Basin Analysis

Sedimentary basins: 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.

The relative motion of plates produces deformation, volcanicity, and seismicity concentrated along their boundaries, which are classified as divergent boundaries, such as the mid-ocean ridge spreading centers 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 centers.

• 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:

(1) Basins due to lithospheric stretching, belonging to the rift-drift suite

(2) Basins formed primarily by flexure of continental and oceanic lithosphere.

PLATE MOTION:

Plate tectonics can operate because the lithosphere is composed of a number of coherent "**plates**" (Fig. 1).

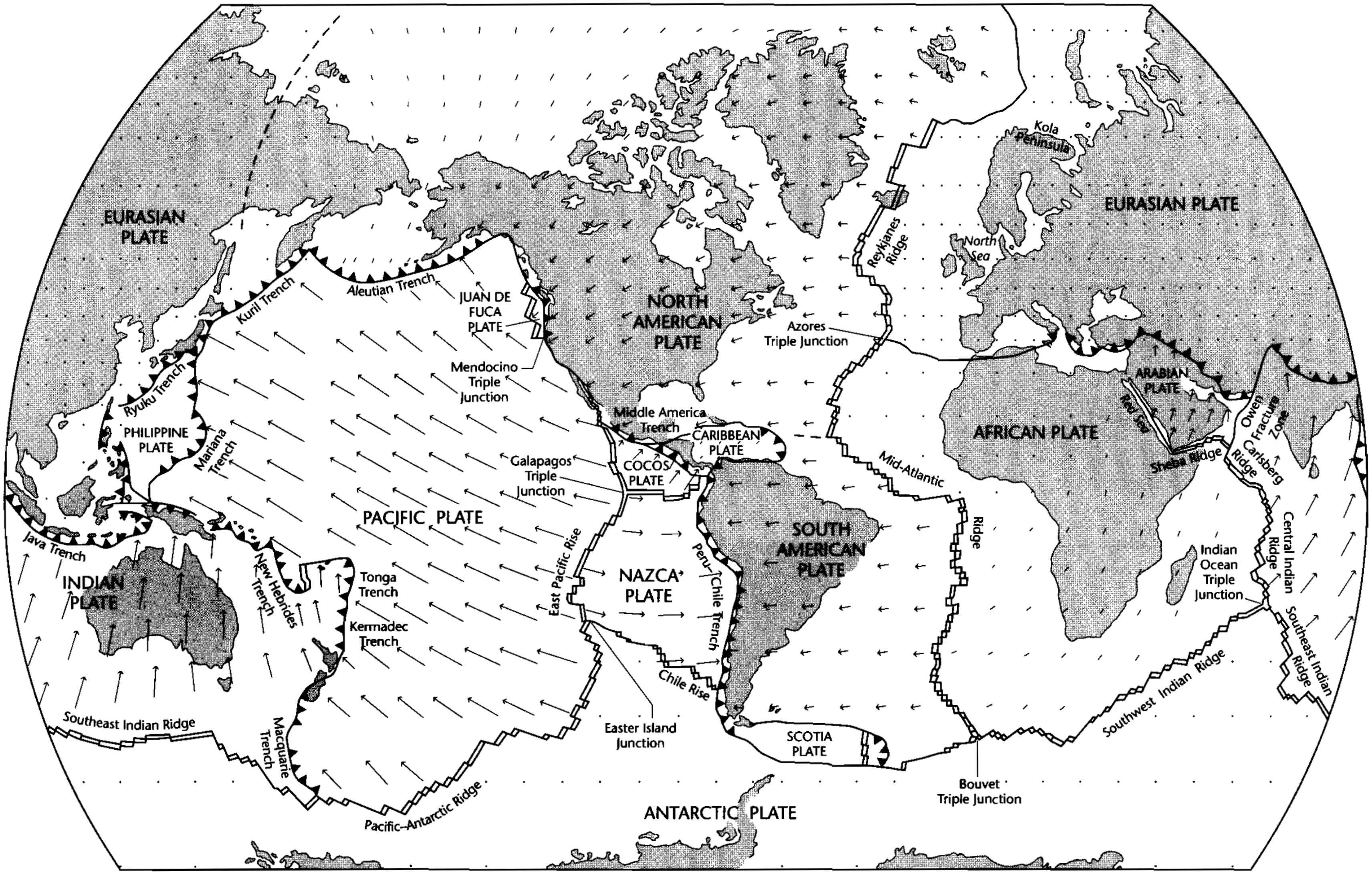

Fig. 1. The lithospheric plates, showing mid-ocean ridges, trenches, and transform boundaries (Le Pichon et al. 1973) and absolute motion vectors from Minster and Jordan (1978). Length of arrows is proportional to the plate speed. The fastest plate motion is in the western Pacific and Indian Oceans, whereas Africa, Antarctica, and Eurasia are almost stationary with respect to the mantle reference frame. Reproduced courtesy of American Geophysical Union.

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.

Three classes of plate boundary exist: ***divergent, convergent,*** and ***conservative*** (Fig. 2).

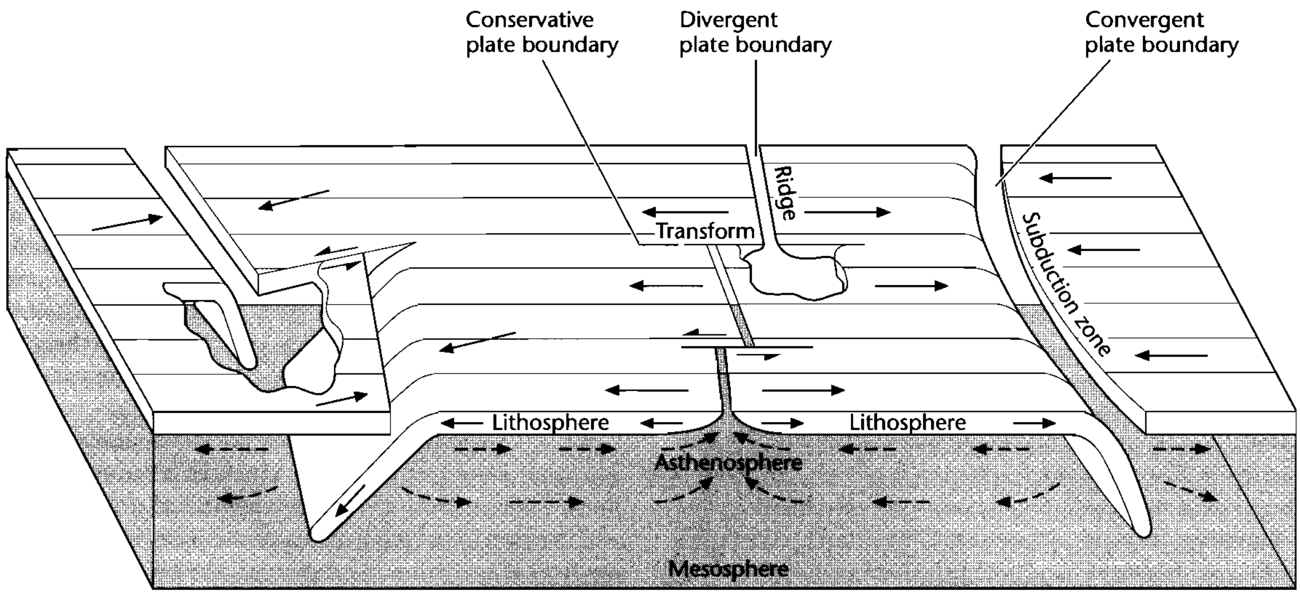

Fig. 2 The three types of plate boundary: convergent, divergent, and conservative (after Kearey and Vine 1996). Reproduced courtesy of Blackwell Publishing Ltd.

Divergent boundaries are typified by the mid-ocean ridge spreading centers of the ocean basins. Here, the recognition of magnetic bands correlated with a magnetic reversal 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:

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

2- 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.

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 (Fig. 3).
Many sedimentary basins can be fitted into a particular phase of the Wilson cycle.

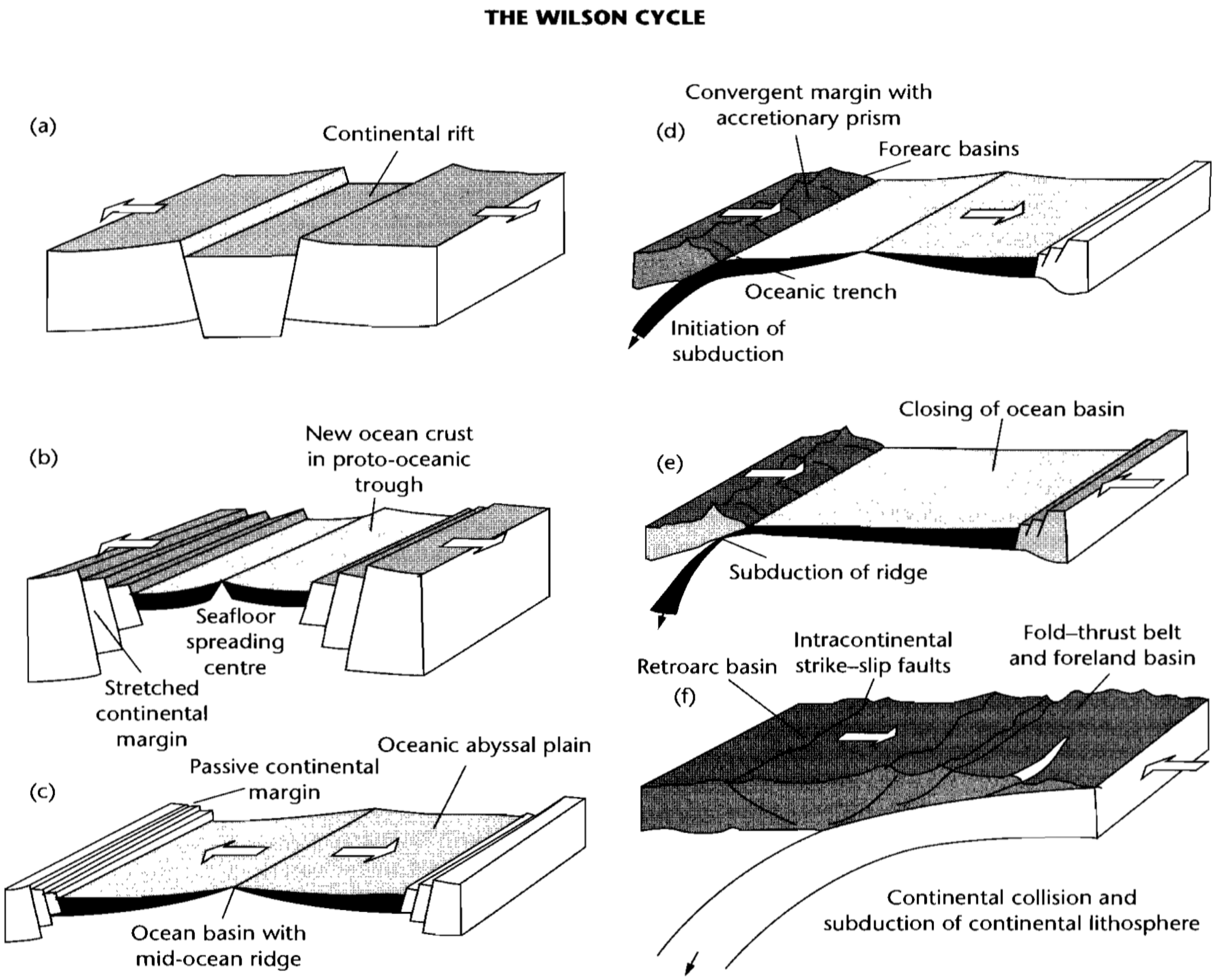

Fig. 3 The Wilson cycle of ocean formation and ocean closure. Continental extension (a) is followed by the creation of a new oceanic spreading centre (b) and ocean enlargement (c). Subduction of ocean floor (d) leads to closure of the ocean basin. Subduction of the oceanic ridge (e) takes place before continent–continent collision (f).

Classification schemes of Sedimentary Basins:

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. 4). The evolution of a basin could then be 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.

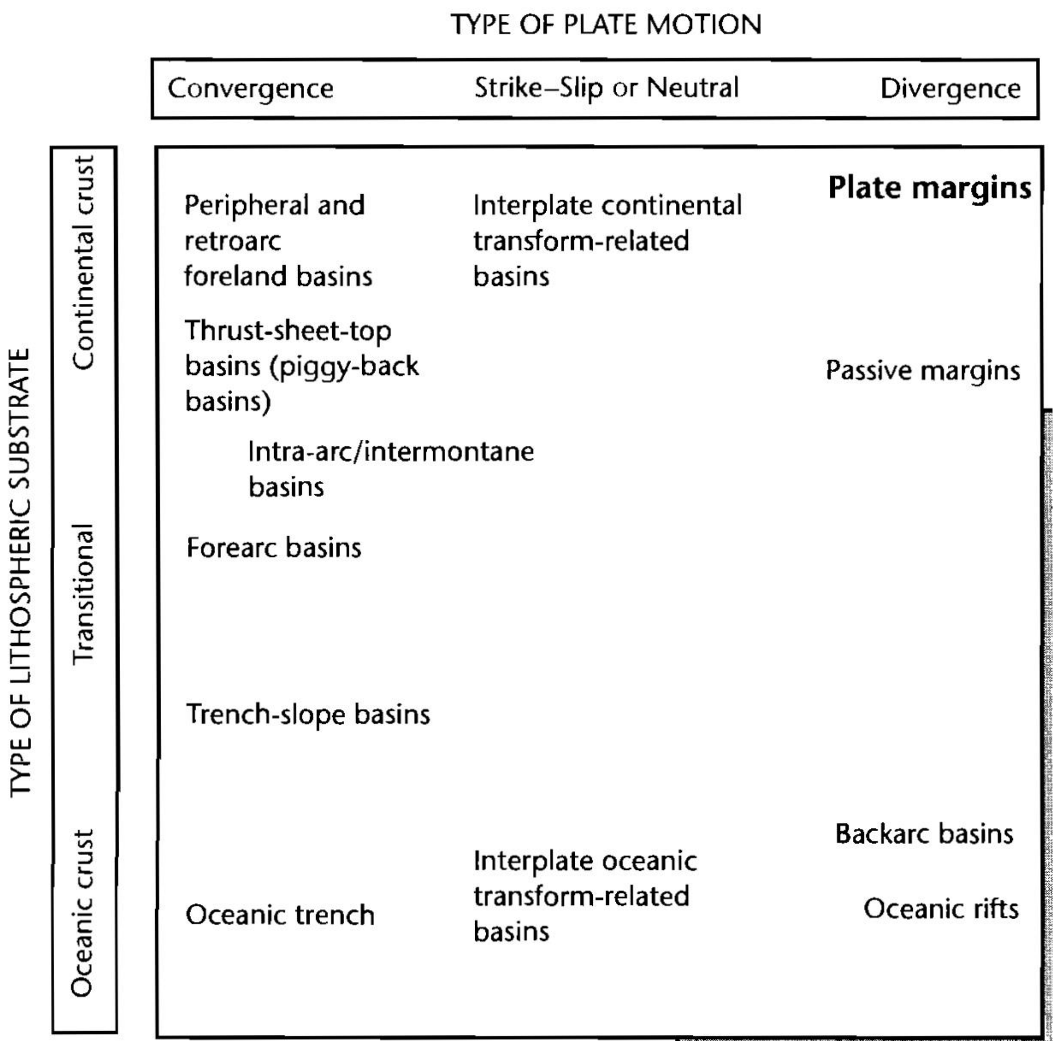

Fig. 4 Classification of basins using the type of lithospheric substrate, type of plate motion, and location with respect to the plate boundary.

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. 5) 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** these are:

1- The basin-forming tectonics.
2- The depositional sequences filling the basin.
3- The basin-modifying tectonics.

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 which can be summarized as:

1- Crustal thinning such as caused primarily by stretching or surface erosion.
2- Lithospheric thickening, such as caused by cooling following stretching or accretion of melts derived from the asthenosphere.
3- Sedimentary and volcanic loading causing isostatic compensation.
4- Tectonic (supracrustal) loading causing isostatic compensation.
5- Subcrustal loading caused by subcrustal dense loads such as obducted mantle flakes or crustal densification due to phase changes.
6- Asthenospheric flow primarily due to the subduction of cold lithospheric slabs.

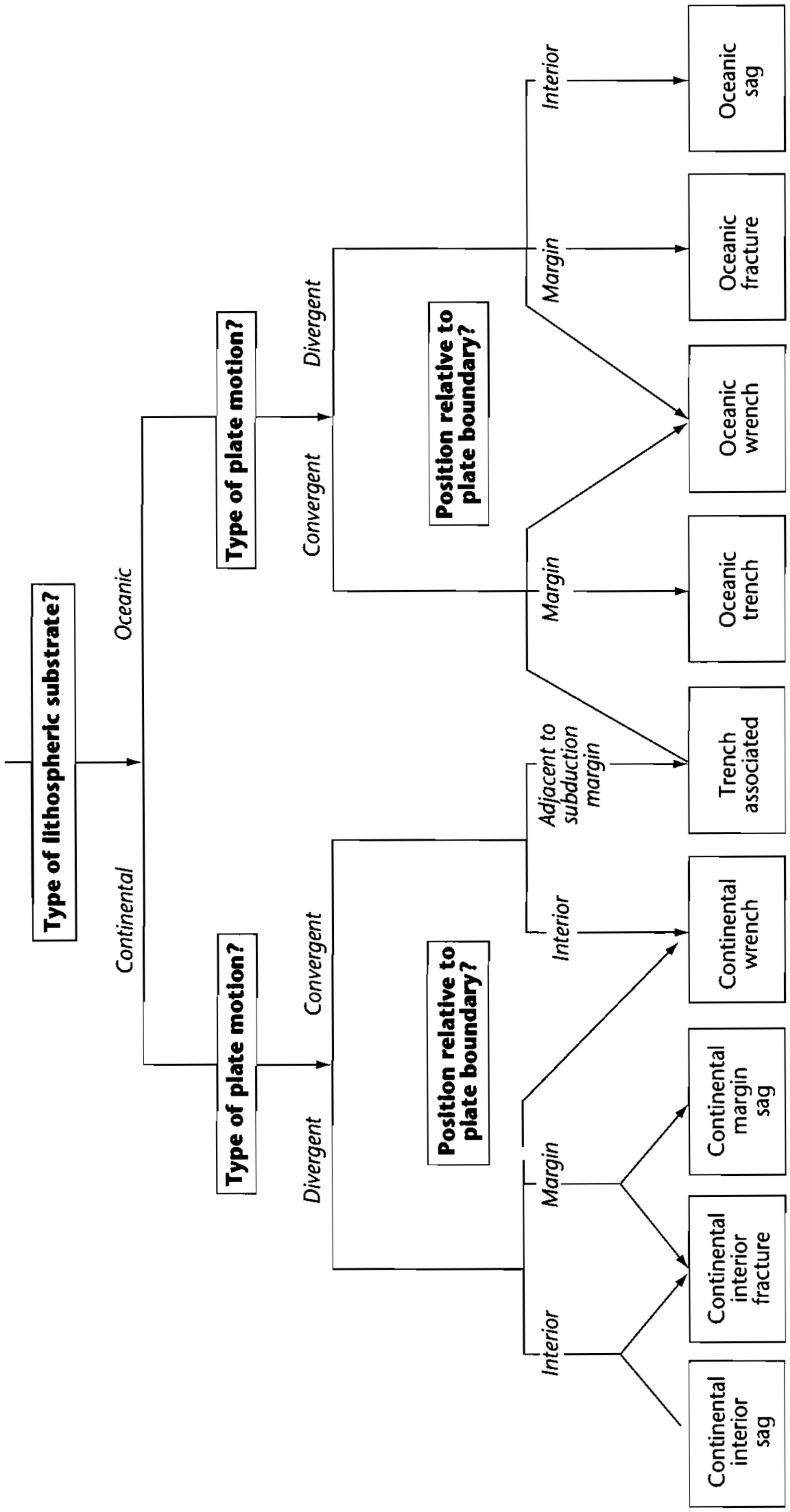

Fig. 5 Basin classification scheme based on Kingston et al. (1983a, b). Not all basin types appear in this scheme. Most notably, foreland basins are missing.

Basins due to flexure:

Flexure is the long wavelength deflection of a lithosphere of finite strength caused by the application of an external force system. A general flexural equation can be established for the case of a thin elastic plate overlying a weak fluid subjected to vertical applied forces, horizontal forces, and torques or bending moments. The general flexural equation can then be used in different geodynamic situations by applying different boundary conditions.

Flexure of the lithosphere is most clearly demonstrated at oceanic islands, seamount chains, ocean trenches, and foreland basins. Lithospheric flexure also supports sediment loads in most sedimentary basins. The flexural basins associated with ocean-continent and continent-continent plate margins are particularly well represented in the geological record. In an Andean-type setting the flexural basin on the subducting oceanic plate is the ocean trench and the flexural basin on the upper plate is a retroarc foreland basin. In Alpine-type settings the flexural basin on the subducting plate or lower plate is called a peripheral- or pro-foreland basin, whereas the flexural basin on the overriding or indenting plate is termed a retro-foreland basin. Some foreland basins are related to subduction zone roll-back (subduction zone retreat) and are associated with prominent backarc extension. The formation and evolution of foreland basins is intimately related to the processes of shortening, exhumation, and extensional collapse in the adjacent orogenic wedge.

The deflection of the oceanic lithosphere along seamount chains such as the Hawaiian Islands can be explained by either the flexure of a continuous plate loaded by a vertical applied force (representing the excess mass of the seamount chain), or by the flexure of a plate broken beneath the vertical applied force.

Foreland basins are elongate or arcuate, highly asymmetrical basins closely associated with continental collision zones. The alternative term "foredeep" was introduced by Price (1973). Dickinson (1974) proposed two genetic classes of foreland basin (Fig. 6):

1- Peripheral foreland basins situated against the outer arc of the orogen during continent-continent collision (e.g., Indo-Gangetic Plain, north Alpine foreland basin).

2- Retroarc foreland basins situated behind a magmatic arc linked with subduction of oceanic lithosphere (e.g., Andean examples, Late Mesozoic-Cenozoic Rocky Mountain Basins, North America). Both classes of fore-land basin overlie cratonic lithosphere and are associated with crustal shortening in tectonically active zones. Some highly arcuate thrust belts and spatially restricted foreland basins are related to subduction zone roll-back and are commonly associated with important backarc extension (Fig. 6). The Apennines-Tyrrhenian Sea of the western Mediterranean is an example.

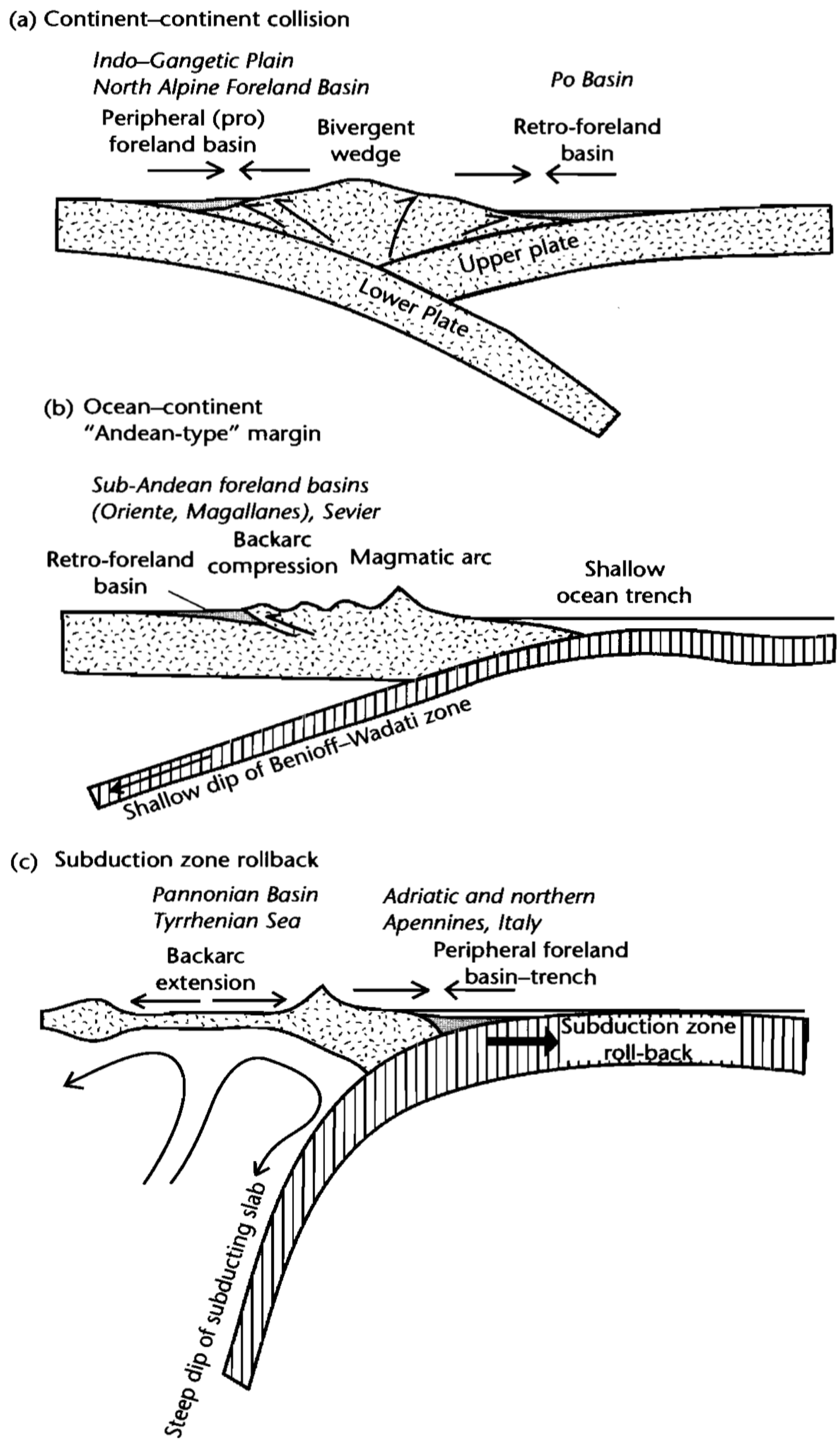

Fig. 6 Schematic illustration of peripheral foreland basins, retro-foreland basins, and basins related to subduction zone roll-back.

Basins associated with strike slip deformation

Sedimentary basins generally form by localized extension along a strike-slip fault system which itself may be related to either divergent, convergent, or oblique relative plate motion. Less commonly, loading resulting from local crustal thickening may cause flexural subsidence.

Although strike-slip basins form in a wide variety of geodynamical settings such as oceanic and continental transforms and arc and suture collisional boundaries, they are best known from intracontinental and continental margin environments.

Zones of strike-slip tectonics are characterized by active seismicity on strike-parallel and strongly oblique faults, with zones of infrequent, large earthquakes along locked segments, and frequent small earthquakes along unlocked segments. Some of the world's best known and most hazardous faults are strike-slip.

Heat flows are generally low, suggesting that major strike-slip faults are weak and therefore generate little frictional heat.

Strike-slip zones may be associated with entire plate boundaries such as the San Andreas Fault system of California and the Alpine Fault system of New Zealand, microplate boundaries, intraplate deformations, or small fractures of limited displacement.

Sylvester (1988), drawing considerably on Woodcock's (1986) genetic scheme (Fig. 7), proposed a classification of strike-slip faults into interplate and intraplate varieties. He recommends use of the term "transform" fault for deep-seated interplate types and "transcurrent" fault for intraplate strike-slip faults confined to the crust.

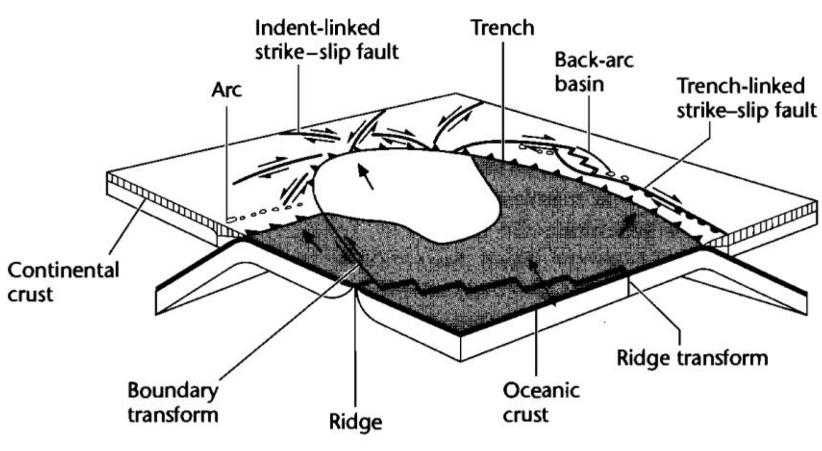

Fig. 7 Genetic classification of major classes of strike–slip fault according to plate tectonic setting (after Woodcock 1986). See also Table 6.1.

The sedimentary basin-fill

The sediment routing system is the integrated geomorphic system involving sediment liberation, transport, and deposition. Understanding this complex system in terms of its physical laws and response times is important for the evaluation of sediment supply to sedimentary basins, which controls the stratigraphic architecture of the basin-fill.

WEATHERING

Weathering is the decay and disintegration of rock *in situ* at the Earth's surface. A weathering mantle forms between pristine bedrock and the land surface, known as regolith. Water moves up and down in the regolith by capillary action and gravity respectively (Fig. 8).

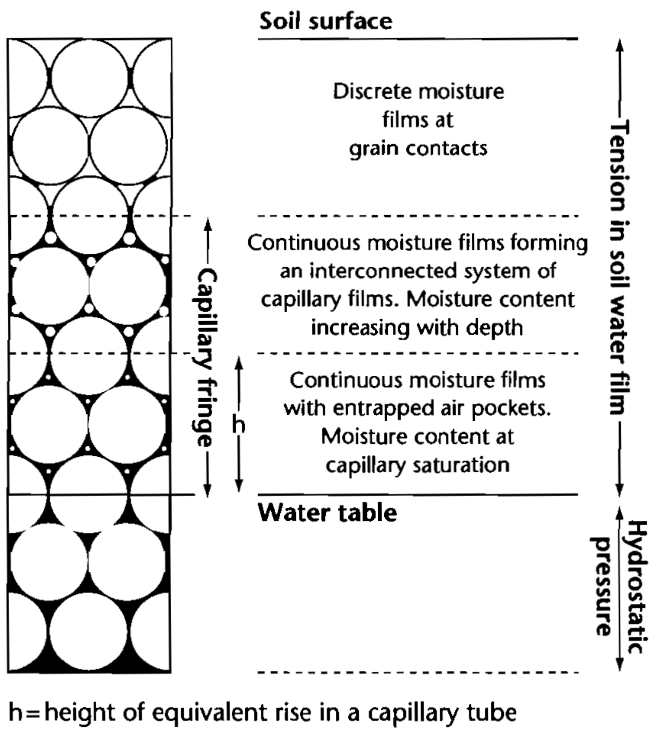

Fig. 8 The zones of moisture in the regolith, showing the extent of the capillary fringe (after Carson 1969).

Chemical weathering processes:

Chemical weathering involves the chemical breakdown of bedrock and the formation of new mineral products.
The main chemical weathering processes are:

1- **Solution** involves the action of water as a solvent. The tendency of a mineral to dissolve in water is expressed by its equilibrium solubility. This is affected by the temperature and pH of the local environment. As an example, quartz (SiO_2) has a low solubility below a pH of 10, but is highly soluble in very alkaline waters above this value. Alumina ($A1_2O_3$) is only soluble in conditions seldom found in nature, below a pH of 4 and above a pH of 9. As a result, alumina accumulates as a residue during weathering, whereas silica may be slowly leached. Calcium carbonate ($CaCO_3$), in contrast, has a steadily decreasing solubility in alkaline waters. However, the low solubility of $CaCO_3$ in pure water is rarely applicable in the natural environment because dissolved CO_2 in water causes $CaCO_3$ to be replaced by calcium bicarbonate $Ca(HCO_3)_2$, which is highly soluble.

2- **Oxidation and reduction** involves the gain or loss of charge by the addition (reduction) or loss (oxidation) of negatively charged electrons. The oxygen dissolved in water is the most common oxidizing agent. Oxidation results in the formation of oxides and hydroxides, as in the oxidation of sulfides such as iron pyrite (FeS_2) under anaerobic conditions to produce sulfuric acid and iron hydroxide. The oxidation of organic matter in soils by bacteria produces CO_2 and therefore generates acidity. The acidity is then used in the hydrolysis of minerals. The tendency for oxidation and reduction to take place is indicated by the **redox potential** (Eh), measured in millivolts.

3- **Hydration** involves the absorption of water into the crystal lattice, making it more porous and therefore more susceptible to weathering.
A common example is the transformation of the iron oxide hematite to the hydrated iron hydroxide limonite.

4- **Acid hydrolysis** is the reaction of a mineral with acidic weathering agents, where the acidity is mainly derived from the dissociation of atmospheric CO_2 in rainwater and soil zones by respiration of plant roots and bacterial decomposition of plants, producing in both cases carbonic acid (H_2CO_3). Hydrolysis involves the replacement of metal cations in the crystal lattice such as K^+, Na+, Ca^{2+}, and Mg^{2+} by the hydrogen or hydroxyl ions of water. The released cations combine with further hydroxyl ions, commonly to form clay minerals. Examples are the hydrolysis of albite (plagioclase felspar $NaAlSi_3O_8$) to kaolinite $(AI_4Si_4)O_{10}(OH)_8$.

The regolith (the layer of unconsolidated rocky material covering bed rock)

The mineralogic composition of the regolith is determined not just by the type and intensity of chemical weathering processes, but also by the parent bedrock.

There are considerable differences in the way basalts and granites weather under the same climatic regime. Clay minerals are particularly diagnostic of both weathering processes and parent material. A primary factor is the extent of leaching, which strips minerals of their metal cations and eventually of their silicon and iron, leading to a stable aluminum-rich residue (gibbsite).

A large throughput of water is necessary for advanced stages of leaching, leading to regoliths dominated by gibbsite, kaolinite, and aluminum oxides and hydroxides.

This is only accomplished in regions of high precipitation rates such as the equatorial and humid tropical regions. The process is commonly termed **laterization**. Where leaching is less intense, the cations released by the breakdown of bedrock promote the formation of cation-bearing clay minerals such as those of the illite and smectite groups. These zones of moderate leaching typify the temperate zones of Asia, Europe, and North America. In semi-arid climates, there are very low rates of chemical weathering and little accumulation of weathering products except for carbonates and salts as hard crusts and concretions.

The clay mineral assemblage is not only a function of climate, but also of position within the regolith profile. This is because the flux of water generally decreases with depth in the regolith, caused by a downward reduction in permeability. Residual minerals such as kaolinite and gibbsite are consequently found at the top of weathering profiles, whereas smectite and illite may be found at deeper levels in the same regolith where leaching is less intense.

The thickness of the regolith depends on the trade-off of two rate effects:

1- Rate of bedrock weathering, which is strongly controlled by climatic factors.

2- Rate of removal by denudation, which is controlled by climatic, tectonic, and topographic factors.

There are a number of factors controlling the rate of chemical weathering:

- **Organic activity** in soils generates soil acidity (CO_2) through decomposition, aids the retention of water through the build-up of organic matter and biologic activity promotes permeability.
- **Climate** controls weathering reactions through the effect on chemical kinetics of temperature, imparting a latitudinal gross pattern of chemical weathering rates.
- **Kinetics** of mineral reactions; chemical weathering requires pore waters to be undersaturated with respect to the mineral being weathered.
- **Bedrock** composition controls the stability of the mineral components to weathering through their degree of polymerization, as illustrated by the Goldich series. Monomer silicates like olivine are most easily weathered, and framework silicates such as quartz are most resistant;
- **Topography** influences the rate of removal of regolith by erosion, which invigorates chemical weathering by subjecting new, fresh bedrock to weathering.
- **Time** is required for chemical changes to take place.

Basin Stratigraphy:

The stratigraphy in a sedimentary basin is the result of the interplay of the generation of space or accommodation and the influx of sediment. Stratigraphic geometries and gross depositional environments are therefore determined by the tectonic mechanisms causing subsidence, local patterns of faulting, the nature of sediment routing systems, and sea-level change.

Relative sea-level change and accommodation: definitions

Eustasy is global sea level measured from the sea surface to a fixed datum, such as the center of the Earth. **Relative sea level** is sea level measured relative to a moving datum, often a distinctive horizon such as a horizon within a sediment pile, or the lower contact with basement. Relative sea level is therefore affected by processes such as tectonic uplift and subsidence, compaction and eustasy.

Water depth is of course the vertical distance between the sea surface and the seabed. Although water depth commonly changes during a relative sea-level change, it may also be influenced by the sediment input into a sedimentary basin with no relative sea-level change.

Eustasy, relative sea level, and water depth are all distinctive concepts (Fig. 9).

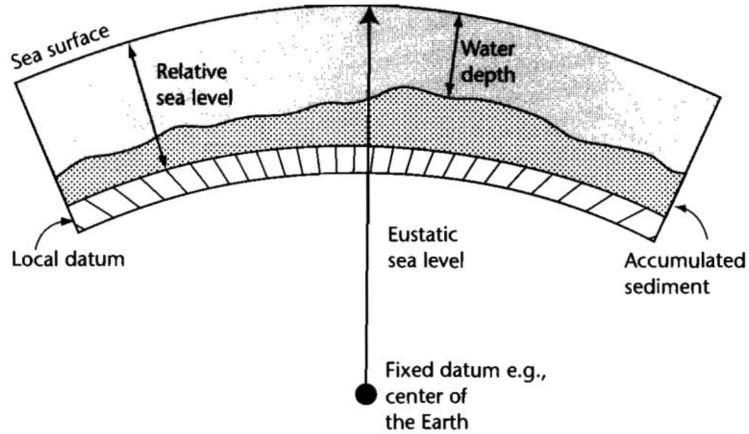

Fig. 9 Definitions of terms used in process stratigraphy (after Jervey 1988; Emery and Meyers 1996): eustatic sea level, relative sea level, and water depth.

Accommodation (the space made available for sediment to accumulate) (Fig. 10) is controlled by base level, since sediment can only accumulate long term up to base level. Base level may be a graded stream profile on land, or a graded shelf profile on the continental shelf.

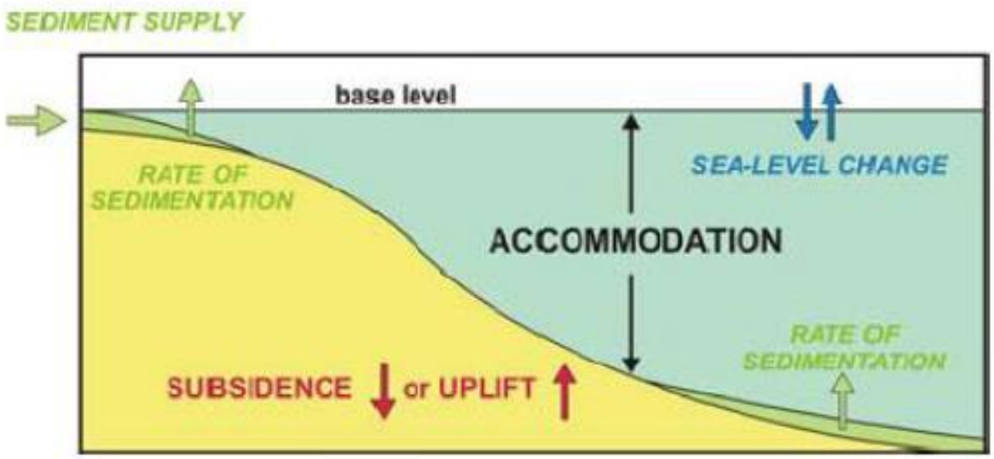

Figure (10) The accommodation and factors that effect on it

STRATIGRAPHIC CYCLES:

megasequences

megasequences representing the stratigraphic packages deposited during distinct phases of plate motion. Consequently, a megasequence may correspond to a period of continental extension with synrift and postrift components, or to a period of convergent plate motion characterized by flexure of the continental lithosphere.

Megasequences are also bounded by extensive but not global major unconformities. Since sedimentary basins are deformations of the lithosphere under a set of platescale driving forces, the concept of the megasequence is of primary importance in basin analysis.

Depositional sequences

The primary meso-scale units of stratigraphy are termed depositional sequences. They are coherent packets of strata that are genetically related and which can be traced for considerable distances across a basin.

Systems tracts

Depositional sequences can be subdivided into smaller units of stratigraphy that have distinct stacking patterns of chronostratigraphic increments. These smaller units are termed systems tracts and they are themselves composed of parasequences, or Paracycles. The tracts of depositional systems have been related to specific intervals of cycles of relative sea level.

Although systems tracts were defined as the stratigraphic response of a system to the competing effects of rate of relative sea-level change and rate of tectonic subsidence, they are clearly sensitive to variations in sediment supply.

Systems tracts are broadly divided into the following classes according to their relationship to specific segments of the relative sea-level change curve (Fig. 11)

1- Lowstand systems tract When relative sea-level fall is rapid, no space is available for further sedimentation, the former shelf is incised by streams (producing incised valley systems) and sedimentation is transferred to the basin floor and slope. Base-of-slope fans (lowstand fans) are nourished by sediment bypassed through the shelf and slope by valleys and canyons. Slope fans result from deposition on the middle or base of the continental slope and may be coeval with the basin-floor fan.

2- Transgressive systems tract. During a rapid relative sea-level rise, the underlying lowstand or shelf margin system tracts are transgressed (the transgressive surface). Where the transgressive surface is erosional, it is called a ravinement surface. Sets of parasequences making up the transgressive system tract are commonly retrogradational (Fig. 12 c), that is, they back-step onto the basin margin, with strong onlap in a landward direction and downlap onto the transgressive surface in a basinward direction. As the rate of relative sea-level change slows down, the sets of parasequences change from being retrogradational to being aggradational (fig 12 b), the surface at which this occurs being that of the maximum flooding. Condensed sections occurin the basin during times of transgression.

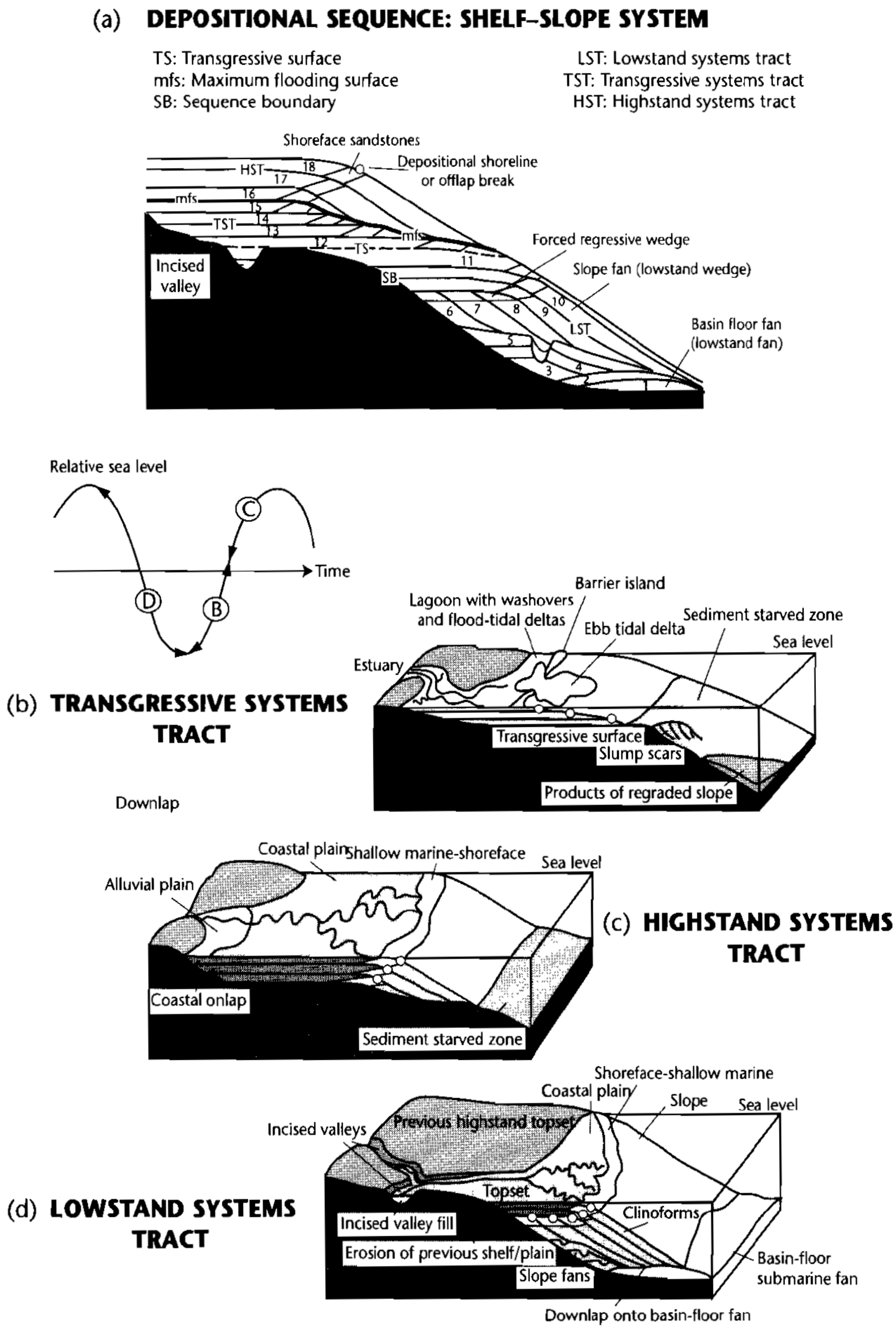

Fig. 11 Characteristic systems tracts and their relation to the relative sea-level curve, according to the Exxon group (Posamentier et al. 1988), modified by Emery and Myers (1996). These block diagrams are extremely idealized and have a strong vertical exaggeration. (a) Arrangement of systems tracts within a depositional sequence in a coastal plain–continental shelf-slope system. Numbers are relative ages of chrons; (b), (c) and (d) show transgressive, highstand, and lowstand systems tracts respectively. Open circles are depositional shoreline or offlap break. Reproduced courtesy of Blackwell Publishing Ltd.

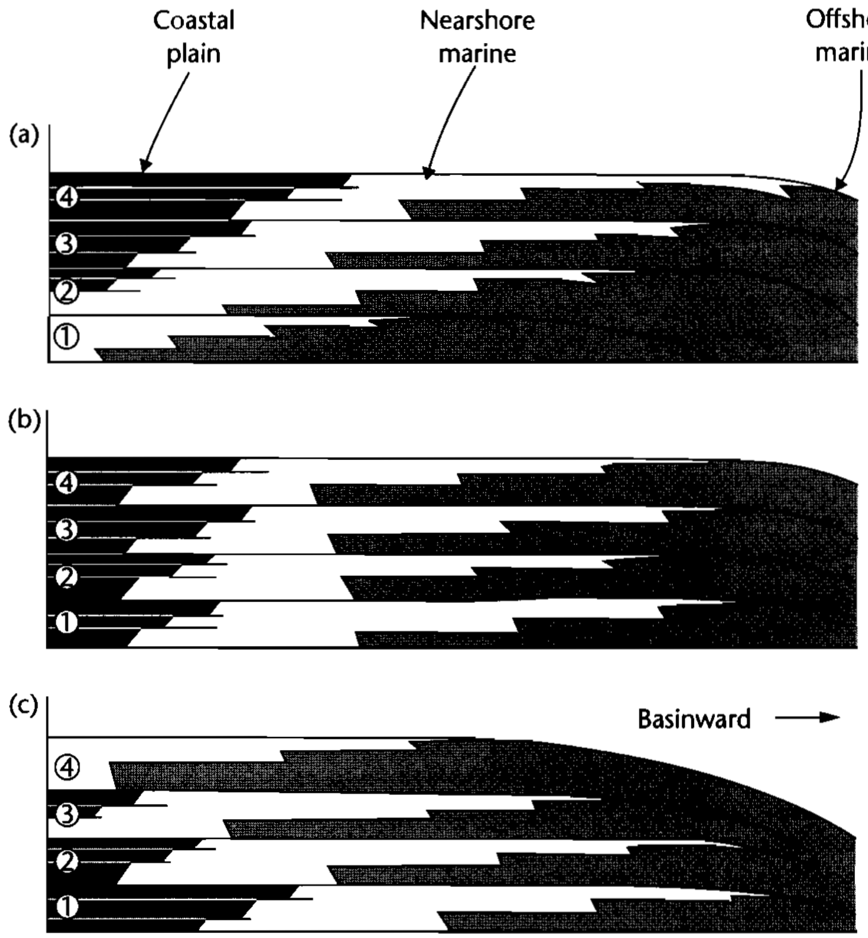

Fig. 12 Stacking patterns of parasequence sets (after Van Wagoner et al. 1988) are indicative of the trend in time of sediment supply and relative sea-level change or accommodation. (a) Progradational parasequence set with a basinward migration of the shoreline, characteristic of the highstand systems tract and lowstand prograding wedge; (b) Aggradational parasequence set showing no movement of the shoreline, characteristic of the shelf-margin systems tract; (c) Retrogradational parasequence set, with a landwards migration of the shoreline, characteristic of the transgressive systems tract. Reproduced courtesy of Society for Sedimentary Geology.

3- Highstand systems tract: After the maximum flooding, the relative sea-level rise slows and sets of aggradational parasequences are succeeded by progradational parasequences with clinoform geometries (Fig. 12).

Basins related to convergent plate motion

1- Morphological and tectonic elements at arc-related margins:

The main components of convergent arc-related systems are, from overriden oceanic plate to overriding plate (Fig. 13).

• An outer rize on the oceanic plate recognized as an arch in the abyssal plain. This is the flexural forebulge of the descending oceanic plate.

• A trench or deep trough, commonly >10 km deep situated oceanward of the arc. The sediments of trenches are dominated by fine-grained turbidites and pelagic deposits.

The bathymetric expression of the trench much depends on the sediment supply into it and, associated with this, the rate of encroachment from the arc of the accretionary wedge. The trench is the bathymetric expression of the deflected (flexed) oceanic plate.

• A subduction complex composed of tectonic stacks of fragments of oceanic crust, its pelagic cover and arc-derived turbiditic sediments, together with perched or accretionary basins ponded on top of the accretionary wedge. The subduction complex makes up the inner slope of the trench. Where accretion rates are high, the subduction complex may rise to shelf depths or even become emergent;

• A forearc basin between the ridge or terrace formed by the subduction complex and the volcanic arc.

• The magmatic arc caused by partial melting of the overriding plate and possibly subducted plate when the latter reaches between 100 and 150km depth. The volcanism is predominantly andesitic. Small intra-arc basins may form by extensional or strike-slip tectonics or in collapsed calderas.

• The backarc region floored by oceanic or continental lithosphere. Where the lithosphere is oceanic, the backarc region typically undergoes extension. Backarc basins are some of the most rapidly extending regions of the Earth's crust today, a prime example being the Aegean Sea of the eastern Mediterranean. Where the lithosphere is continental, as in Andean-type margins, the backarc (or retroarc) region is typically a zone of flexural subsidence related to major fold-thrust tectonics along the arc boundary.

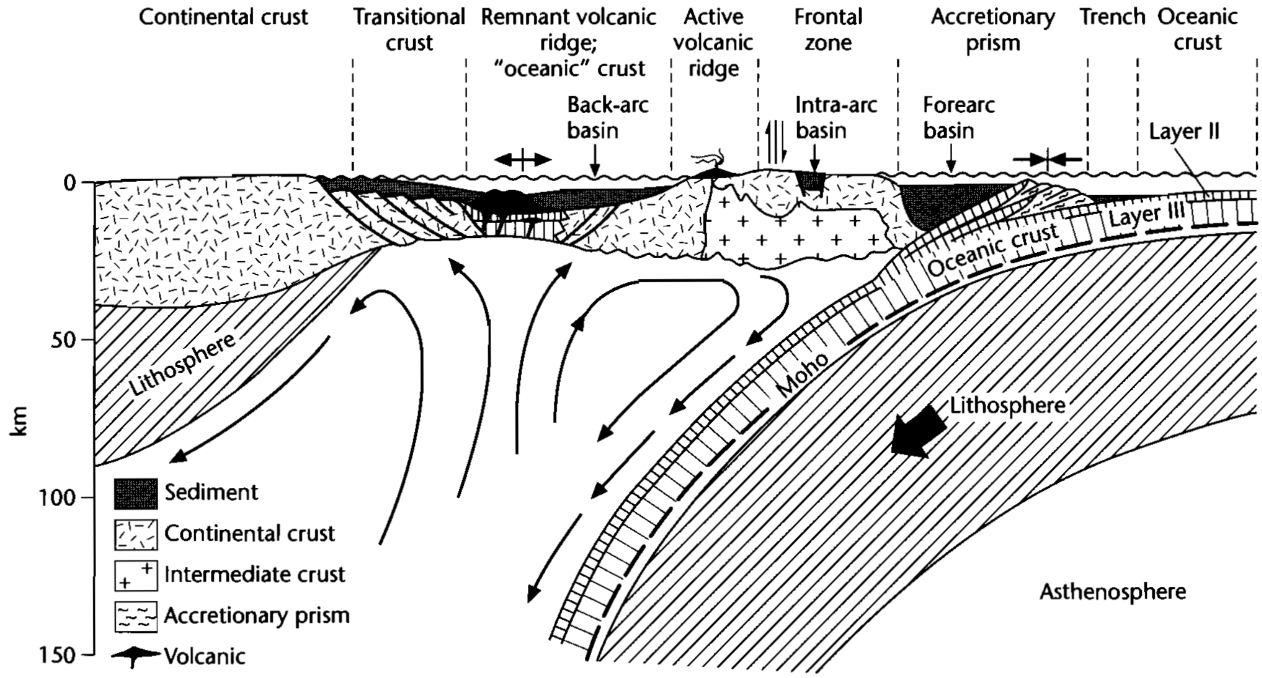

Fig. 13 A convergent ocean–arc boundary showing the location of the trench, accretionary wedge, forearc basin, intra-arc and backarc basins. Modified from Dickinson and Seeley (1979). The intra-arc or intra-massif basin and the forearc basin both contain deep marine to nonmarine sediments. The accretionary basin on the subduction complex contains tectonic slices of abyssal plain, slope and trench deposits together with ophiolites and metamorphics.

2- Foreland basins (peripheral and retroarc types)

The foreland basin system contains four depositional zones (Fig. 14):

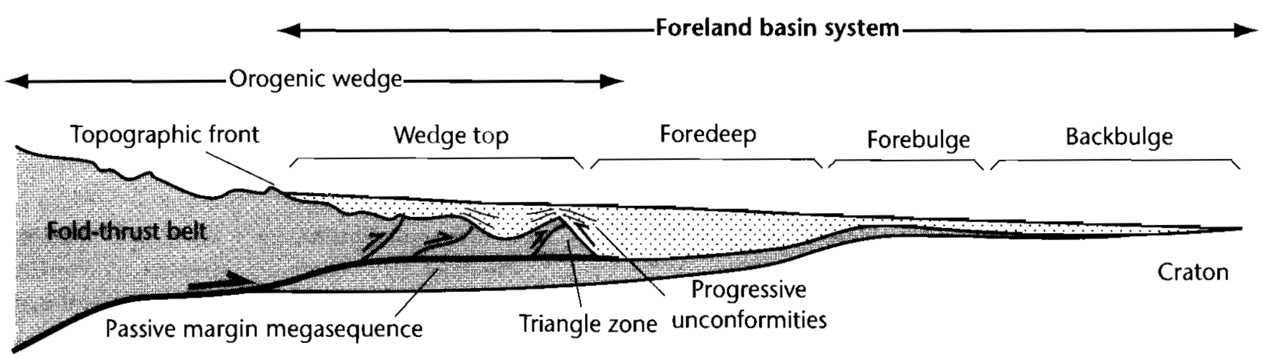

Fig. 14 The four depositional zones of a foreland basin system as envisaged by DeCelles and Giles (1996).

1- Basins that rest on moving thrust sheets as a thrust-sheet top or piggy-back basin, which receives sediment from the eroding orogenic wedge.

2- A basin ahead of the active thrust system in a foredeep, which is supplied with sediment from both the continental foreland and the orogenic wedge.
3- Sediment may also accumulate on the flexural forebulge if ccommodation is available, for example because the foreland lithosphere is submerged below sea level as a result of negative dynamic topography.
4- A shallow, broad backbulge basin filled with shallow marine and continental sediments; ongoing convergence should cause the backbulge depozone to be uplifted and eroded in the flexural forebulge.

3- Ocean trenches and accretionary basins

Ocean trenches are one of a series of sedimentary basin types found at convergent arc-related or ocean-continent boundaries. Here we draw the distinction between:
(i) Trenches situated on the downbent oceanic lithosphere.
(ii) Accretionary basins perched on the accretionary subduction complex.
(iii) Forearc basins located between the arc and the subduction complex, and
(iv) Backarc basins found on the landward side of the arc (Fig. 13).

4- Forearc and backarc basins

Three types of sedimentary basin occur in the Forearc region:
(i) The accretionary basins located on subduction complexes mentioned in the previous section.
(ii) Intra-arc or intra-massif extensional basins, common where a broad forearc region has developed over continental crust, as in the Andes. These elongate basins typically follow volcanic lines or fundamental tectonic lineaments and are filled mainly with nonmarine fluviatile and lacustrine sediments dominated by volcanic constituents.
(iii) Large basins located between the subduction complex and the magmatic arc, comprising "residual" basins with a basement of stretched continental crust or obducted oceanic crust, and "constructed" basins underlain by the landward portion of the subduction complex.
The oldest sediments of "residual" forearc basins are generally deep-water pelagic deposits, whereas those of "constructed" types may be shallower. Submarine fans build into the basins transversely from the magmatic arc. Intra-oceanic forearcs tend to be sediment-starved and remain deep marine, whereas sediment-nourished examples near major continental catchments may rapidly become shallow marine.

Depositional environments in backarc basins on oceanic crust are also deep marine, except along their margins where fluviatile, coastal, and shallow marine depositional environments may exist. Karig and Moore (1975) pre sented a model for the evolution of backarc basins based on the western Pacific examples. Initially, volcaniclastic wedges shed from the arc interfinger with a background of pelagic clays. As subsidence continues and outstrips sediment supply, the seafloor commonly descends below the carbonate compensation depth (CCD), so that the more evolved basin accumulates siliceous rather than calcareous clays.

Strike-slip basins:

Basin geometries are deep but relatively narrow, with high syndepositional relief causing conglomerates and breccias to be banked up against faulted basin margins. Sedimentation rates are rapid. Lateral facies changes are also rapid, so that marginal breccias may pass laterally directly into lacustrine mudstones. Fault movements cause syndepositional unconformities to form in individual basins and different stratigraphies to develop in closely adjacent basins, making correlation difficult.
The best known intracontinental transform is the San Andreas system, and one of the best documented pull-apart basins in this system is Ridge Basin, California.

The basin-modifying tectonics

The third major element used to classify basins is basin-modifying tectonics. Basins or cycles formed by one type of tectonic movement may be changed during their history by other structural events. There are three types of basin modifying tectonics:
1- Episodic wrenches (L).
2- Adjacent foldbelts (FB).
3- Complete folding of a basin area (FB3) which is foldbelt formation.

Definition of Episodic Wrenches and Foldbelts

Episodic wrenches are designated by the single letter "L," and represent a wide variety of lateral movements not connected with basin/cycle origin. Episodic wrenches modify basins formed by other means and are found in basins with all ages of basement rocks. It is believed that old zones of weakness in the basement, such as old sutures, interior fracture zones, plate boundaries, etc, move periodically or episodically in response to plate movements.

Generally, the origin of an episodic wrenching or lateral (L) movement is fairly easy to ascertain, given good plate-tectonic reconstructions. In some places, the effect of varying intensities of (L) movements is shown in (Fig. 15).

Foldbelts are caused by convergence of two or more plates. Basin areas caught in this convergence may be completely or only partly folded. Basins not completely folded are not considered to be foldbelts and are said to have been episodically wrenched. Basins completely folded are called foldbelts (FB3).

Tectonic Modifiers

Wrenches (L) and Adjacent Foddbelts (FB)

The tectonic modifiers of primary basin types are listed in (Fig 15) in order of increasing magnitude, downward from "a" to "f" Each of these effects is found associated with both episodic wrenches (L) and adjacent foldbelts (FB). The very weak "a" effect is known to occur within or adjacent to a basin with minimal structural effect. La would mean that a wrench passed through or adjacent to a basin but caused no faulting or folding visible at the surface or on seismic reflection. Porosity and permeability, however, may be affected. FBa would mean that a foldbelt was formed on the side of a basin but had no effect, of faulting or folding, on the basin itself.

Lb and FBb illustrate the "b" effect-still very weak on the scale. The "b" effect triggers salt or shale diapirs and growth faults within the basin and can cause open folds in basins adjacent to FBb. We believe that without a tectonic event of "b" intensity or stronger, salt and shale may not be triggered to flow. The movement could be described best as "jiggling." There are numerous examples, world-wide, of basins with thick salt layers and plenty of load that have never flowed to produce. Nor do they exhibit any other evidence of post-salt deposition structuring. It can be concluded that slight plate motions or jiggling are required to initiate or trigger salt and

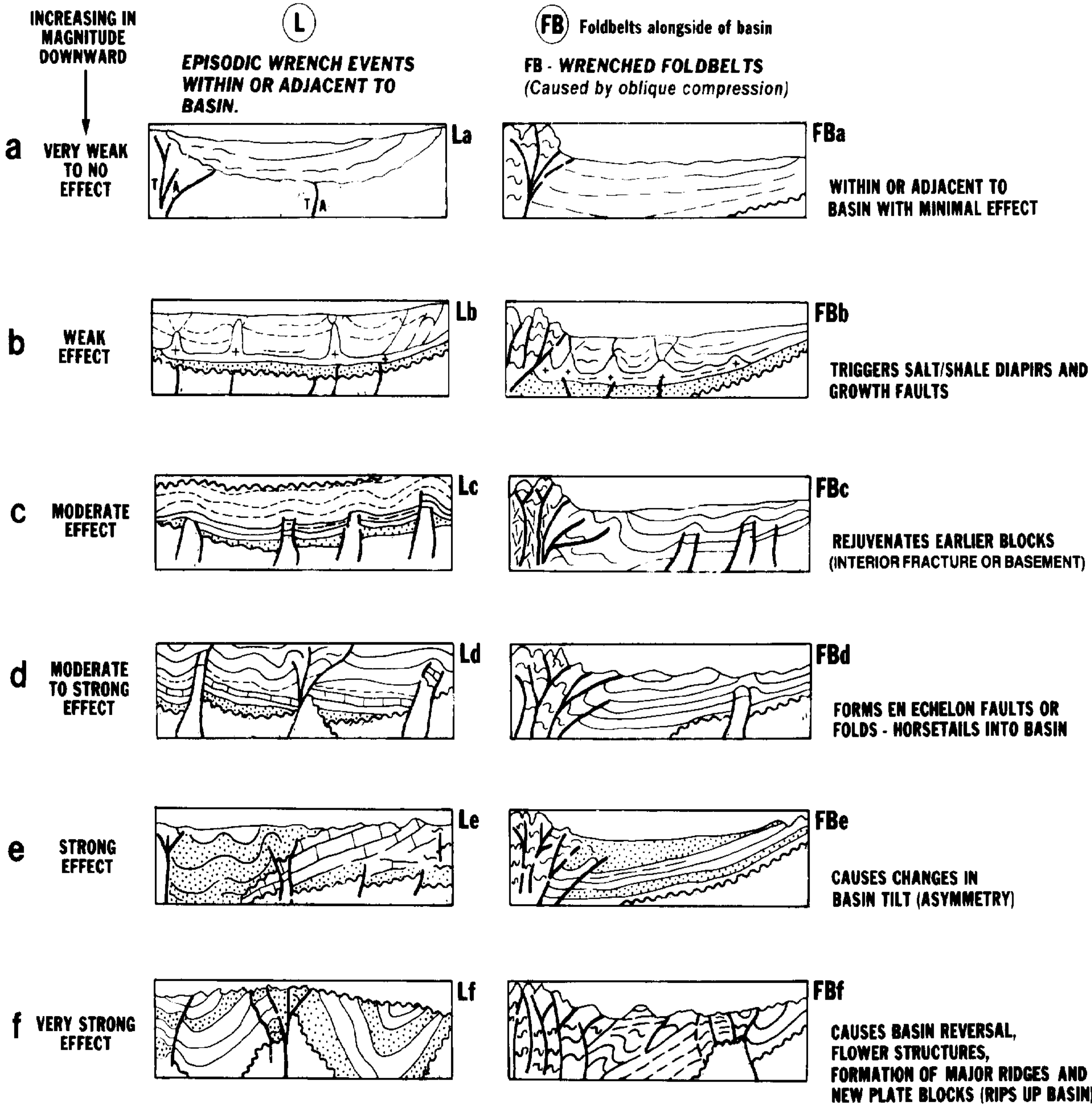

Fig. 15 -Tectonic modifiers of primary basin types (L and FB).

shale diapirs and growth-fault movements. Salt and shale diapirs, once triggered by an Lb event, could continue to grow by static load without further jiggling to keep them moving.

Lc and FBc illustrate the "c" effect which is rejuvenation of preexisting blocks, either interior fractures or basement. This jostling of older blocks can cause structural growth which may or may not reach the surface. Many of the world's giant fields owe their structural growth to "c" effects. It is important to note that the "a," "b," and "c" effects of domes or arches may not reach the top of the structured cycle, which was also the old ground surface.

The first modifier to reach the old ground surface as wrench-generated faults and folds is the "d" effect, which is rated as a moderate to strong event. It is convenient to divide the list of modifiers into the ones causing "weak" effects (a, b, c) and the ones causing "strong" effects (d, e, and f).

Ld and FBd are examples of classic wrenching. Relative plate movement is enough to cause en echelon faults or folds to be well developed. Here are found the first flower structures recognizable on seismic records or visible on the surface. Horsetails (a series of en echelon faults or folds) may be seen as fanning out of foldbelts or wrench zones into basins. The Ld flower structures are fairly modest ones and do not bring basement to the surface, as they do in stronger wrench events. Le and FBe are strong plate tectonic effects which significantly alter basin tilt, causing marked basin asymmetry. Tilt in one direction, or change of tilt direction, is an "e" effect.

Lf and FBf effects are the strongest episodic wrench events we record in a basin. Basins are turned inside out or "reversed" with synclines becoming anticlines.

Flower structures bring basement or very old rocks to the surface. These "f" effects also see the formation of major ridges or arches in a basin, the breakup of plates under basins, and consequent formation of new smaller basins out of the old megabasin. The basin can be ripped up extensively and still be called an "f" effect. However, if the basin is completely folded, destroyed, or altered by faulting, we would go one step further than "f" and call it a foldbelt (FB3).

It should be noted that any of the characteristic tectonics of the weaker modifiers may be found in any stronger one. For example, salt and shale diapirs and growth faults (Lb) may be found along with rejuvenated block movements (Lc). All of the tectonic parameters, salt domes, jostled blocks, flower structures, basin tilts, etc, may be found in Lf.