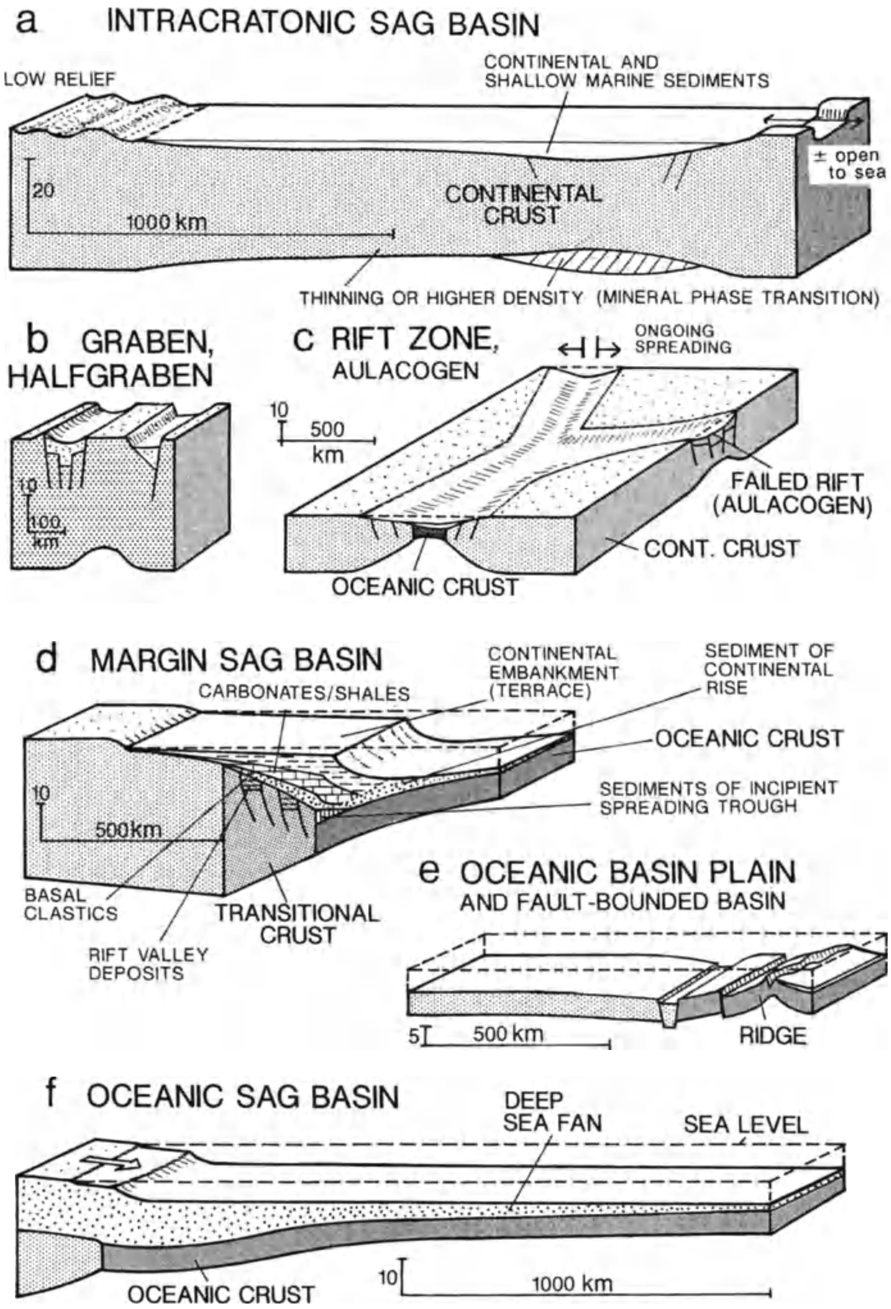

Fig. 1.1a-f. Tectonic classification for extensional basins on continents, continental margins, and on oceanic crust. See text for explanation. (After Dickinson and Yarborough 1976; Kingston et al. 1983; Mitchell and Reading 1986)

12.1). If the underlying mantle is relatively hot, the lithosphere may expand and show updoming prior to or during the incipient phase of rifting. Substantial thinning of the crust by attenuation, which is often accompanied by the upstreaming of basaltic magma, thus forming transitional crust, causes rapid subsidence in the rift zone. Subsequent thermal contraction due to cooling and high sedimentary loading enable continuing subsidence and therefore the deposition of thick sedimentary infillings.

Failed rifts and aulacogens (Fig. 1.1c). If divergent plate motion comes to an end before the moving blocks are separated by accretion of new oceanic crust, the rift zone is referred to as "failed". A certain type of such failed rifts is an aulacogen. Aulacogens represent the failed arm of a triple junction of a rift zone, where two arms continue their development to form an oceanic basin. Aulacogen floors consist of transitional crust, which may include some oceanic crust, and allow the deposition of thick sedimentary sequences over relatively long time periods. These sedimentary fills are

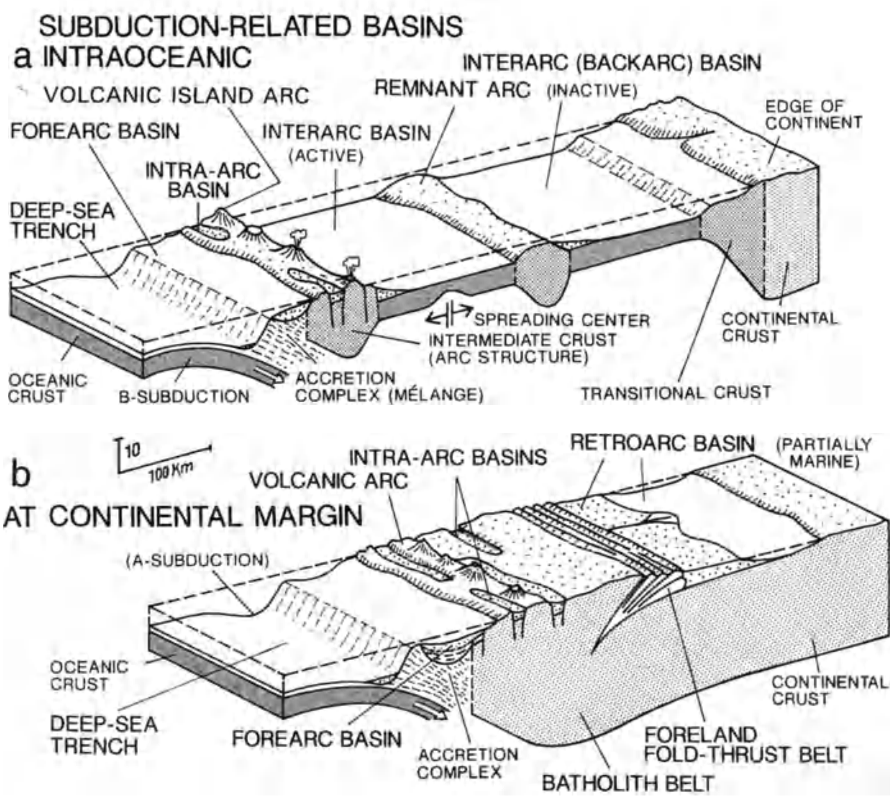

Fig. 1.2a,b. Tectonic classification of subduction related basins (Fig. 1.1 continued). See text for explanation

often affected by subsequent convergence along fault zones. Basins similar to aulacogens may also be initiated during the closure of an ocean and during orogenies.

Passive margin basins (Fig. 1.1d). The initial stage of a true oceanic basin setting (or a proto-oceanic rift system) is established when two divergent continents separate and new oceanic crust forms in the intervening space. This does not necessarily mean that such a basin type fills with oceanic sediments, but it does imply that the central basin floor lies at least 2 to 3 km below sea level. When such a basin widens due to continued divergent plate motions and accretion of oceanic crust (drifting stage), its infilling with sediment tends to more or less lag behind ocean spreading. Terrigenous sediments are deposited predominantly along the two continental margins of the growing ocean basin. The marginal "basins" are commonly not bordered by morphological highs. They develop on top of seaward thinning continental crust with seaward increasing (differential) subsidence (Sect. 8.4). They therefore represent asymmetric depositional systems in which sediments commonly build up in the form of a prism (Fig. 1.1d and Sect. 12.2). Some of these marginal basins may be affected and bordered by transform motions (tension-sheared basins). In a sediment-starved environment, subsided transitional crust can create deep plateaus (sunk basins). In general, subsidence of these marginal basins tends to decrease with passing time, unless it is reactivated by heavy sediment loads.

Oceanic sag basins or nascent ocean basins occupy the area between a mid-oceanic ridge, including its rise, and the outer edge of the transitional crust along a passive continental margin (Fig. 1.1f). They commonly accumulate deep-sea fan or basin plain sediments. Due to the advanced cooling of the aging oceanic crust, subsidence is usually low, unless it is activated by thick sedimentary loading near the continental margin. Fault-bounded basins of limited extent are common in conjunction with the growth of mid-oceanic ridges (Fig.1.1e).

Basins related to subduction. Another group of basins is dominated by convergent plate motions and orogenic deformation. Basins related to the development of subduction complexes along island arcs or active continental margins include deep-sea trenches, forearc basins, backarc basins (Fig. 1.2a and b), and smaller slope basins and intra-arc basins.

Deep-sea trench floors are composed of descending oceanic crust. Therefore, some of them represent the deepest elongate basins present on the globe. In areas

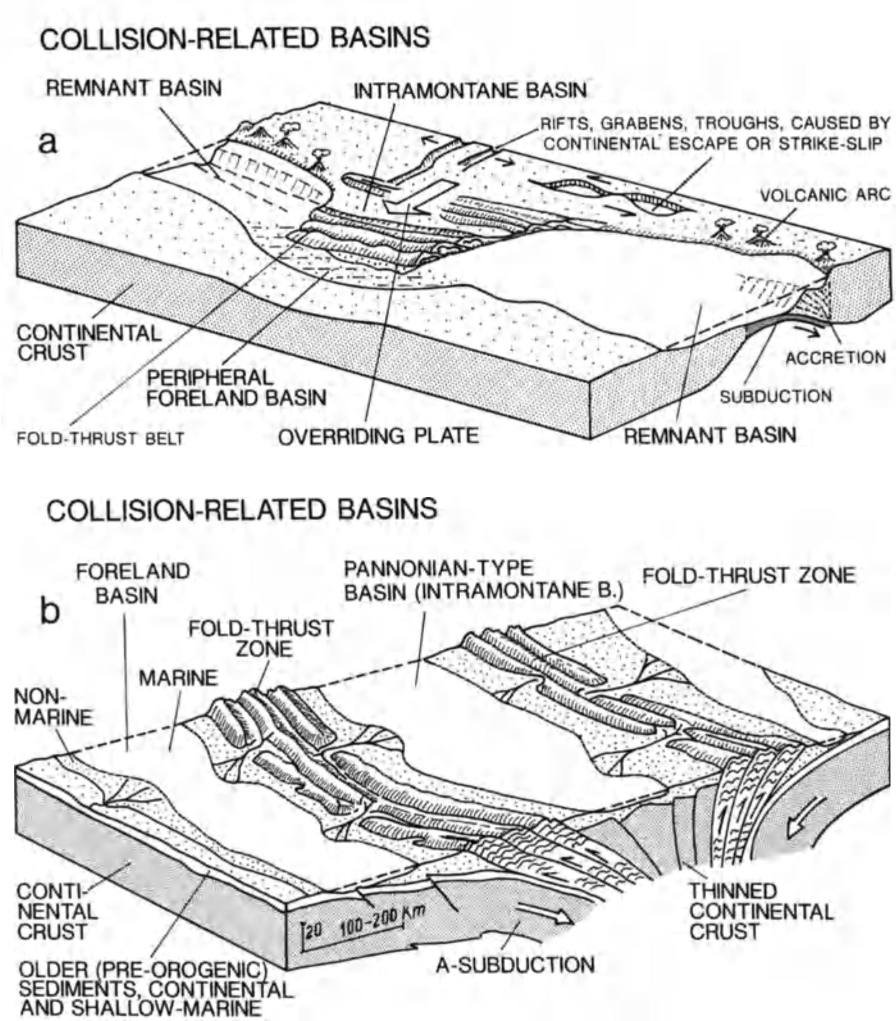

Fig. 1.3a,b. Tectonic classification of collission-related basins (Fig. 1.2 continued). See text for explanation

of very high sediment influx from the neighboring continent, however, they are for the most part filled up and morphologically resemble a continental rise. Deep-sea trenches commonly do not subside as do many other basin types. In fact, they tend to maintain their depth which is controlled mainly by the subduction mechanism, as well as by the volume and geometry of the accretionary sediment wedge on their landward side (Sect. 12.5.2).

Forearc basins occur between the trench slope break of the accretionary wedge and the magmatic front of the arc. The substratum beneath the center of such basins usually consists of transitional or trapped oceanic crust older than the magmatic arc and the accretionary subduction complex (Sect. 12.5.3). Rates of subsidence and sedimentation tend to vary, but may frequently be high. Subsequent deformation of the sedimentary fill is not as intensive as in the accretionary wedge.

Backarc and interarc basins form by rifting and ocean spreading either landward of an island arc, or between two island arcs which originate from the splitting apart of an older arc system (Fig. 1.2a). The evolution of these basins resembles that of normal ocean basins between divergent plate motions. Their sedimentary fill frequently reflects magmatic activity in the arc region.

Terrane-related basins are situated between microcontinents consisting at least in part of continental crust (Nur and Ben-Avraham 1983) and larger continental blocks. The substratum of these basins is usually oceanic crust. They may be bordered by a subduction zone and thus be associated with basins related either to subduction or collision.

Basins related to collision. Partial collision of continents with irregular shapes and boundaries which do not fit each other leads to zones of crustal over- or

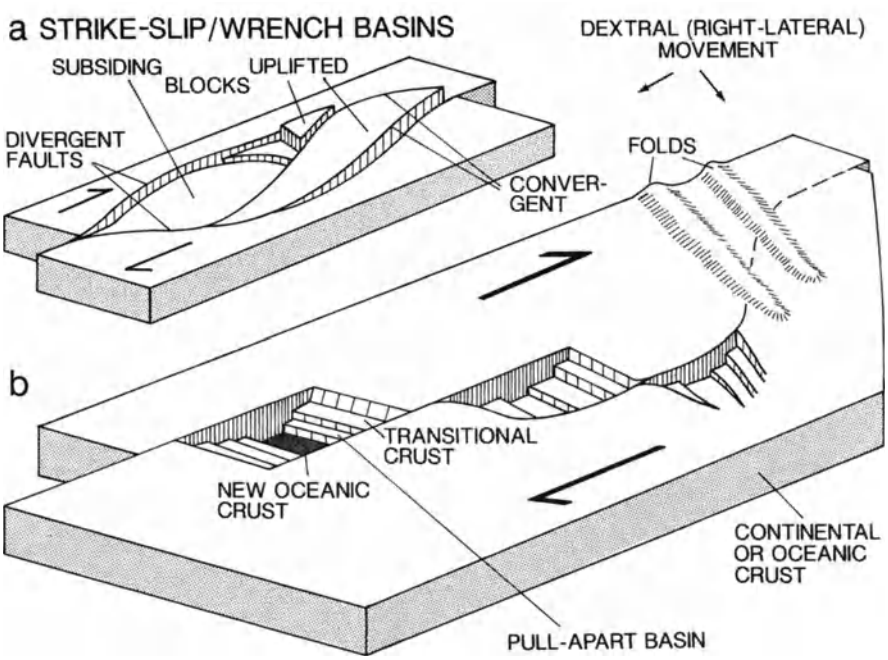

Fig. 1.4a,b. Strike-slip/wrench basins (Fig. 1.3 continued). See text for explanation

underthrusting. Along strike, however, one or more oceanic basins of reduced size may still persist (Fig. 1.3a).

These *remnant basins* tend to collect large volumes of sediment from nearby rising areas and to undergo substantial synsedimentary deformation (convergence, also often accompanied by strike-slip motions).

Foreland basins and *peripheral basins* in front of a fold-thrust belt are generated by depressing and flexuring the continental crust ("A-subduction", after Ampferer, Alpine-type) under the load of the overthrust mountain belt (Fig. 1.3a and Fig. 1.3b). The length of these asymmetric basins tends to increase with time, but influx of clastic material from the rising mountain range often keeps pace with or exceeds subsidence and thus causes basin filling (Sect. 12.6). Collision of two continental blocks may lead to "continental escape" of parts of the overriding plate as well as produce extensional graben structures or rifts perpendicular to the strike of the fold-thrust belt (Fig. 1.3a).

Retroarc or intramontane basins (Figs. 1.2b and 1.3a) occur in the hinterland of an arc orogen ("B-subduction" zone). They may affect relatively large areas on continental crust. Limited subsidence appears to be caused mainly by tectonic loading in a backarc fold-thrust belt.

Pannonian-type basins originate from post-orogenic divergence between two fold-thrust zones (Fig. 1.3b). They are usually associated with an A-subduction zone and are floored by thinning continental or transitional crust.

During crustal collision, some foreland and retroarc basins are broken up into smaller blocks, whereby strike-slip motions may also play a role (Fig. 1.3a).

Some of the blocks are affected by uplift, others by subsidence, forming basinal depressions. The mechanics of such *tilted block basins* were studied, for example, in the Wyoming Province of the Rocky Mountain foreland (McQueen and Beaumont 1989). So-called *Chinese-type basins* (Bally and Snelson 1980) result from block faulting in the hinterland of a continent-continent collision. They are not directly associated with an A-subduction margin, but it appears unnecessary to classify them as a special new basin type (Hsü 1989).

Strike-slip and wrench basins (Fig. 1.4a and b). Transform motions may be associated either with a tensional component (transtensional) or with a compressional component (transpressional). Transtensional fault systems locally cause crustal thinning and therefore create narrow, elongate pull-apart basins (Sect. 12.8). If they evolve on continental crust, continuing transform motion may lead to crustal separation perpendicular to the transform faults and initiate accretion of new oceanic crust in limited spreading centers. Until this development occurs, the rate of subsidence is usually high. Transpressional systems generate wrench basins of limited size and endurance. Their compressional component can be inferred from wrench faults and fold belts of limited extent (Fig. 1.4b).

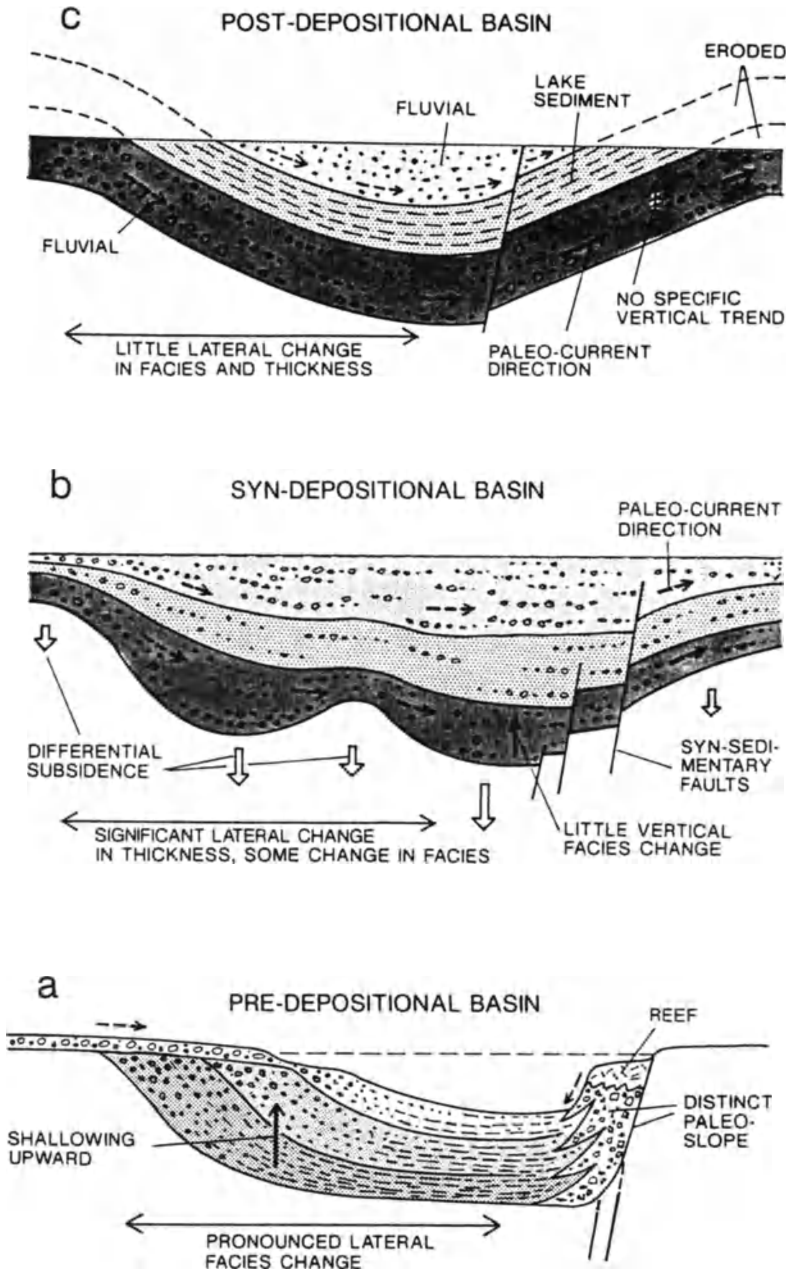

Fig. 1.5. a Rapid, pre-depositional tectonics creates a deep morphological basin which is later filled up with post-tectonic sediments. The geometry of former basin can be derived from transport directions and facies distribution. **b** Syn- depositional tectonic movements control varying thickness of fluviatil and shallow-marine sediments an generate a basin-fill structure, although a morphological basin barely existed. **c** Post-depositional basin-like structure created by tectonic movements after the deposition of sheet-like fluvial and lake sediments; part of the syntectonic basin fill is removed by subsequent erosion

In order to identify these various basin categories, one must know the nature of the underlying crust as well as the type of former plate movement involved during basin formation, i.e., divergence or convergence. Even in the case of transform movement, either some divergence or convergence must take place. Small angles of convergence show up as wrenching or fold belts, and small angles of divergence appear as normal faulting or sagging.

One should bear in mind that all these basin types represent proto-types of tectonically controlled basins. They offer a starting point for the study and evaluation of basins, but there are no type basins which can be used as a complete model for any other basin (Burchfiel and Royden 1988). Even within a single broad tectonic setting, the development of smaller individual basins may display great variation. As soon as basins are analyzed in greater detail, the broad tectonic basin classification listed above becomes less useful. In addition, over long time periods, a sedimentary basin may evolve from one basin type into another (polyhistory basins) and thus exhibit a complex tectonic and depositional history (Sect. 12.9).

1.3 Tectonics and Basin-Filling

Although basin-generating tectonic movements and basin-filling depositional processes generally interact, one can distinguish three different modes (partially end members) of this relationship (Fig. 1.5, based on Selley 1985a):

Pre-depositional basins. Rapid tectonic movements predate significant sediment accumulation and create a morphological basin, which is filled later by post-tectonic sediments (Fig. 1.5a). The water depth in the basin decreases with time, although some syn-depositional subsidence due to sediment loading is likely (Sect. 8.1). Sediment transport as well as vertical and lateral facies associations are substantially influenced by the basin morphology.

Syn-depositional basins. Sediment accumulation is affected by syn-depositional tectonic movements, e.g., differential subsidence (Fig. 1.5b). If the sedimentation rate is always high enough to compensate for subsidence, the direction of transport and the sedimentary facies largely remain unchanged, but the thicknesses of certain time slices varies. In Figure 1.5b they increase toward the center of the basin. In this case, the basin structure is syn-depositional, but there was hardly a syn-depositional morphological basin controlling the sedimentary facies of the basin. If sedimentation is too slow to fill up the subsiding area, a morphological basin will develop. Then, the distribution and facies of the succeeding sediments will be affected by the morphology of the deepening basin (transition to the situation shown in Fig. 1.5a).

Post-depositional basins. The deposition of sediments largely predates tectonic movements forming a distinct basin structure. Hence, there is no or little relationship between the transport, distribution, and facies of these sediments and the later evolved basin structure (Fig. 1.5c). In most cases, however, some relationship between a syn-depositional subsidence phase and the subsequent tectonic overprint cannot be excluded.

Of course, there are transitions between these simplified models of the interaction between basin-generating tectonics and basin-filling processes (see Chap. 12). Certain basins may show a complex history and therefore contain sediments affected by both pre- and syn-tectonic movements.

1.4 Basin Morphology and Depositional Environments

1.4.1 General Aspects

The geometry of an ultimate basin fill is controlled mainly by basin-forming tectonic processes, but the *morphology of a basin* defined by the sediment surface is the product of the interplay between tectonic movements and sedimentation. Therefore, as already mentioned, a purely tectonic classification of sedimentary basins is not sufficient for characterizing depositional areas. It is true that a sedimentary basin in a particular tectonic setting often experiences a specific development and subsidence history (Chaps. 8 and 12), but its morphology, including water depth, may be controlled largely by other factors, such as varying influx and distribution of sediment from terrigenous sources (Chap. 11).

For example, a fluvial depositional system can develop and persist for considerable time on top of subsiding crust in various tectonic settings (Miall 1981). Fluvial deposits are known from continental graben structures, passive continental margins, foreland basins, forearc and backarc basins, pull-apart basins, etc. Fluvial sediments accumulate as long as rivers reach the depositional area and supply enough material to keep the subsiding basin filled. Although the basin-forming processes and subsidence histories of these examples differ fundamentally from each other, the sedimentary facies of their basin fills display no or only minor differences. In order to distinguish between these varying tectonic settings, one has to take into account the geometry of the entire basin fill, as well as vertical and lateral facies changes over long distances, including paleocurrent directions and other criteria. Syndepositional tectonic movements manifested by variations in thickness, small disconformities, or faults dying out upward (cf. Fig. 1.5b) may indicate the nature of the tectonic processes involved.

The *erosional base level* and hydrographic regime within a basin are additional important factors controlling sediment dispersal and modifying basin morphology. They largely determine the development of special sedimentary facies as demonstrated in the elementary model of Fig. 1.6. In a fluvial environment, sediments cannot accumulate higher than the base level of erosion and the elevation added by the gradient of the stream. If there is more influx of material into the depositional system than necessary for compensation of subsidence, the sediment surplus will be carried farther downstream into lakes or the sea.

This signifies that the level up to which a basin can be filled with sediments may depend on the geographic position of the basin in relation to the erosional base. In Tibet, for example, the floors of present-day fluvial basins (intramontane basins and graben structures) are elevated higher than 3000 m in comparison to the coastal fluvial plains elsewhere.

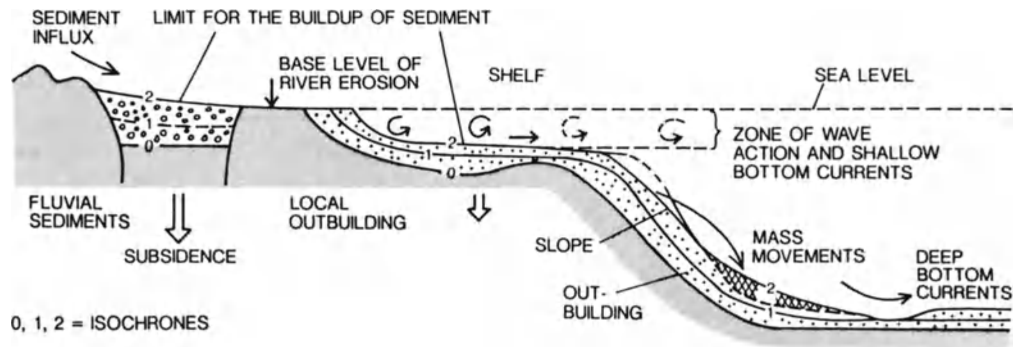

Fig. 1.6. Base level of erosion, hydrodynamic regime in the sea, and gravity mass movements as limiting factors controlling upbuilding and outbuilding of sediments. Note that the model may be modified by sea level changes

The morphology of water-filled basins may significantly change as a result of depositional processes. Lakes and low-energy marine basins frequently show prograding deltaic facies, causing pronounced basinward outbuilding of sediment (Sects. 2.1.1, 2.2.2, 2.5.1, 3.4.1). Consequently, the area occupied by the deeper basin, where finer-grained sediment is deposited, decreases with time, although the initial, tectonically controlled basin configuration persists. By contrast, high-energy basins are less influenced by sediment outbuilding (Fig. 1.6).

For example, terrigenous sediments transported into high-energy shelf seas tend to be reworked and swept into deeper water by wave action and bottom currents, except for some local seaward migration of the shoreline. Even on deep submarine slopes and in the deep sea, there is no general outbuilding or upbuilding of sediments, because gravity mass movements and deep bottom currents redistribute large quantities of material.

These few examples demonstrate that the most appropriate classification scheme for sedimentary basins depends primarily on the objectives of the study. If tectonic structure and evolution of a region are the main topics, then basin fill geometry and subsidence history derived from the thickness of stratigraphic units (Sect. 8.4) are of primary importance. If, on the other hand, the depositional environment, sedimentary facies, and paleogeographic reconstructions are of major interest, then the basin classification used should not be strictly tectonic. Such a classification should also take into account changes in basin morphology caused by depositional processes, the chemical and hydrodynamic regimes of the basin, and peri-basin characteristics such as the size and nature of the drainage areas on land delivering terrigenous and dissolved materials into the basin.

1.4.2 Different Methods in the Study of Modern and Ancient Sediments

Many workers distinguish between recent and ancient examples of depositional environments (e.g., Davis 1983; Reading 1986a), because the interpretation of spaleoenvironments from the fossil record is subject to greater uncertainties. Furthermore, the methods of investigation and the possibilities of observing certain physical and biological sedimentary structures differ between soft sediments and lithified rocks. Soft material, for example, is suitable for the determination of primary grain size distribution, which in the case of lithified rocks is frequently problematic. On the other hand, any kind of structure is commonly much better visible in ancient rocks than in soft sands and muds. The surface of recent sediments on land and under water can be well observed, but in many cases, for example in fluvial environments, such temporary surfaces are rarely preserved in the sedimentary record. In contrast, indurated beds alternating with weaker material frequently show excellently preserved lower and upper bedding planes with trace fossils, various marks, and imbrication phenomena which are difficult to observe in soft sediments. Diagenesis may, however, also obscure primary bedding features. In addition, there are special sediments in the past, particularly far back in the Earth's history, for which no present-day analogies are known. Such environments are mentioned in Section 6.5.

In spite of such various problems between recent and ancient sediments, the depositional environments of both groups are treated jointly in this book, except for some special deposits. After a brief overview in this chapter, the most important groups of depositional environments are described in simplified facies models in Chapters 2 through 6.

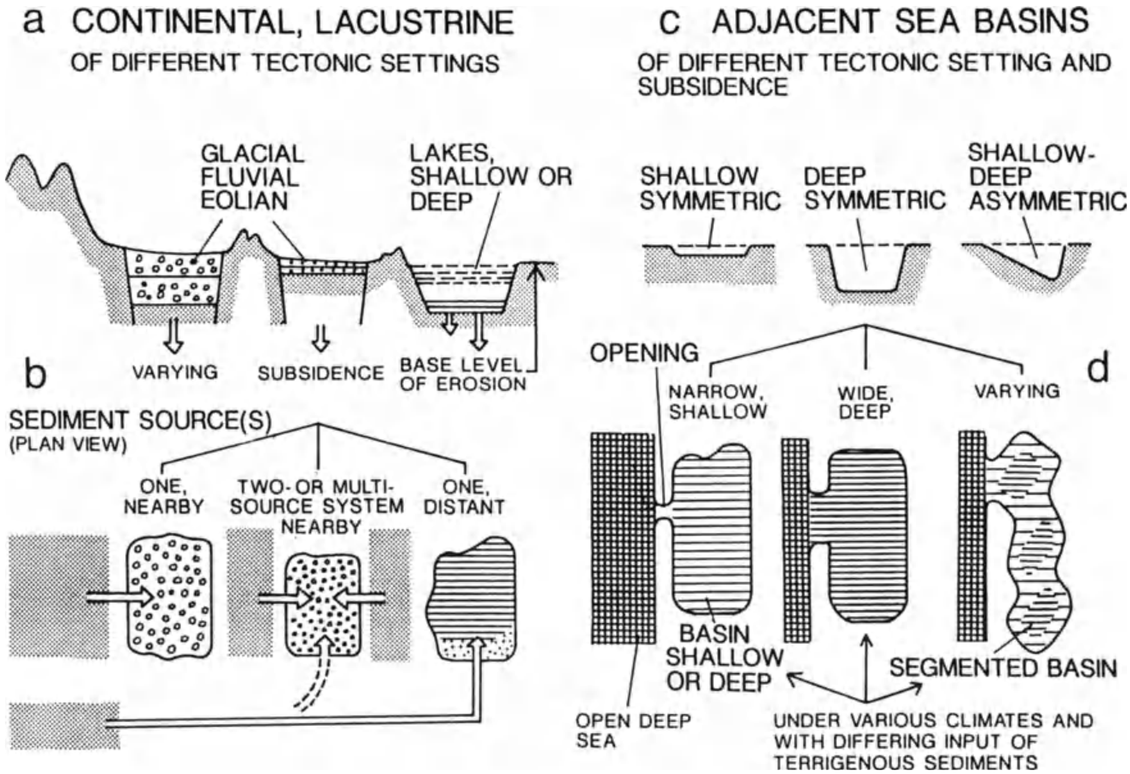

Fig. 1.7a-d. Overview of various depositional environments, based primarily on basin morphology (**a** and **c**) and peri-basin characteristics (**b** and **d**). All these basins are strongly affected by variations in terrigenous input under differing conditions of relief and climate. For further explanation see text

1.4.3 Depositional Environments (Overview)

On the surface of our present-day globe, on land and below the sea, hundreds of depositional areas are known which meet the definition of sedimentary basins as described in Section 1.1. If we add to this list medium to large ancient sedimentary basins whose fill is still largely preserved, we have some thousand sedimentary basins. Taking into account this large number and the many factors controlling a sedimentary environment, it appears at first glance that an enormous number of differing depositional environments should exist. This is in fact the case, but nevertheless it is possible to subdivide this great quantity into a limited number of distinct groups which have many characteristics in common.

Such depositional environment models have been extensively described in several textbooks (e.g., Reineck and Singh 1980; Blatt et al. 1980; Walker 1984a; Walker and James 1992; Selley 1985a and b; Reading 1986a, 1996; Boggs 1987), and single groups of environments have been dealt with repeatedly in special publications, memoirs, short course notes, etc. (cf. Chaps. 2 and 3).

In Figures 1.7 and 1.8 the various types of sedimentary basins are predominantly classified according to their depositional environment and basin morphology. However, peri-basin geomorphology and climate also play a role. One can distinguish between several principal groups, for example:

- Continental (fluvial, glacial, eolian), lacustrine, and deltaic environments.
- Adjacent sea basins and epicontinental seas of varying salinity.
- Marine depositional areas of normal salinity.

As an alternative, a group of "transitional" environments may be defined between continental and marine environments (e.g., Davis 1983). This group includes marine deltas, intertidal environments, coastal lagoons, estuaries, and barrier island systems (cf. Chap. 3). In Part II of this book, a more diversified classification is used with the following main groups:

- Continental sediments.
- Coastal and shallow sea sediments (including carbonates).
- Sediments of adjacent seas and estuaries.
- Oceanic sediments.
- Special sediments and environments.

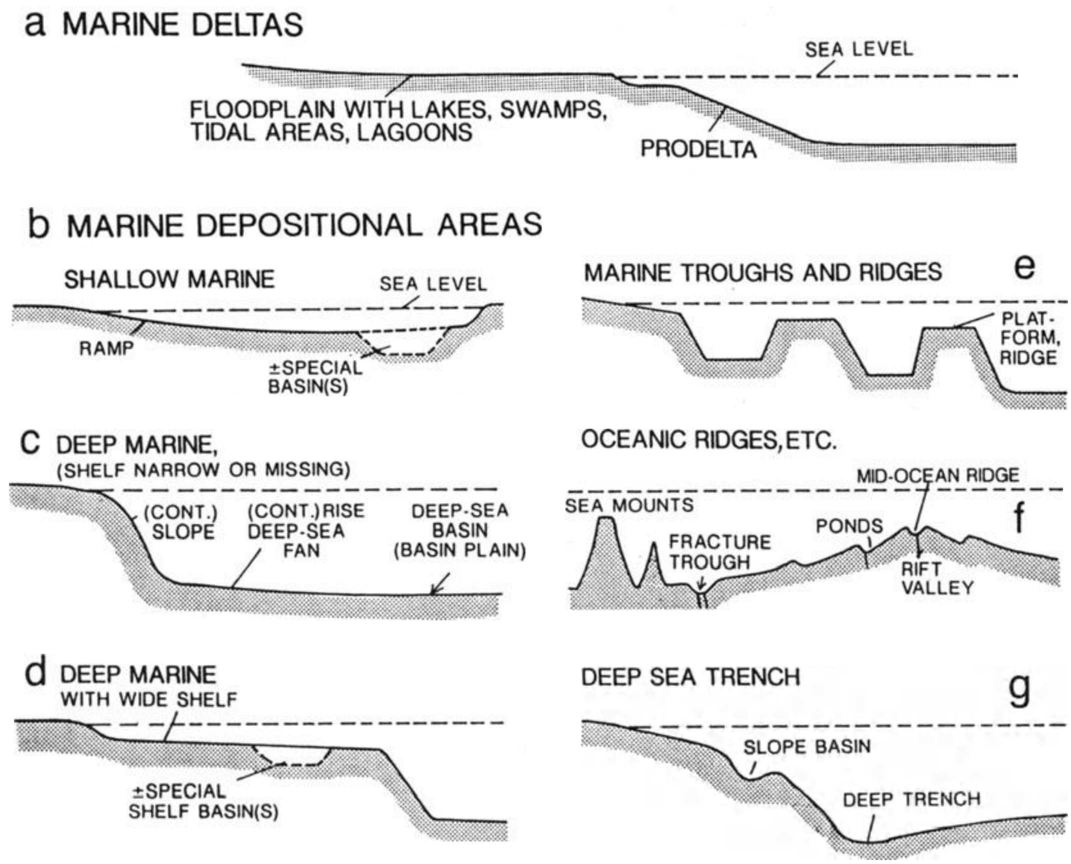

Fig. 1.8a-g. Marine delta (**a**) and overview of other marine basins (**b** through **g**). For further explanation see text

In addition, Chapter 7 deals with depositional rhythms and cyclic sequences which may occur in all groups of depositional environments.

1.4.4 Elementary Principles for Basin Filling

The *fluvial environment* is controlled by its erosional base level and sediment supply from more elevated regions. As long as sediment supply is sufficient to compensate for subsidence, regardless of the type of tectonic setting, the river gradient and thus a more or less constant average net transport direction through the fluvial basin can be maintained (Fig. 1.7a), and the sedimentary facies does not change significantly. A topographic depression, i.e., a syn-depositional morphological basin (Fig. 1.5b) can only develop when fluvial transport lags behind basin subsidence.

This clear relationship between gradient and transport direction is modified in the *glacial and eolian environments*. Subglacial abrasion often leads to erosional depressions, over-deepened valleys, and ice-filled troughs, which are later filled with water creating short-lived lakes. Similarly, eolian deflation can generate local depressions in the land surface which, if the groundwater table rises, may be transformed into salt pans. However, such erosional features are normally filled up again with sediments within a short time span. On the other hand, eolian sand can accumulate large "sand seas" reaching elevations well above the surrounding landscape. In addition, wind-blown sand and dust can migrate into different directions, partially upslope.

The influence of peri-basin morphology on *fluvial-lacustrine sedimentation* is described in Figure 1.7b. Terrigenous material entering the basin may come from nearby or distant sources. Consequently, the sediment will be texturally immature or markedly mature. Similarly, its mineralogical composition may be either fairly uniform or mixed. In addition, the climate in the source area(s) exerts a strong influence (Sect. 2.2.4). Where sediment accumulation cannot compensate for subsidence, long persisting, deepening lakes or shallow seas evolve (see below).

Marine deltas represent a transitional, highly variable depositional environment between continental and marine conditions (Fig. 1.8a). The subaerial part of such a delta is controlled by fluvial and possibly lacustrine processes, whereas its coastal and subaqueous regions are dominated by the hydrodynamic and chemical properties of the sea. Large terrigenous sediment supply causes prograding of the deltaic complex toward the sea; high sedimentation rates and subsidence enhanced by the sediment load enable the formation of thick, widely extended deltaic sequences. Marine delta complexes provide a particularly good example of depositional environments which are controlled predominantly by various exogenic factors (Sect. 3.4).

Adjacent sea basins and *epicontinental seas* are connected with the open sea and therefore exchange water with the ocean (Fig. 1.7c and d). The extent of this water exchange and thus the salinity of the basin water strongly depend on the width and depth of the opening to the ocean. In humid regions, adjacent basins with a limited opening tend to develop brackish conditions, while arid basins frequently become more saline than normal sea water (Sect. 4.2). Adjacent basins and epicontinental seas on continental crust are commonly shallow, but basins on oceanic or mixed crust may also be deep. The shapes of these basins largely vary; some of them show symmetric or asymmetric cross sections; some represent basins subdivided by shallow swells into several subbasins (segmented basins). In the latter case, markedly differing depositional subenvironments have to be taken into account. Most of these adjacent basins are strongly influenced by the climate and relief of the peri-basin land regions (cf. Fig. 1.7b), which control the influx of terrigenous material from local and/or distal sources. In summary, adjacent basins may exhibit a particularly great variety of sedimentary facies (Sect. 4.3).

The sediments of *shallow seas and continental shelves (Fig. 1.8b)* are also considerably affected by processes operating in neighboring land regions. These generally provide sufficient material to keep these basins shallow. Strong waves, surface and bottom currents usually tend to distribute the local influx of terrigenous sediment over large areas. Especially in shallow water, high-energy conditions prevent the deposition of fine-grained materials, partially including sands. Therefore, such environments often persist over long time periods without being filled up to sea level (Fig. 1.6). This is also true for widely extended shallow-marine basins, as long as excess sediment volume (in relation to space provided by subsidence) can be stored in special depressions (Fig. 1.8b and d) or be swept into a neighboring deeper ocean basin. The margin of such basins is commonly characterized by a kind of ramp morphology.

Deeper marine basins are usually bordered by a shelf zone of varying width followed by a wide and normally gentle slope (continental slope, Fig. 1.8c). The foot of the slope in deep water (continental rise) is still gently inclined basinward; it is built up to a large extent by redeposited material derived directly from the slope (slope apron) or by sediments funnelled by submarine valleys and canyons into the deep sea (deep-sea fans). The terms *continental slope* and *continental rise* are commonly used to describe corresponding features of the present-day passive, Atlantic-type continental margins. These terms, however, imply a plate-tectonic interpretation. *Deep-sea basins* or basin plains are the deepest parts of marine environments except deep-sea trenches and some other special features associated with the behavior of oceanic crust (see below).

Large volumes of terrigenous material can be collected by the troughs in a *submarine horst and graben topography* bordering the continent (Fig. 1.8e). Similarly, deep-sea trenches at the foot of relatively steep slopes and slope basins are sites of preferential sediment accumulation (Fig. 1.8g). Thick, ancient flysch sequences are mostly interpreted as depositions in such basins. Less important sediment accumulation systems are small basins ("ponds"), which occur along oceanic ridges, and infillings of narrow troughs due to fracturing of the oceanic crust (Fig. 1.8f).

The thin, frequently incomplete sedimentary records on the tops of *submarine ridges, platforms, and seamounts (Fig. 1.8f)* strongly contrast with all other marine sediments. These deposits are mostly biogenic, chemically or biochemically precipitated and usually contain only very small proportions of terrigenous or volcanoclastic materials. Although such limited sediment accumulations can hardly be referred to as basin fills, they do constitute an important and diagnostically significant part of larger marine depositional environments.

The direct influence of tectonic basin evolution on sedimentary facies is only evident in areas, where tectonic movements are rapid and nonuniform, such as at the basin margins, or where sediment accumulation lags far behind subsidence faulting, or thrusting. This situation is common in *continental rift and pull-apart basins* during their early stages of evolution, in subduction-related settings, in remnant and foreland basins, and in deep marine environments along oceanic ridges or transform faults far away from large land masses. These problems are further discussed in Chapter 12.

1.4.5 Some General Trends for Sediment Accumulation and Facies

From the previous discussion one can draw some general, straightforward rules for the sediment accumulation and facies in various depositional environments:

- The influence of terrigenous sediment sources on basin fillings decreases in the following order: high-relief continental environments - lowlands - shallow seas - deep sea.

- Similarly, the sedimentation rate tends to decrease from highland continental basins to the central parts of large oceanic basins.
- Basins with low sedimentation rates tend to accumulate sediments relatively rich in biogenic components. Such basins may persist for long time periods and are therefore often markedly affected by synsedimentary tectonic movements.
- Chemical sediments (evaporites) of some extent commonly form in lowlands (lakes) and special portions of adjacent shallow seas, but rarely in the other depositional environments.
- The sedimentary facies of many basin fills do not reflect tectonic basin evolution and specific structural elements. Only in some basin types and/or during the most rapid phase of basin evolution do tectonic movements directly control sedimentary facies. However, the geometry of basin fills, sedimentation rates, and syn- and post-depositional deformations characterize the tectonic style and evolution of the basin considered.

1.4.6 Facies Architecture

The principal sediment characteristics of the various depositional environments include features on different scales. These range from large-scale phenomena, relevant to the facies distribution in the total basin, to micro-scale properties which are studied in a single rock specimen. As Allen (1983) and Miall (1985) have pointed out, the sedimentary basin fill often displays a certain type of stratigraphic architecture, i.e., larger units are built up by a number of smaller, basic units. In single outcrops, generally only the smaller scale units can be observed, which are often not sufficiently diagnostic for the recognition of the true nature of the total basin fill or a large part of it.

The brief summaries for the common depositional environments presented in Chapters 2 through 6 are largely based on these principles. They preferentially show field and outcrop phenomena and how these fit into a larger scale facies model. Micro-scale features and processes are only described in special cases.

1.4.7 Summary (Basin Classification)

- Sedimentary basins originate from various endogenic processes and can be classified on the basis of plate tectonic models. The basin-generating forces also determine the overall geometry of the entire basin fill.
- However, the depositional environments and sedimentary facies of the basins are to a large extent controlled by exogenic processes (e.g. terrigenous sediment influx, autochthonous sediment production, hydrographic regime of the basin, sediment distribution, etc.).
- This means that (1) similar sediment types can occur in basins of completely different tectonic origin, and (2) an additional basin classification is needed which is based on the various depositional environments.
- Certain basin fills reflect pre- and syndepositional tectonics. To understand both basin-forming tectonics and depositional environments, basin fills should be studied at different scales (outcrops, drillholes, geometry of the total basin fill).