Chapter 1
Precambrian Terrains and Mountain Building Processes

1.1 Introduction

Orogenic (Gr. *Oros* means mountain and *genic* means birth) belts or orogens are some of the most prominent tectonic features of continents. These terms are, however, not synonymous to Mountain belt which is a geographic term referring to areas of high and rugged topography. Surely, mountain belts are also orogenic belts but not all orogenic belts are mountains. In this book, the orogenic belts, also called mobile belts, are termed fold belts because they are made up of rocks that show large-scale folds, and faults/thrusts and metamorphism with evidence of melting or high mobility in the core region during orogenesis. These belts are characteristically formed of (a) thick sequences of shallow water sandstones, limestones and shales deposited on continental crust and (b) deep-water trubidites and pelagic sediments, commonly with volcanoclastic sediments and volcanic rocks. Typical mobile belts, rather fold belts as titled here, have rocks that were deformed and metamorphosed to varying degrees and intruded by plutonic bodies of granitic compositions. Some fold belts are also characterized by extensive thrust faulting and by movements along large transcurrent fault zones. Even extensional deformation may be found in such belts. Most belts show a linear central region of thick multiply deformed and metamorphosed rocks bordered by continental margins, but some belts are also having oceanic margin on one side.

The application of the plate tectonic model to the study of the fold belts (or mobile belts) has revolutionized our ideas about how these belts form. We now believe that orogenic belts form at convergent margins because of long periods of subduction beneath the plate margin or because of the collision of two continents, of a continent with an island arc, or of a continent with an oceanic crust. Different types of fold belts form depending on the character of the colliding blocks and on which side one block overrides the other (Press and Siever, 1986).

The subduction of oceanic lithosphere gives rise to two different fold belts, depending upon the nature of the overriding plate. Subduction beneath oceanic lithosphere gives rise to the formation of island arc. Subduction under continental lithosphere gives rise to a linear fold belt on the overthrusting plate margin that runs along the subduction zone. Such fold belts are generally termed *Andean type*

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orogenic belts. The term Cordilleran type is no longer in use since the recognition that the Western Cordillera of North America formed by rather more complex processes than implied by this simple model (cf. Kearey and Vine, 1996). Another type of fold belts are *Collision fold belts* which develop when subduction adjacent to a continental margin brings another continent that make up an integral part of the subducting oceanic lithosphere. The resulting collision of two continents causes the creation of a collision fold belt by stacking of thrust slices of crust. Collision fold belts are complex since the geological record of collision always preceded by Andean type orogenesis. By studying the young orogenic belts, we can discover the relationship between orogenic structures and related plate tectonic activity. Similar structures in older orogenic belts can be used to infer the existence of similar plate tectonic activity in the geological past. It is commonly found that fold belts subdivide the stable Archaean cratons and show wholly ensialic deformation, with no rock associations that could be equated with ancient basins. In some Proterozoic fold belts, however, ophiolitic complexes have also been recognized, which can be attributed to processes associated with subduction zones, for example Wopmay orogen at the northwestern margin of the Laurantian shield (Hoffman, 1980). There are many examples in literature (Kroener, 1981; Condie, 1982) in which Precambrian orogenic belts have been explained in terms of modern plate tectonic processes. Gibb et al. (1983) have proposed that the structural province boundaries of the Canadian shield can be explained in terms of plate margin interaction during Proterozoic times. These structures making suture zones were formed by continent-continent collision of the juxtaposed continental masses, following the consumption of oceanic lithosphere. In this scenario we have to look for temporal and spatial occurrences of various major rocks as a function of plate motion or collision.

Orogenic belts are commonly at the margin of the stable shield or craton. For example, the Caledonian Mountain belt (Lower Palaeozoic) is seen on the western margin of the Baltic shield (mainly in Norway and Great Britain); the Hercynian or Variscan (Upper Palaeozoic) belt extends roughly from Western Europe to the Pacific with a notable N-S appendix in the Ural Mountains. The Tertiary fold belt of the Himalayan ranges is also trending E-W bordering the northern part of the Indian shield. The Alpine ranges occur mainly to the south of the Hercynian belt that it partly overlaps (Fig. 1.1). A similar overlap of the mountain belts is seen around the Siberian shield, with orogenic belts from Caledonian to Hercynian and to Tertiary, arranged more less concentrically around the shield; the ages of the fold belts decreasing radially outward. For example, the Hercynian belt of Western Europe includes Lower Palaeozoic rocks that were previously deformed in the Caledonian orogeny. In addition, the core of the Central Alps includes rocks metamorphosed in the Hercynian orogeny and re-metamorphosed in the Alpine orogeny. Since the older basement is rarely exposed, it is not possible to affirm that the Cenozoic cordilleras are not underlain by rocks as old as the Caledonian shield. Moreover, the older events may have been erased by later orogenies. These problems need to be recognized while discussing the evolution of the fold belts in the Indian shield.

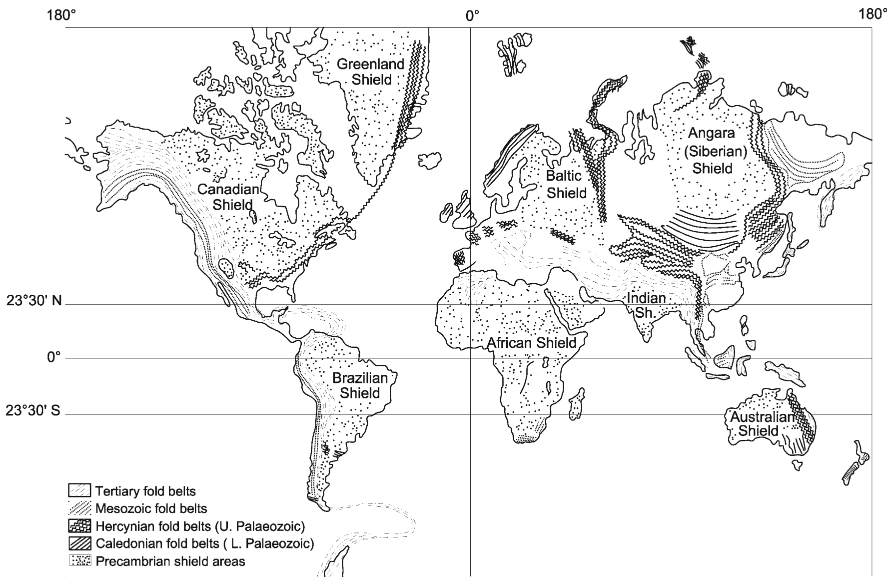

Fig. 1.1 Tectonic units of the continents: the Shields and Orogenic belts (after Umbgrove, 1947)

1.2 Archaean Terrains

The continents (including the shelf region) and oceans are the major features of the Earth surface. Continents are made of low-density sialic rocks and contain ancient (Precambrian) crystalline rocks-association of gneiss-granite-granulite/greenstone that forms the continental nucleus of Archaean age (2.5–3.8 Ga). These Archaean terrains are distinguishable because of metamorphic grade into two regions. One, high-grade granulite-gneiss regions that show amphibolite or granulite facies metamorphism, and second, greenstone belts that have low-grade (greenschist facies) metamorphosed volcanic rocks and associated sediments. The granulite-gneiss constitutes the bulk of the Archaean terrains. They are mostly quartzofeldspathic gneisses derived by deformation of felsic igneous rocks (e.g. granite becomes a gneiss on deformation) of tonalite-trondhjemite composition. There are minor metasedimentary rocks such as quartzites, metacarbonates, and metamorphosed iron formations. Deformed mafic-ultramafic complexes or layered gabbro-anorthosite complexes and orthoamphibolites make up the remaining gneissic regions. The high-grade gneisses show ample evidence of partial melting to produce migmatite and charnockite. The latter rock is characterized by the mineral hypersthene. These granulitic rocks range in composition from acidic, hypersthene granite (charnockite *sensu stricto*) through intermediate hypersthene-granodiorite (enderbite) to hypersthene gabbro (norite) or basic charnockite. The high-grade granulite-gneiss terrains are complexly mixed on a scale of tens to hundreds of kilometers with low-grade greenstone belts that contain mafic-ultramafic to silicic volcanic rocks and shallow intrusive bodies, volcanogenic and shallow water sediments.

The contact between greenstone belts and high-grade granulite-gneiss areas is complex. At places, the contacts are shear zones that modify or mask the original relationship. Elsewhere, greenstone rocks are deposited on older gneissic basement. In still other regions, gneissic/granitic rocks intrude the greenstone belt. The sedimentary rocks of the Archaean terrains have one of two broad associations: either they are immature volcanogenic sediments that are characteristic of the greenstone belts, or they are quartz-carbonate-iron formations assemblage occurring with deformed mafic-ultramafic layered igneous complexes in gneissic terrains. Another feature of the Archaean terrains is that their constituent rocks are intensely deformed and show more than one generations of folds (see Nisbet, 1987).

The oldest rocks on the Earth are as old as 3800 Ma and occur in widely separated Archaean terrains that are mainly tonalitic gneisses, metamorphosed basalts, and minor ultramafic and calc-alkaline volcanics. These metaigneous rocks occur as supracrustal sequences or greenstone belts. The two types of rocks are considered by some geologists as ancient vestiges of sialic continents and mafic oceanic crust (cf. Sharma and Pandit, 2003).

Greenstone belts of mafic to ultramafic volcanics with intercalated sediments generally occur in a synclinorium type structure, 40–250 km wide, and 120–800 km long. Most often, the greenstones are divisible into three stratigraphic groups. A lower group consists of komatiitic to mafic volcanics rocks with pillow structures, having a bulk composition similar to ocean ridge basalts. A central group of andesites and calc-alkaline volcanic rocks are similar in trace elements and REE to the rocks found on island arcs. An upper group comprises clastic sediments such as graywackes, sandstones, conglomerates, and banded iron formations with chert and limestones.

The origin of the greenstone belts is thought (non-uniformitarian model) to develop as a downwarp in the primitive crust in response to the load of injected high-Mg lavas and subsequent filling by sediments (Fig. 1.2). It is believed that the crust was thickened by compression of the trough and crustal melting at depths gave rise to diapiric granites (Windley, 1984). Most workers suggest a plate tectonic model for the formation of the greenstone belts and propose that they originated in subduction related setting, particularly in ensialic marginal basin. For Rocas Verdes complex in Southern Chile, Tarney et al. (1976) proposed that greenstone belts developed in ensialic marginal basins. Their model is shown in Fig. 1.3. According to this model, the greenstone belts developed on thin continental crust in a back-arc environment. They were covered by sediments that came from the flanking continents and volcanic arc. As the basin was subjected to compression, volcanics and overlying sediments were deformed and intruded by granites generated at depth. The back arc environment also explains how Archaean oceanic crust was preserved; if it originated in a normal ocean basin, it would probably have been destroyed by subduction.

Commenting on the granulite-gneiss belts, Windley (1984) pleaded that the geochemical characteristics of these belts could be replenished only when some form of plate tectonics was operative during Archaean times. It is by this mechanism that the mantle-derived continental crust was created during the Archaean (see also Sharma

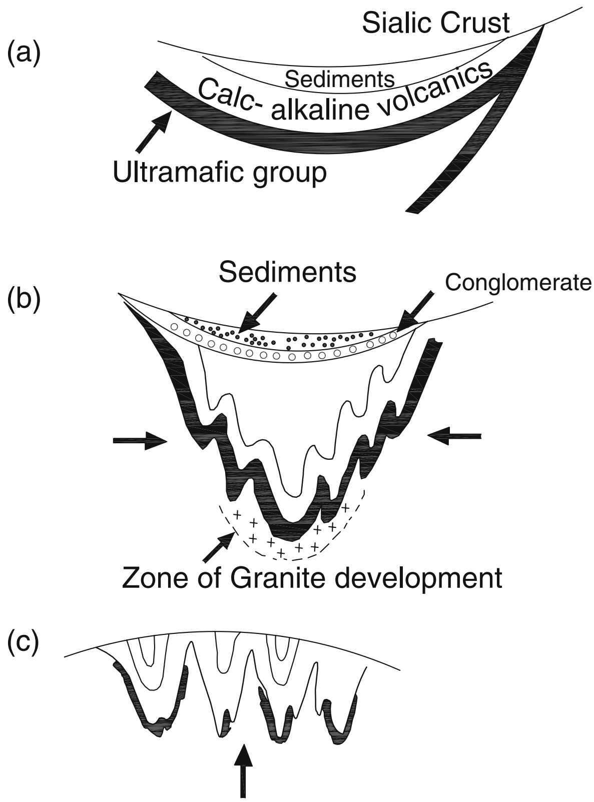

Fig. 1.2 Non-uniformitarian evolutionary model of Greenstone belt (after Windley, 1984). (**a**) Synformal structure; (**b**) crustal thickening; (**c**) uplift of deformed greenstones

and Pandit, 2003). Windley argues that this mechanism was probably not identical to present day plate tectonics as the Archaean time is documented to have high heat flow and thinner lithosphere. However, by combining the marginal-basin and subduction related tonalite batholith models, the Archaean plate tectonics could provide a mechanism for the complementary origin of both greenstone and granulite-gneiss belts. The proposed mechanism of Windley (1984) considers the Archaean crust being made up of a large number of small, relatively thin continental crustal blocks. Subduction of oceanic crust between these gave rise to back-arc spreading and formation of ocean-type crust (Fig. 1.4). It is by subduction only that calc-alkaline volcanics and tonalite magma were generated. It is suggested that by the end of Archaean, these small continental fragments welded into large sialic blocks characterized by alternate belts of greenstones and granulite-gneiss formed, respectively, from the oceanic crust and eroded tonalite batholiths. In this event, bigger and thicker plates dominated the Proterozoic lithosphere.

There are workers (see Moores and Twiss, 1995) who are reluctant to accept these simplistic applications of modern plate tectonic models to the Archaeans, due

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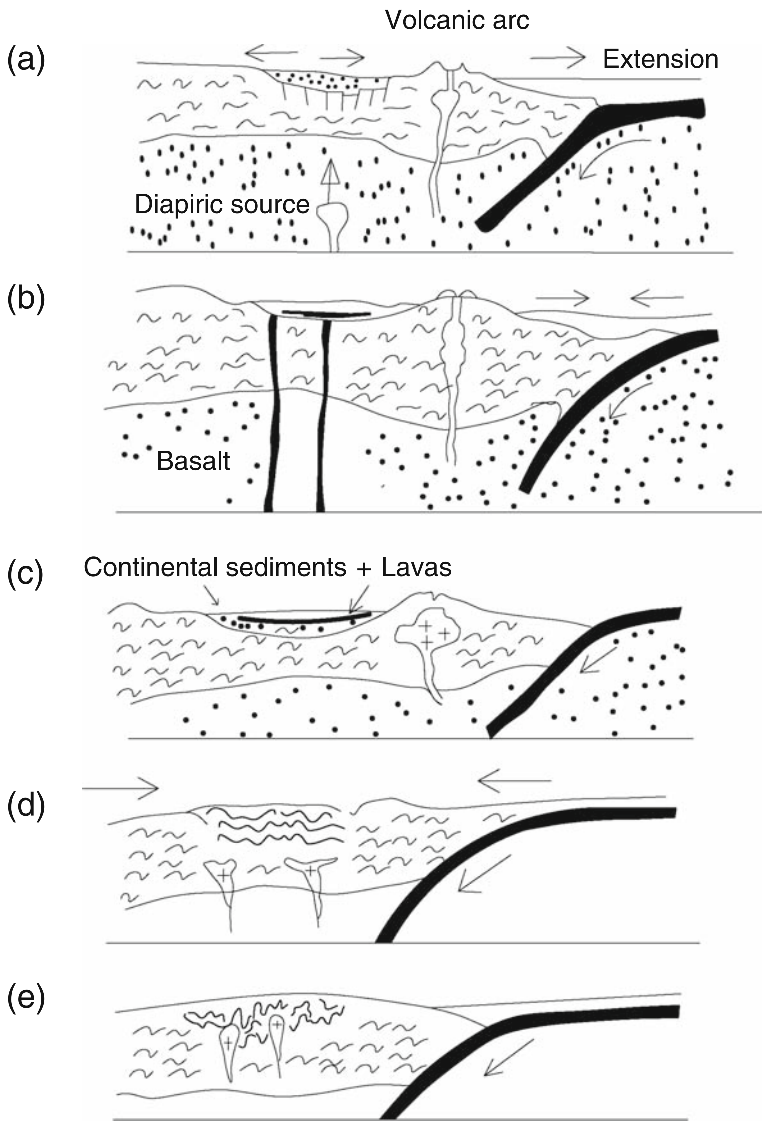

Fig. 1.3 Plate tectonic model for the formation of Greenstone belt exemplified by Rocas Verdes Complex, South Chile (after Tarney et al., 1976). (**a**) Extensional phase and basin formation; (**b**) volcanic phase and emplacement of basalt; (**c**) continental sediments and arc lavas as supracrustals; (**d**) deformation due to basin closure and intrusion of synorogenic tonalite; (**e**) emplacement of late potassic Granite

to several reasons. There are extensive differences between modern and Proterozoic tectonics. Although greenstone belts bear some resemblance to modern island arc or marginal sequences, significant differences cast doubt on a simple correlation. Ultramafic lavas (komatiites), abundant in the Archaean, are almost entirely

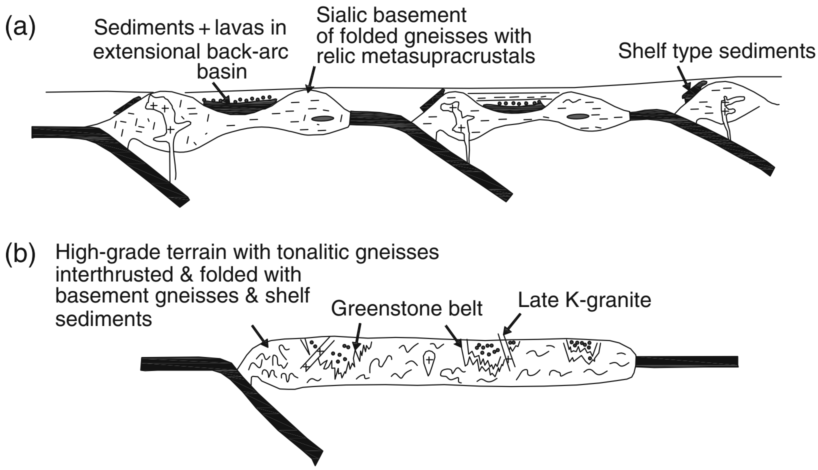

Fig. 1.4 Plate tectonic model for the evolution of greenstone-gneiss-granulite in the Archaean (redrawn from Windley, 1984). (**a**) Several small continental plates with ensialic back-arc basins. The Andean-type margins are invaded by mantle-derived tonalitic batholiths and plutons; (**b**) aggregations of the continents to produce a large continental plate consisting of greenstone and granulite belts

lacking in modern marginal basins or island arcs. If the greenstone belts represent marginal basins, they lack the accompanying mature arc that is ubiquitously associated with modern marginal basins. If they are island arcs, they differ in composition from modern island arcs and seem to lack the compositional zonation possessed by modern arcs that reflect the polarity of the subducting slab. The quartzite-carbonate-mafic-ultramafic association present in Archaean gneissic terrains has no clear modern equivalent. Many workers (e.g. Moores and Twiss, 1995) suggested that the sedimentary rocks represent shelf sequences and the associated mafic-ultramafic complexes may represent the Archaean equivalents of oceanic crust. Despite these differences, however, mantle convection almost certainly did occur and some forms of plate tectonics in early Proterozoic and Archaean times cannot be ruled out (cf. Sharma and Pandit, 2003).

1.3 Proterozoic Terrains

The Archaean terrains became stable for first time in Proterozoic (< 2500–600 Ma) and served as basement for the deposition of Proterozoic sediments. Precambrian regions of the Earth's crust that have attained tectonic stability are called the *cratons*. The evidence of this relative stable tectonic environment is documented in the occurrence of mature sediments such as quartzites, quartz pebble conglomerates, deposited on eroded, rather peneplaned Archaean basement rocks. These mature sediments define regionally extensive stratigraphic units. Quartzites are often intercalated with abundant iron formations (interstratified magnetite/hematite rocks, iron carbonates and iron silicates). These are mostly undeformed cratonic sediments, like the platform sequences, lying upon a pre-existing older basement, indicating the existence of large stable continental region in Proterozoic time. They host many vast Precambrian placer gold and uranium deposits as well as most iron deposits of the world. These sedimentary sequences extend over large areas in the Indian shield but are not the theme of this book. Interested reader may refer to the relevant source either in journals or in published books/memoirs. Another type of Proterozoic rocks making up the fold belts are either peripheral to the cratonic areas or are sandwiched between two cratonic regions within the Indian shield. In other words, the cratons were sutured or accreted by the Proterozoic fold belts. These fold belts with their characteristic geological setting and tectono-thermal evolution constitute the main theme of this book. They can be called fold belts to indicate major belts of pervasive deformation and tectonic mobility. Fold belts are most prominent tectonic features of the Indian shield. These belts form a linear or arcuate belt of deformed and metamorphosed rocks. They were the loci for the development of former intracratonic basins and rifts as they are formed of (i) thick series of shallow-water sandstones, limestones and shales, all deposited on continental crust and (ii) deepwater turbidites and pelagic sediments, commonly with volcaniclastic sediments and volcanic rocks deposited on ocean floor. Typical fold belts have been deformed and metamorphosed to varying degrees and intruded by plutonic rocks of granitic composition. Some fold belts may display structural symmetry on either side of the core made up of high degree of metamorphic rocks and plutonic bodies (Moores and Twiss, 1995).

The Proterozoic fold belts or orogenic belts are of two types. Some are multiply deformed regions of both supracrustals and Archaean basement rocks. Others exhibit thick sedimentary sequences deposited in linear troughs, presumably along ancient continental margins, subsequently deformed and recrystallized into linear fold-and-thrust belts, like those of Phanerozoic belts. In addition, large anorthosites occur as distinctive igneous-metamorphic rock-suite that appeared in Proterozoic time (1000–2000 Ma). These anorthosites are intimately associated with granulites and the Eastern Ghats mobile belt would represent a good example of this type of fold belt. Other Proterozoic fold belts that are covered in this book are the Aravalli fold belt of Rajasthan in western India, the Satpura, Mahakoshal, Sakoli, and Dongargarh fold belts in central India, the Singhbhum mobile belt in eastern India and finally the Pandyan mobile belt in southern Indian shield. For each of these fold belts of the Indian shield, we will consider the plate tectonics theory for their development, because these Proterozoic fold belts show deformation and metamorphism which can be attributed to collision tectonics. Being intensely eroded, these fold belts display multiply deformed and metamorphosed thick sequences of miogeosynclinal rocks and the basement in the core or root zone. High angle fault zones are also numerous in almost all these fold belts. Therefore, the study of these Proterozoic fold belts would give some information about plate tectonic interactions for the first 95% of Earth's history. By studying these features, we have an opportunity to decipher part of the tectono-thermal history of the Earth that is preserved nowhere else. Before the acceptance of plate tectonics, the formation of the fold belts was conceived essentially by vertical movements and it is interesting to know the early views on the origin of fold belts, known then as mountain ranges.

1.4 Mountain Building Processes

Earlier, the mountain belts were explained by *contraction hypothesis* according to which the wrinkles formed during cooling of the Earth and decrease of its radius became the mountains of the Earth's crust. According to this hypothesis, the continents were essentially stationary on the Earth's surface. However, vertical movement was accepted to explain the presence of marine rocks in high mountains like the Himalaya. The driving force for such vertical movements was lately thought to be isostasy—the theory initially proposed in 1865 by George Airy to explain gravity measurement in India.

1.4.1 Geosynclinal Theory

In the middle of the nineteenth century, the American geologist James Hall proposed the geosynclinal theory. According to this theory the vertical movements of tectonic origin developed a subsiding linear trough or a geosyncline in which thick, shallow water sediments were deposited along both sides of the trough (miogeosyncline) and deep water sedimentary and volcanic rocks in the centre (eugeosyncline). Following the deposition of these thick sequences, deformation occurred, with "eugeosynclinal" sediments becoming the most deformed and metamorphosed. Deformation was conceived to be symmetrical, with thrusting of all rocks in both directions away from the centre over the flanking undeformed continental platforms. The mountain building process (orogenesis) was conceived to have three main phases (Fig. 1.5).

1.4.1.1 Geosynclinal Phase

Here a linear depression appears in the crust into which sediments derived from surrounding regions of the crust were deposited, commonly accompanied by extrusion and intrusion of basic volcanic rocks (see Dickinson, 1971). Sedimentation continues in a more or less steady state, but not uniform over the whole basin in either composition or thickness due to variations in depth of water, proximity to source area, etc. In the shallower part of the basin, rocks such as limestones and sandstones tend to predominate, and in the deeper parts, a much thicker sequence of "flysch"-type sediments is generally found. It is in this deeper part that basic volcanic activity is generally concentrated (Fig. 1.5a). These differences from place to place in sedimentary type (or facies) are most useful to the geologist in helping him to reconstruct the form of the ancient basin, or, conversely, to establish directions and amounts of relative movements that have taken place during folding and thrusting.

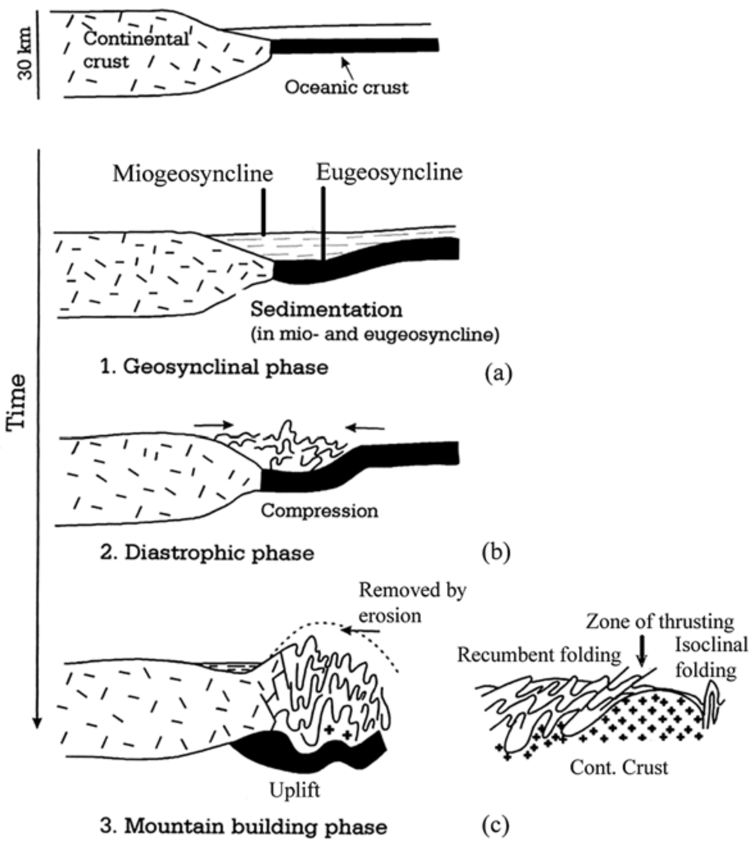

Fig. 1.5 Geosynclinal model for the development of an orogenic belt

1.4.1.2 Diastrophic Phase

The diastrophic or tectogenic phase involves strong deformation of the sediment-filled basin (Fig. 1.5b). The structures formed in this phase are in detail unique to each geosyncline; commonly occurring features are: first, wide spread formation of folds of various sizes from a few cm to many km, and second, thrust faulting in many scales. Flat-lying rock masses moved for very long distances along large faults. The folding and low-angle thrusting are intimately related and seem to proceed together. The deeper levels, including the basins in some of the pre-existing crust, become the sites of intense metamorphism, commonly invaded by granite bodies. At the upper and marginal levels, metamorphism is less intense or absent, but the pre-orogenic basement can become involved in the deformation and appear as slices of crystalline rocks, generally in the cores of the large folds in the covering sediments. The metamorphism of the basin sediments along with the basement have been classed into three *depth zones*: Epizone (characterized by minerals like chlorite, muscovite, epidote), Mesozone (characterized by garnet, staurolite, hornblende, etc), and Catazone

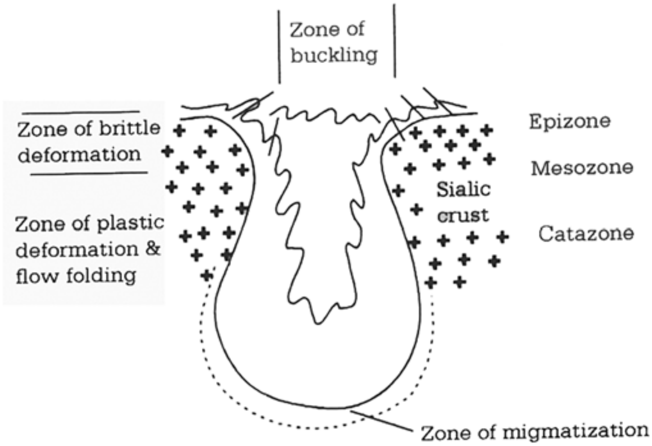

Fig. 1.6 Depth zones and distribution of deformed and metamorphosed rocks in a crustal section

or Katazone (characterized by sillimanite, cordierite, hypersthene). In the lower part, the geosyncline becomes mechanically weak or mobile during metamorphism. The anatectically-formed granite can rise in the upper or middle zones in form of a diapir or dome (Fig. 1.6).

1.4.1.3 Orogenic Phase

This is the mountain building phase characterized by uprising of the deformed and metamorphosed geosynclinal sediments and basement crust. Sometimes the upper folded layers seem to become detached from its position and slide under the influence of gravity as extensive thrust sheets to lie over the less deformed parts of the geosynclinal margin and even onto the undeformed "foreland", as in the Garhwal Himalaya (Gansser, 1964). During the later part of the orogenic phases, marginal and intermediate depressions form which rapidly fill-up with coarse sediments (molasse), produced by rapid erosion of the rising mountain range (Fig. 1.5c).

The greatest failing of geosyncline theory was that the tectonic features were classified without having an understanding of their origin. In this theory, both continents and oceans were considered permanent. A major weakness of this model was that there was no explanation for what caused geosynclines to develop and subsequently to deform. Another problem with the geosyncline model was that it could not explain the very large amount of shortening exhibited by the nappes in the Alps. This led the Austrian geologists Ampferer and Hammer in 1911 to formulate the hypothesis of "Verschluchung" (the downbuckling or subduction) of the European platform beneath the Alpine geosyncline. In 1905, the German geologist, Gustav Steinman, recognized the intimate association of serpentinites, pillow lavas and radiolarian cherts in the Alpine orogen and proposed that the sediments were laid down in deep water over a rapidly subsiding continental platform. This recognition led to the belief that geosynclines also had deep-water environment of deposition.

Lately, some authors like Kearey and Vine (1996, p. 7) do not recommend the use of the geosynclinal terminology, e.g. eugeosyncline and miogeosyncline for sediments with and without volcanic members respectively. They state that "the term geosyncline must be recognized as no longer relevant to plate tectonic processes." Nevertheless, it must be noted that the relation of sedimentation to the mobilistic mechanism of plate tectonics allowed the recognition of two specific environments in which geosynclines formed, namely passive continental margins and subduction zones. However, the terms mio- and eu-geosyncline are useful terms to indicate specific depositional environment and would be retained here.

1.4.2 Plate Tectonics Theory

A direct challenge to the permanency of continent and ocean basins came in 1915 when Alfred Wegener proposed his hypothesis of *continental drift.* This hypothesis witnessed long controversies. Nevertheless, two important findings are quite interesting. In 1930, a study of gravity at sea by the Dutch geophysicist F.A. Vening Meinesz disclosed that deep-sea trenches in the Caribbean and in Indonesia were associated with negative gravity anomaly to suggest downbuckling of the crust into the mantle—a concept that was called tectogene. The formation of tectogene, according to Arthur Holmes, is in response to a downgoing convection current in the mantle. At the same time in 1928, the Japanese seismologist, K. Wadati, recognized that earthquake sources beneath Japan were located along an inclined planar zone that extends from the trench east of Japan and dips westward under the islands. The revolution began in 1960 with the circulation by Hess of a manuscript entitled, "Evolution of ocean basin and History of ocean basins." In this manuscript, Hess (1962) argued that the oceanic crust was young and was created over rising limbs of mantle convection cells. Taking support of the palaeomagnetic data of Blackett, Irving and Runcorn and the drift theory of Wegener, R.S. Dietz coined the term *Sea floor spreading.* In short, the sea floor spreading can be likened to a conveyer belt in which lithospheric plates were transported laterally carrying the continents with them.

The development over the last four decades of the Plate tectonics theory gave us a way of understanding mountain building processes. Therefore, we must have a relevant knowledge of this theory that would form the required background about the evolution of fold belts of younger and older ages. In plate tectonics, unlike earlier orogenic hypothesis, the sediments deposited in widely separated depositional sites are telescoped and sutured onto the convergent plate margins. In plate tectonics it is not essential that the strata must be deformed and metamorphosed, as it depends on the location of the deposition, e.g. sediments deposited away from a trench will escape deformation during plate convergence. Furthermore, the regional metamorphism of the basinal sediments is not depth controlled but depends on the particular gradient that can prevail in the given crustal segment, irrespective of the depth.

The basic concept of the Plate Tectonics is that the rigid outer shell of the Earth, *lithosphere,* lying above the Low Velocity Zone (LVZ) or Asthenosphere, is divided by a network of boundaries. These boundaries were originally distinguished on their seismicity and later by the presence of mountain belts/ocean ridges, subduction zones/trenches, and by chain of volcanoes and earthquake belts, separating the rigid outer shell into separate blocks which are termed *Lithospheric Plates.* The plates (80–150 km thick) could be exclusively continental or oceanic or both. Today there are six major plates and a number of smaller plates. The hypothesis of sea-floor spreading, put forward by Hess (1962) led to establish that the lithospheric plates were transported laterally (carrying the continents over them) by convection currents in the mantle. The plates move away from *ocean ridges,* which are the sites of seismic and volcanic activity and abnormal high rates of heat flow. Because of tensional forces, the ridge develops a central rift valley, which is devoid of any sediment cover, but further away from the ridge, sediments become more abundant and also thicker and older.

1.4.2.1 Plate Margins

Three types of plate margins are recognized, depending on their movements. In following description, we examine the nature of the plate boundaries and their geological characteristics (Fig. 1.7).

Constructive Plate Margin

The constructive plate margins develop when continents rift and move away to form new ocean basins by upwelling and solidification of magma generated by decompression melting of the upper mantle underneath. These margins initiate at

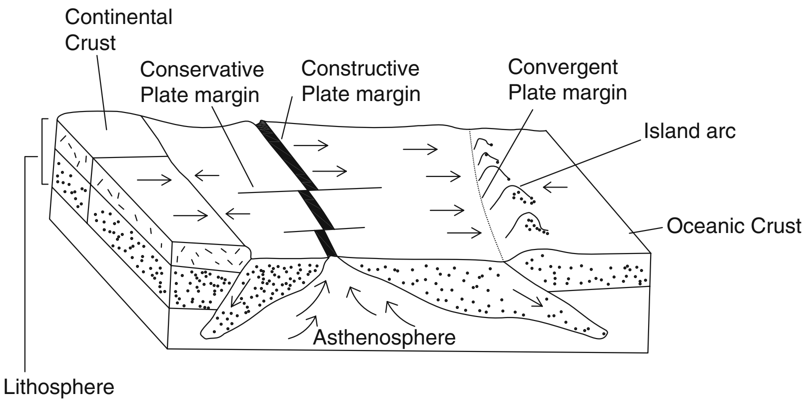

Fig. 1.7 Block diagram showing the different plate margins. Mantle material rises below the Constructive plate margins (ocean ridges) and plate material descends into the mantle at Convergent (destructive) plate margins (ocean trenches). At Conservative plate margins (transform faults) plates slide past each other, without getting created or destroyed

a divergent plate boundary or ocean ridges and are also referred to as divergent plate margins. The ocean ridges can be conceived as the sites of creation of new oceanic material and can be called Atlantic type margins, considering the geographic region where it is characteristically developed now. These are, in fact, passive margins and include a coastal plane and a submarine topographic shelf of variable width, generally underlain by a thick sequence (10–15 km) of shallow-water mature clastic or biogenic sediments. Shelf region passes into a steep slope toward ocean basin. Normal faults are characteristically found in sediments along these margins. The continental rift zones, e.g. African rift or Red Sea rift, represent incipient constructive boundaries.

Convergent Plate Margin

The convergent plate margin is located at ocean trench where two plates converge and when denser of the two plates sinks, along an inclined plane or *Benioff Zone,* below the other to be eventually resorbed into the mantle. These are also called the destructive plate margins, typified by the Andean type margin. These margins exhibit an abrupt topographic change from deep-sea trench to a high belt of mountain. Shelf region is very narrow or absent. Mountain chains along these margins are characterized by a chain of volcanoes of andesitic composition. Intense deformation near the trench results into thrust complex. Most subduction zones now are situated at island arc within the oceans but certain subduction zones border the continents. There are thus two types of convergent boundary now. The first follows the deep ocean trenches and the second follows the belt of young mountain ranges of Alpine-Himalayan chain.

Conservative Plate Margin

The conservative plate margins are faults where two adjoining plates move laterally past each other so that lithosphere is neither created nor destroyed. The direction of relative motion of the two plates is parallel to the faults. Conservative plate margins occur within both oceanic and continental lithosphere, but the commonest conservative plate margins are oceanic *transform faults* (Wilson, 1965). The best-documented example of a large continental transform fault is the San Andreas Fault of California, and therefore these margins are called the Transform or California type. These margins are characterized by sharp topographic differences between ocean and continents. They are marked by active strike-slip faulting, sharp local topographic relief, poorly developed shelf and irregular ridge and basin topography and many deep sedimentary basins.

The oceanic faults make a very prominent feature of the oceans. The ocean ridge crests are repeatedly offset by parallel sets of such faults. The displacement along these faults could be dextral (right lateral) or sinistral (left lateral). Since the spreading direction of the ocean ridges remains parallel to these faults, the divergent motion away from the ridge axis would be "transformed" to a transcurrent motion along such a fault, and then perhaps transformed again to convergent motion at a

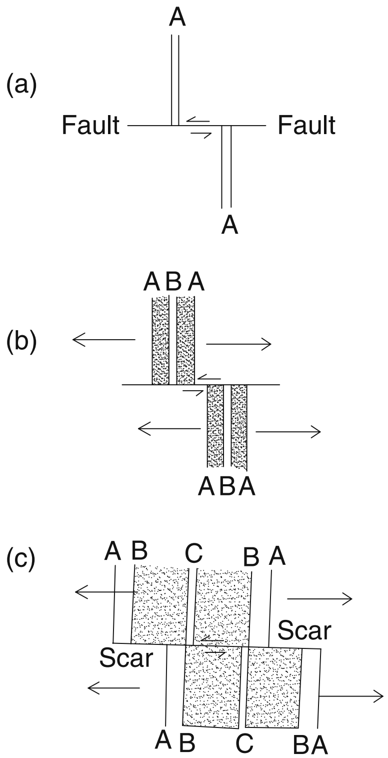

Fig. 1.8 Representation of development of a ridge-to-ridge transform (adapted from Verhoogen et al., 1970, Fig. 3.75). (**a**) Oceanic ridge (*double line*) along which new crustal material added is displaced by a sinistral strike-slip fault (*single line*); (**b**) addition of magma between line A and A and slip on the fault; (**c**) addition of material between lines B and B; scar of previously active sections of the fault extend laterally into the surrounding crust. Note that the sense of slip on the transform fault is opposite of that inferred from the displacement of the ridge by strike-slip faulting

trench (Fig. 1.8). Therefore, Tuzo Wilson (1965) called these faults *transform faults*, fundamentally different from the strike-slip faults on land. A transform fault must be parallel to the direction of relative motion of the lithospheric plates on either side and is therefore controlled by the relative velocity of the two plates. Strike-slip fault, on the other hand, is controlled by stress. By using transform faults and the spreading rates, the relative angular velocities of several plate pairs have been established (varying from 1 to 5 or more cm per year).

The plate margins also have distinct geologic characteristics. Constructive plate margins are characterized by basaltic volcanic activity along an axial rift. By contrast, destructive plate margins occurring at the junctions between oceanic versus oceanic or oceanic against continental plates are characterized by the presence of island arcs or continental volcanic chains, associated with intrusive magmas of intermediate composition and metamorphism at depth. Conservative plate margins, shown as transform faults, are generally devoid of volcanic activity.

Plate tectonics is an expression of thermal convection within the Earth. The convection occurs by the mass movement at surface and at depth. According to one hypothesis, the convection may be whole mantle whereas other hypothesis considers that the convection is shallow mantle in nature. Rising convection currents at ocean ridges (where the ocean plates move away from the ridge) brings heat to the surface.

The cold moving plates on encountering the opposing moving plate descends along a trench (due to descending convection currents) and heats up to be absorbed in the deep mantle. The trench dips below an adjacent continental margin or an island arc. This inclined zone is an activity of earthquake and is called a *Benioff Zone*. It is a critical piece of evidence in favour of the process of subduction.

In short, the ocean ridges can be conceived as the sites of creation of new oceanic material and the trenches as the loci of destruction.

1.4.2.2 Physiography of Plate Tectonic Units

This section gives a brief account of geological features and environment of different physiographic units of the plate tectonics, including the plate margins.

All convergent margins, regardless of their length, age, stage of development, display an overall similarity in topography. The orogens developed at plate convergence possess a number of features in common, for example a trough, a foreland, or undeformed plate on either side, fold-and-thrust belt, an internal crystalline core zone of deformed and metamorphosed sedimentary and volcanic rocks, granitic plutons, and a suture zone that may not be recognizable in all older fold belts. The *sutures* are the boundaries between crustal blocks that were carried on two different plates and that may have been juxtaposed by plate movements (Dewey, 1977). The recognition of sutures in the continental geology is essential for the older fold belts as it is for the Phanerozoic ones. Most sutures are recognized by the presence of ophiolites but these rocks may not always be present in the Proterozoic fold belts. In the ancient fold belts, sutures can be identified either by characteristics of the boundary itself or by a major discontinuity across the boundary, such as marked changes in lithology, geologic history, structural style, palaeomagnetic vectors, or faunal assemblage. Orogenic belts record the subduction of one plate beneath another, as well as collisions between crustal blocks, such as two continents, a continent and an island arc, or a continent and an oceanic plate. Because all oceanic crust older than 200 Ma (Early Jurassic) has been subducted, Proterozoic fold belts are the main repositories of information about early plate tectonic interactions. By studying these features, we can possibly decipher the tectonic history of the Earth preserved in these fold belts and nowhere else. The following description of physiography of the different elements of the plate tectonic environment will be helpful for a greater appreciation of the plate tectonic theory. Those interested in more details should consult Moores and Twiss (1995). The description given below starts across the margin of the downgoing plate (Fig. 1.9) and covers all elements successively.

Outer Swell

The Outer Swell is a topographic bulge that develops in the down-going plate before the plate bends into the mantle (see Fig. 1.9). This bulge or swell is a few hundred meters above the abyssal plain. It is suggested that this outer swell is the result of elastic bending of the plate as it starts descending into the mantle. It should obviously have normal faults parallel to the trench.

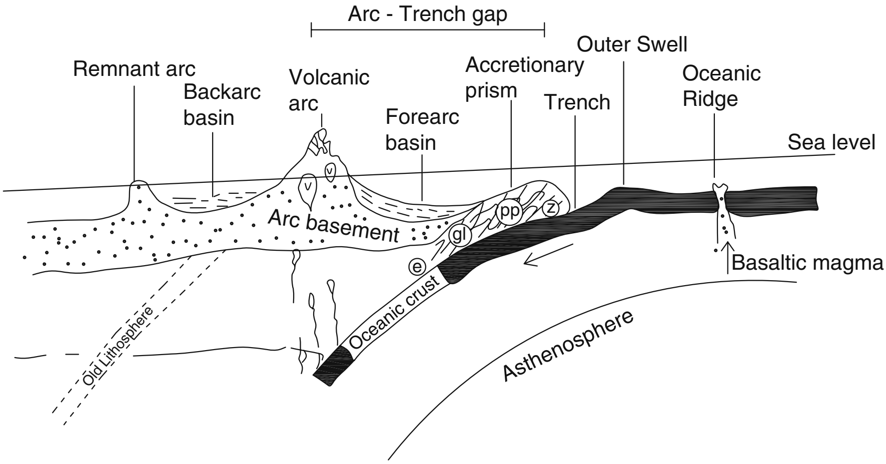

Fig. 1.9 Physiographic-cum-geologic units in a convergence of the Ocean-Ocean plates (see text for details). e = eclogite facies; gl = glaucophane schist facies; pp = prehnite-pumpellyite facies; z = zeolite facies

Trench

It is a deep topographic valley that forms downwards towards the boundary between the subducting and overriding plates (see Fig. 1.9). It is typically 10–15 km deep and is continuous for thousands of kilometers. Most trenches contain flat-lying turbidite sediments deposited by currents flowing either down into the trench away from the overriding plate, or along the axis of the trench.

Arc-Trench Gap

It includes the entire region between the trench and the Volcanic Arc. It consists of the steep inner trench wall that flattens into an area of gentle slope called the fore-arc basin (or Upper Trench Slope). The two areas, namely the trench and the fore-arc basin, are separated by a small topographic ridge—the Outer ridge, not to be confused with the outer swell, described earlier. The fore-arc region may be underlain either by a thick wedge of mostly deformed sedimentary rocks, known as the Accretionary Prism, or by deformed arc basement rocks covered in places by thin layer of sediments.

Accretionary Prism

It forms on the inner wall of an ocean trench. It develops when trench-fill turbidites (flysch), and perhaps also the pelagic sediments and underlying oceanic crust, are scrapped from the descending oceanic plate by the leading edge of the overriding plate to which they are welded. The accretionary prism is the main site of crustal deformation in a subduction zone. The deformation begins at the foot of the inner trench wall. Rocks in the accretionary prism are cut by numerous imbricate thrusts that are sympathetic to the subduction zone, i.e. they dip in the same direction. Because of these imbricate thrust faults, mostly listric, the accretionary sediments define a series of wedge-shaped pockets within which are developed complex folds verging towards the trench (see Fig. 1.9). The deformed rocks in the accretionary prisms are mostly sediments derived from either overriding or down-going plates. In some cases, seamounts from the down-going plate are being incorporated into the accretionary prisms. Sediments from the arc region are added to the top of the accretionary prisms when they are deposited in basins in the fore-arc area. They are also carried by the turbidity currents into the accretionary prism as subduction carries them back toward the arc and the basal thrust faults propagate out into the undeformed sediments (cf. Moores and Twiss, 1995). This process is called *off-scrapping*. It is by scrapping that the deep-sea sediments that accumulated on the downgoing plate are incorporated into the accretionary prism. Offscrapping results in progressive widening of the accretionary prism, i.e. outward growth of the prism. With continued subduction, older thrust wedges are gradually moved upward and rotated arcward by addition of new wedges to the base of the accretionary prism. As a result, the older thrusts become more steeply dipping with time. Offscrapping, also called underplating or subcretion, is also responsible for decrease in the dip of the subduction zone as the Volcanic Arc becomes more mature.

Deformation structures in the accretionary prisms reveal more than one generation, more clearly visible in soft sediments. In some cases, the deformation in the accretionary prism is so intense that any pre-existing stratigraphic continuity is destroyed. Such chaotic deposits are called *mélanges*. During the deformation, a fore-arc basin may develop between trench and the island arc in a trough. At some subduction zones, continental basement or old rocks of the island arc can be traced out to the lower trench slope, nearly devoid of accretionary prism. Here, only thin sediments accumulate in basins on the basement rocks and very little, if any, sediment occurs in the trench. The arc basement is commonly cut by normal faults down thrown on the trench side. Many trenches show relatively undeformed sediments on down-going plate that extend for several kilometers beneath deformed rocks of the trench inner wall.

Fore-Arc Basin

It is a trough between trench and island arc (see Fig. 1.9). It covers the oldest rocks of the accretionary prism. The sediments deposited in this basin are derived from the volcanic arc and hence clastic and carbonates with some fine-grained turbidite deposits.

Volcanic Arc

Beyond the fore-arc region in the active volcanic arc itself, older rocks occurring below a topographically higher region constitute the Frontal Arc, also called the *Arc Basement.* This should not be confused with the fore-arc region. In the Island Arc, also known as Volcanic Arc or Magmatic Arc, the arc basement is a shallow marine platform or an emergent region of older rocks. In continental arcs, the continental platform of older rocks stands 1–5 km above sea level. Most island arcs consist of volcanoes that have an elevation of 1–2 km above sea level, irrespective of the elevation of the basement. In contrast, the elevation of the continental arcs depends on the elevation of the basement. The arc basement, i.e. Frontal arc consists of older, deformed and metamorphosed rocks on which modern Arcs are built. However, the character of the basement is quite variable. Continental arcs such as the Andes are built on a continental basement complex.

In Island arc, we have two types of sediments, namely the clastic and carbonate. The clastic sediments consist of debris from active volcanoes. They range from fine sand to coarse conglomerates and breccias deposited near the source regions. Pumice is the common constituent of these sediments. In continental arcs, sediments are predominantly subaerial and subordinate marine. Most island arc sediments are marine, reflecting deposition in deep sea fan complexes that are shed from active volcanic islands. In tropical regions, active island arc contains fringing carbonate reefs. These carbonate deposits are flooded intermittently with volcanogenic debris. Thus, carbonate and volcanic deposits interfinger together with each other. The volcanic arc in the region of active magmatic activity is marked on the surface by a chain of volcanoes, generally andesitic in composition. Large plutonic bodies of batholithic dimensions are unexposed in recent Arc but abundant in ancient continental arcs. In plan view, the volcanoes are spaced along the Arc at a fairly regular interval of approximately 70 km. In both continental and oceanic arcs, igneous complexes are correlatable with depth or to the subducting plate. Increase in Al, Na and K for a given rock type has been correlated with the depth of Benioff Zone in a number of Arc complexes. Such compositional variations are of great value for inferring the direction of dip of ancient subduction zone. However, for ancient arc, these compositional variations must be used with caution for a number of reasons, as stated by Moores and Twiss (1995). These are: (1) the same abundance in elements in a single magmatic system as it crystallizes and differentiates; (2) the same differences in the same place through time as the subduction changes its dip or its depth or as it migrates with respect to the magmatic center; (3) the high susceptibility in alteration of many major and minor elements in rocks during metamorphism and weathering. These processes severely affect the elements, namely Na, K, Si, and to lesser extent Al, that are used to characterize the volcanic rocks of the Arc; (4) some Arcs do not show any variation in composition with depth; (5) the cause of variation of chemistry with depth is not clearly understood because of unclear origin of the magmas and degree of contamination of magmas. Metamorphism in arc and arc basement (frontal arc) reflects the high heat flow that existed in the region. In continental arcs, the arc basement is a continental crust and is generally above the minimum melting temperature of the granite. This fact alone could explain the abundance of granitic batholiths in continental arc regions. Under the volcanic arc a high-temperature, low-pressure metamorphism occurs and when these rocks occur adjacent to the fore-arc region near trench with low T, high P environment, we have *paired metamorphic belts.* The paired metamorphic belts are often parallel to the plate boundary over extended distances and the two belts are of the same age and are typically separated by major faults. In Miyashiro's classic model for Japan, the high-pressure belt is located in the accretionary wedge between a subduction zone and an island (magmatic) arc, whereas the low-pressure belt is up to several hundred kilometers farther inland from the plate margin.

Back-Arc Basin

The back-arc basins are small ocean basins behind the volcanic arc (Fig. 1.9). They are also called marginal basins. In island arcs, these basins have oceanic crust (structure and abyssal depth) similar in composition to those of the ocean basins. Here the basins are bordered by remnant arc or Third arc that are linear topographic ridges composed of thicker crust, comparable to island arcs. These back-arc basins receive the sediments derived from the island arcs. In continental arcs, the sediments are deposited in the basins that form on top of the continental platform toward the continental interior. These epicontinental basins are called Retro-arc basins or foreland basins. Back-arc basins of oceanic island arcs have two characteristics. Some are composed of entrapped old oceanic crust; others are composed of younger lithosphere. The latter typically show seismic activity, high heat flow, and sea-floor spreading. The composition of volcanic rocks in modern back-arc basins is variable. Some rocks occurring away from the arc have similar chemistry as the mid-oceanic ridges. The volcanic rocks found near the arc are similar to the rocks of the main arc. The remnant arcs also show the same igneous and sedimentary rocks that characterize active volcanic arc, but they are older than the associated active island arcs and the marginal basins. The remnant arc occurring as ridges seems to be the remnants of a pre-existing arc that split to form the sides of a developing marginal basin. Therefore, the origin of remnant arcs is closely related to the origin of the back arc basins themselves. In the back arc basins, clastic sediments derived from the arc show interfingering relationship with the lavas and pelagic sediments. The back-arc sediments show folds and thrusts toward the interior of the continent, although the back arc region of the oceanic and continental arcs is marked by extensional tectonics and subsidence relative to the arc itself.

The back-arc basins are considered to have developed in response to tensional tectonics. Several origins of the back-arc basins have been proposed (see Moores and Twiss, 1995). Some authors believed that an existing island arc rifted along its length in response to tensional forces and that the two halves, corresponding to the island arc and the remnant arc, separated to give rise to the marginal basin (Fig. 1.9). Others suggested that the island arc, being most ductile, undergone initial rifting; subsequent widening gives rise to a back-arc basin.

Other back-arc basins owe their origin to the formation of new oceanic crust behind an island arc. There are three ways by which marginal basins could originate (Fig. 1.10). First, when forceful injection of a diapir from the subducting slab causes extension of the overlying oceanic crust (Fig. 1.10a). Second, when convection currents driven by the drag of the down-going slab cause extension of the oceanic crust behind the island arc (Fig. 1.10b). Third, when an overriding plate retreats by change of global plate motion, change of compression to extension of the oceanic

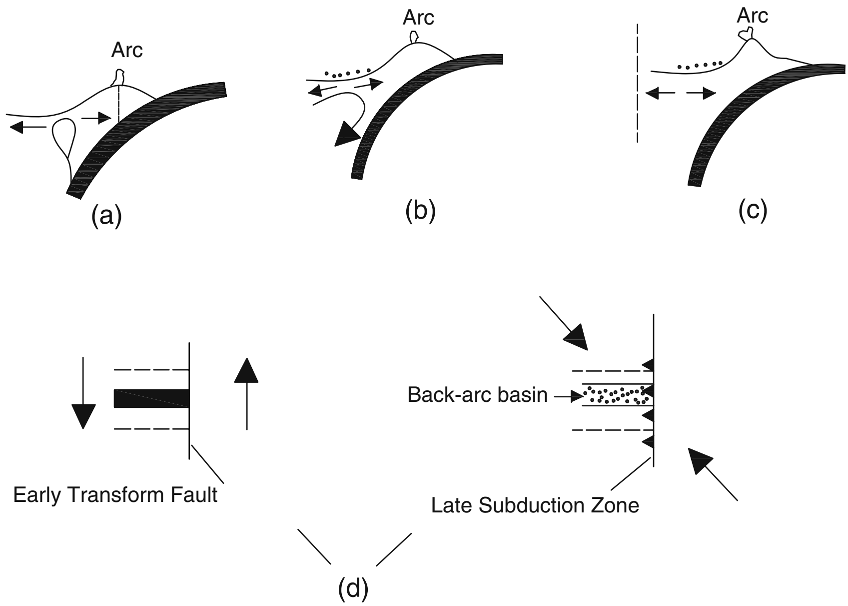

Fig. 1.10 Different mechanisms for the formation of Back-arc basin (after Moores and Twiss, 1995). (**a**) Forceful injection of a diapir from the subducting slab and extension of the overlying crust; (**b**) Convection currents driven by the drag of the downgoing slab resulting into extension of oceanic crust behind the island arc; (**c**) Retreat of overriding plate by global plate motion changes compression to extension of the oceanic crust behind the arc; (**d**) Back-arc basin formed by change of a Transform fault to a subduction zone

crust behind the arc gives rise to the formation of back-arc basin (Fig. 1.10c). Some basins may result from the entrapment of oceanic crust if a pre-existing fault zone becomes an oceanic subduction zone during a change of plate motion (Fig. 1.10d). The Aleutian basin may have formed in this way.

Some back-arc basins, familiarly known as the *marginal basins*, are formed due to rifting or secondary spreading of continental crust, resulting into Japan style marginal basin (Fig. 1.11). The resulting arc is the continental arc, which may be later added by magmatic material appearing as a melt material of the subducting

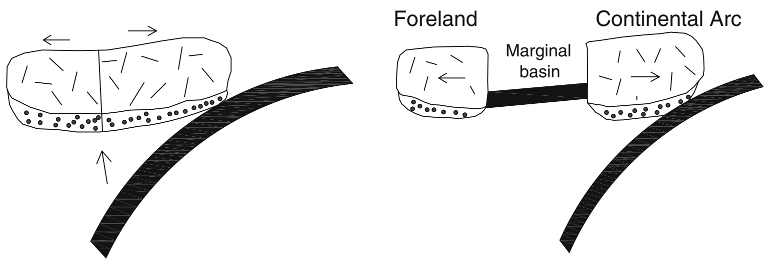

Fig. 1.11 Marginal Basin (Back-arc basin) formed due to secondary spreading (see text)

oceanic plate. The Japan Sea is a narrow ocean between the passive coast of Asia and the active volcanic arc of Japan. The continental arc formed by the secondary spreading is made up of the continental crust, which separated into an island arc and a continent near it. The supracrustal rocks in the marginal basin so formed are similar to those deposited in true geosyncline and are subsequently folded and metamorphosed by convergence of the two blocks that were separated earlier. Like the volcanic arc, the marginal basin is also an area of high heat flow.

1.5 Continental Rifting and Sedimentary Basins

James Hall of New York recognized long ago (1859) that mountains, as the most elevated parts of the Earth's crust, had risen by gigantic inversion of relief from the more depressed regions. Thus, it becomes obvious that a sedimentary basin or a trough (geosyncline) is the first requisite for the orogenesis. Some authors use the term rift for such basin and the process is called *rifting*. The term rift is applied to an elongated depression in the continental lithosphere because of applied extensional forces. Intracratonic basins are generally the consequence of extensional tectonics. In some cases, these extensional basins are asymmetric with an overall increase in thickness towards one of the bounding faults (Dunbar and Sawyer, 1988), indicating that the basin-bounded faults are active during sedimentation inducing subsidence and creating space for preservation of sediments. The basin-fill strata are also affected by intrabasinal gravity faults reflecting synsedimentary downward displacement that generate accommodation space for sediment deposition through out the history of basin evolution (Bally, 1980).

Even prior to the advent of plate tectonics it has been recognized that subsequent to the stabilization of the continental areas or cratons 2500 Ma ago, the Archaean crust was subjected to ductile stretching due to movement of mantle at depth and drag effects of sublithospheric convection currents (Bott, 1981, 1982). The crustal stretching or extension resulted in the formation of depressed regions or geosynclinal basins whose floor could be made of sialic crust or, if stretching was sufficiently large, of oceanic crust (Courtillot, 1982; Courtillot and Vink, 1983). In this depression, sediments start accumulating and eruption of magma results due to decompression melting of the ascending asthenosphere (Fig. 1.12).

Since the strength of the lithosphere under tension is least, the continental crust shows rifting associated with normal faults. The rift zone must be approximately normal to the regional tensile stress system. The rifts often contain volcanic rocks that are alkaline in the initial phase of rifting. The location of the rift is often controlled by pre-existing zones of crustal weaknesses (Virk et al., 1984). The rifting phenomenon occurs in a sequence of events in which first a normal fault develops in the upper crust under tensile stress (Fig. 1.13a). The fault movement warps the crust whose bending has the maximum tension at the location of greatest curvature. In this location, a second fault develops and a rift valley is thereby established (Fig. 1.13b). Continued tension gives rise to continuous subsidence of the keystone,

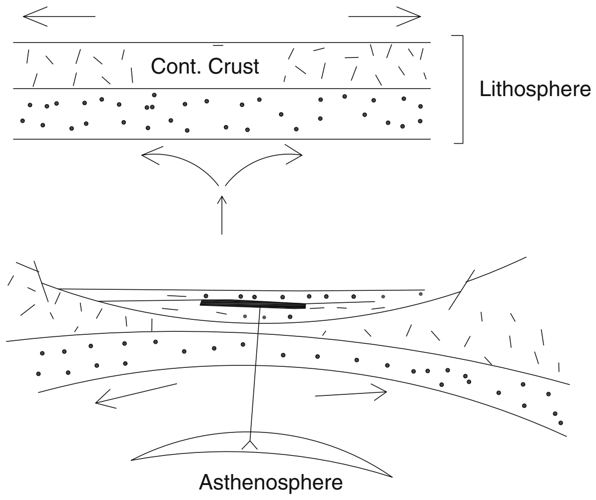

Fig. 1.12 Ductile stretching as a possible mechanism for rift basin formation (see text)

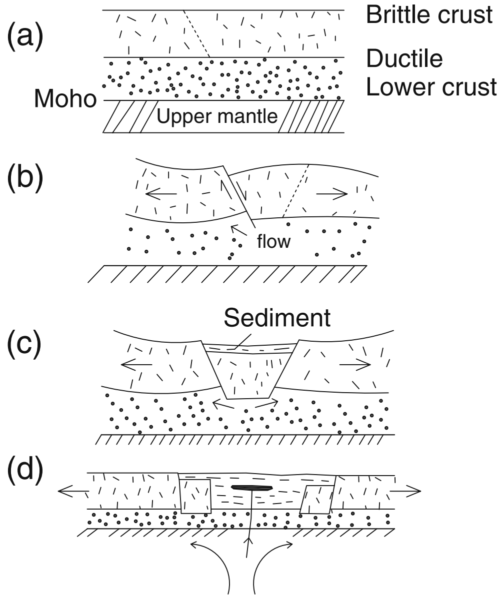

Fig. 1.13 Stages in the evolution of a Rift (see text for details)

aided by the sediment load. These movements are accompanied by compensatory flow in depth (Cronin, 1992; McKenzie and Morgan, 1969). Further tension separates the crustal blocks or lithospheric plates farther away and this horizontal separation of the crustal blocks indicates formation of true oceanic rift or trough. The floor of the rift drops below sea level. As the rift widens, sedimentation begins and

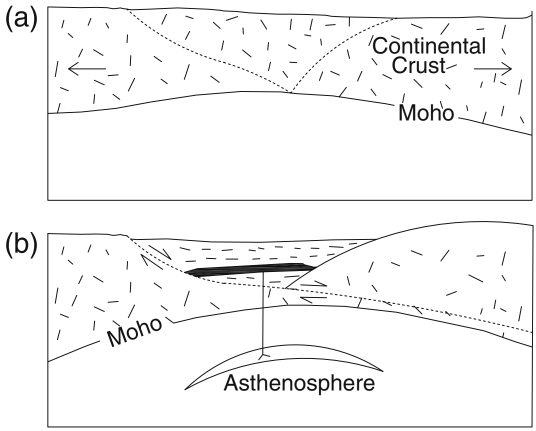

Fig. 1.14 Development of an Extensional basin in a continental lithosphere. (**a**) Rising convection currents generate tensional stresses, causing stretching of the crust; (**b**) deposition of sediments and basaltic lava in the extensional basin

marine conditions prevail. During the sedimentation emplacement of magma/lava also occurs due to melting of the asthenosphere.

The sedimentary basins, particularly in the intraplate environments, also develop by oceanward creep of the continental lower crust with a concomitant rift formation (Fig. 1.14).

1.5.1 Rifting and Triple Junctions

Rifting is commonly associated with domal uplift of the overlying crust because of impinging of mantle plume on the base of the lithosphere (Burke and Dewey, 1973). The plume is the consequence of some sort of thermal anomaly in the upper mantle, at the lithosphere-asthenosphere boundary. Any increase in temperature is likely to raise the boundary and consequently thinning of the lithosphere. Because of this activity and consequent doming of lithosphere, we have three-armed rifts originating from a triple junction, and also have basaltic magmas with the initiation of continental rifting, similar to the East African Rift, Gulf of Aden and the Red Sea that form a triple junction (Oxburgh and Turcotte, 1974; Patriat and Courtillot, 1984). If all three arms of the Rift continue to spread, the result is three separate continents and a mid-ocean Ridge-Ridge-Ridge (R-R-R) junction. On the Earth's surface that comprises of a mosaic of interlocking plates, there are several places where three plates are in contact and these are called triple junctions. The type of triple junction depends upon the various combinations of the basic types of plate boundary: divergent, convergent, and strike-slip. A convergent boundary can have either of two polarities, depending upon which is the overriding plate. A strike-slip boundary can have either of two shear sense; and a divergent boundary will have only one subduction direction. Thus, there are in all five different geometries for the plate boundary. Since three plate boundaries come together at a triple junction, so there are 125 combinations of 5 boundary geometries, taken three at a time (5×5×5). Of these, only 16 are kinematically possible, and 14 can actually exist for any geological significant length of time. These 14 types of junctions are called *stable triple junctions*. Stability of a triple junction does not mean that the location of the junction is fixed on the earth's surface or on a plate boundary (Cronin, 1992). Triple junctions between three ocean ridges, such as that in the South Atlantic between the African, South American and Antarctica plates, are known as ridge-ridge-ridge or RRR triple junction (Fig. 1.15a). A similar notation can be used to recognize triple junctions involving trenches (T), or Transform Faults (F). Therefore, a trench-trench-trench junction is denoted as TTT triple junction (Fig. 1.15c, d) while a ridge-ridge-transform fault junction would be termed RRF triple junction (Fig. 1.15e).

Figure 1.15 also shows the evolution of these three triple junctions from one time to a later time (see right-side figures). The geometric configuration of RRR (Fig. 1.15a) is always stable and the magnetic anomalies within the surface area created have Y-shaped patterns around the spreading ridges (Fig. 1.15b). The triple

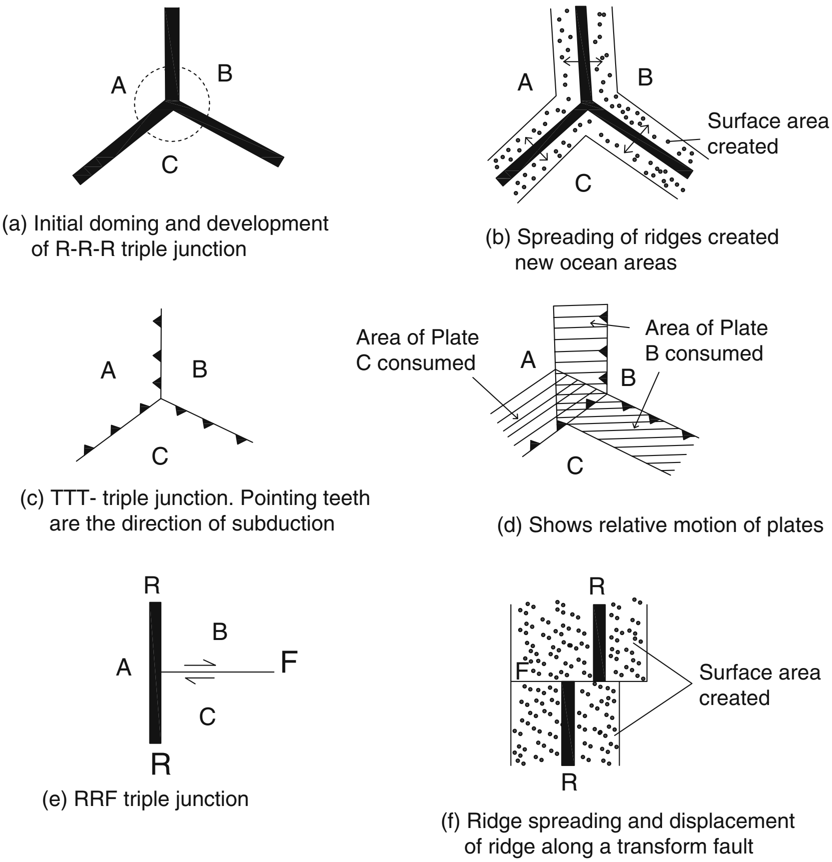

Fig. 1.15 Evolution of triple junction from one time to later time (after Kearey and Vine, 1996); see also text

junction between three trenches (TTT triple junction) is stable only when the relative motion of plates A and C is parallel to the plate boundary between B and C (Fig. 1.15c). That is, if the rates are the same and if the direction of subduction of plate C below plate A is exactly parallel to the boundary between plates B and C (Fig. 1.15d). When the relative motion of plate C is not parallel to the boundary between plates B and C, the triple junction is unstable. The triple junction RRF (Fig. 1.15e) is unstable because there is relative motion between plate B and plate C. Consequently, the RRF triple junction evolves at once to form an RFF junction (see Fig. 1.15f and compare with Fig. 1.8).. The RRR junctions are the only junctions that are always stable. Other triple junctions may be stable in certain restricted situations but are otherwise always unstable. Because RRR triple junctions are more stable than other forms, such triple junctions might still exist in old ocean floor, and can be recognized by bent magnetic anomalies, as in the Y shaped area of Fig. 1.15b.

The stability of the boundaries between plates is dependent upon their relative velocity vectors, as explained below (Fig. 1.16).

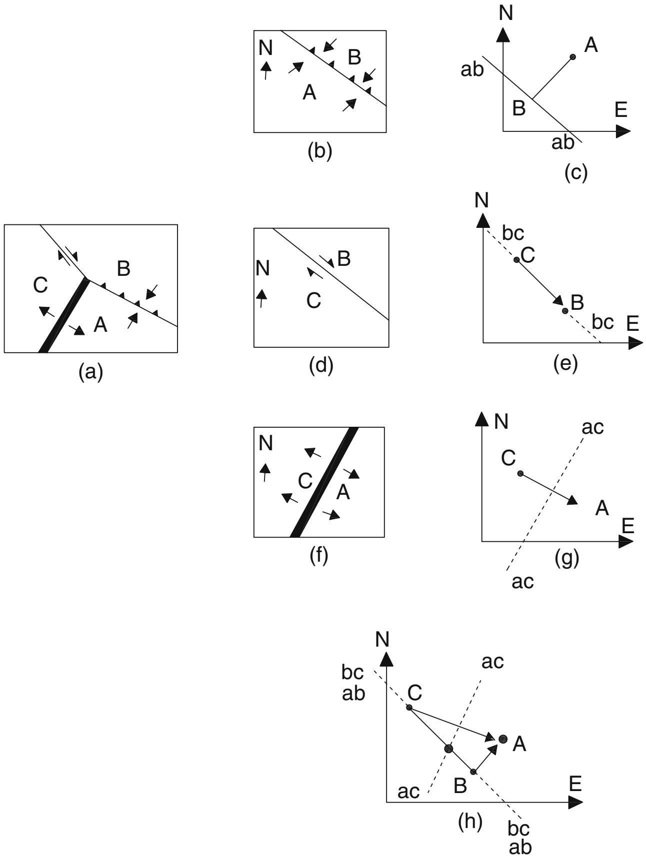

Fig. 1.16 RTF triple junction and velocity vectors as indicators of stability/instability of Plate boundaries (after Kearey and Vine, 1996); see text

When three plates come into contact at a triple junction the stability depends upon the relative directions of the velocity vectors of the plates in contact. Figure 1.16a shows a triple junction at ridge-trench-transform fault (R-T-F). In order to be stable, the triple junction must migrate up or down the three boundaries between the pairs of plates. To appreciate we ascertain this migration at each boundary in turn. In Fig. 1.16b, plate A is underthrusting plate B at a trench in NE direction. Now relative movement between A and B is shown in velocity space in Fig. 1.16c in which the velocity of any single point is represented by its N and E components. The line joining two points represents velocity vectors. Thus, the direction of line AB is the direction of relative movement between A and B, and its length is proportional to the magnitude of their relative velocity. Therefore, line *ab* must represent the locus of a stable triple junction, and B must lie on *ab* since there is no motion of the overriding plate B with respect the trench.

Next, we consider a transform boundary between plates B and C (Fig. 1.16d). Again, the relative movement between B and C is represented in velocity space (Fig. 1.16e). Between these two plates, the line BC is the velocity vector. The locus of a point travelling up and down the fault BC is in the same direction as vector BC, because the relative movement direction of B and C is along their boundary.

Finally, consider a ridge separating the plates A and C (Fig. 1.16f) and its representation in velocity space (Fig. 1.16 g). The relative velocity vector AC is now normal to the plate margin, and hence the line ac represents the locus of a point travelling the ridge. The ridge crest must pass through the mid-point of the velocity vector CA (provided the accretion is symmetrical).

By combining the velocity space representations, the stability of the triple junction can be determined from the relative position of the velocity lines representing the boundaries (Fig. 1.16 h). If they intersect at one point, it means that the triple junction is stable, because that point is able to travel up and down all three plate margins. If the velocity lines do not all intersect at a single point, the triple junction is unstable.

As stated already, only the RRR triple junction is stable for any orientation of the ridges. However, if one branch ceases to spread, it would form a failed arm, called *Aulacogen* (Fig. 1.17). Of the remaining two, one rift spreads more or less normal to the continental margin and the other spreads obliquely. Aulacogens are long-lived, deeply subsiding sedimentary troughs that extend at high angles to the fold belt (Burke, 1980; Hoffman et al., 1974). Since they occur from Late Proterozoic times

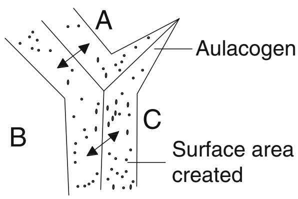

Fig. 1.17 Development of an Aulacogen from a RRR triple junction; see text for details

onward, some geologists have invoked their presence at places where they do not seem to be. Aulacogens are located at re-enterants on continental plate margins, and their initial formation is cotemporaneous with continental rupture. They are characterized by vertical tectonics. The sedimentary fill is about three times thicker than on the adjacent craton. The sediments are undeformed or weakly deformed, in contrast to the extreme tectonism experienced by adjacent orogenic fold belts. The aulacogens have a tendency to reactivation and often contain intrusion of alkaline igneous rocks. Aulacogens provide a complete igneous, sedimentary, and structural history of events associated with an orogenic belt. One should be therefore cautious to declare a rift within a continental plate as aulacogen (cf. Klein and Hsui, 1987; Beaumont and Tankard, 1987). The aulacogens are a very favourable location for the development of a river system, which carries detritus from the craton.

In India, the triple junction is located at 15 degree south latitude and 75 east longitude in the central Indian Ocean. From this junction a branch of the rift system, the Carlsberg ridge extends NW into Gulf of Aden to the Afar RRR triple junction, from which extend the Red Sea and the East African rifts. The second is the SE Indian Ridge that goes to south of Australia. The third is SW Indian ridge which extends in a series of short ridge and long transform fault segments around the southern tip of Africa into the South Atlantic Ocean to the Bouvet ridge-transform-transform triple junction (cf. Moores and Twiss, 1995).

For identification of the triple junction on or near the Indian shield, one has to consider criteria such as age, orientation and igneous activities in the three-armed rifts that are evolved by domal uplift of the crust, as described earlier.

It must be noted that rifting in the brittle zone of the crust, giving rise to *graben* (and associated horst), like the Rhine Graben or Godavari graben, is limited only to a small width (Bott, 1964, 1971). The amount of extension achieved by displacement on a steep normal fault is limited by depth to which crustal block can sink as it is compensated by lateral flow in the ductile lower crust. If the subsiding fault-wedge is bounded by vertical faults, the subsidence is lower than if the fault wedge is bounded by inclined faults. This is easily understandable if one considers that a wedge-shaped block of wood would float lower in water than a block of the same thickness with parallel edges.

1.5.2 Pull-Apart Basin

Typical Pull-apart basins occur at different scales and they form in the sedimentary cover above strike-slip fault in the basement. The most popular mechanism is local extension between two en echelon basement strike-slip fault segments. These can be right-stepping with dextral shear or left-stepping with sinistral shear. The local extension is accommodated in the sedimentary cover by normal faults at the releasing oversteps and bends. In other words, when a strike-slip fault has notable curvature, the curved area, separating the ends of the faults, is thrown into tension

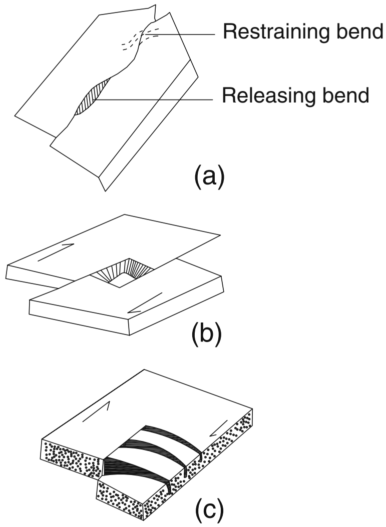

Fig. 1.18 Pull-apart basins due to strike-slip motion and extension. (**a**) Basin at a fault curvature during transtensional regime, (**b**) sharp pull-apart basin by strike-slip motion, (**c**) basins formed as a result of fault termination

or compression. Compression gives rise to an elevated region by crustal shortening, typified by folds and thrust faulting, while tension gives rise to an extensional trough, known as a pull-apart basin (Fig. 1.18). The strike-slip margins are initially straight and parallel, but may sag with time (Fig. 1.18a, b). A similar basin would also develop when one fault terminates and sidesteps to an adjacent parallel fault (Fig. 1.18c). In a pull-apart basin, the faults should be steep near the surface and should show evidence of strike-slip movement. Pull-apart basins progressively grow in the same direction as the fault movement so that the oldest sediments occupy the margins only (Fig. 1.19). As the pull-apart basin grows, its floor is stretched and attenuated whereby igneous material is emplaced in the centre of the basin (revealed by high gravity). However, this may not happen in all circumstance, and some basins may not have been attenuated to an extent to cause rupture of the lithosphere and igneous rocks are completely absent. The pull-apart basin can have any shape. Pull-apart basins are excellent target for hydrocarbons.

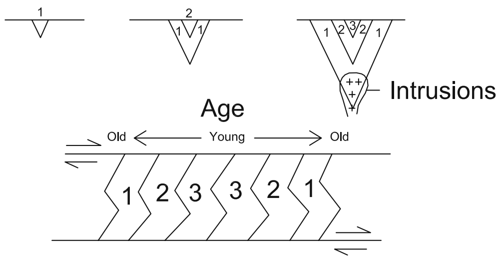

Fig. 1.19 Stages in the formation of Pull-apart Basins (see text for details)

1.5.3 Foreland Basin

It is a flexure depression peripheral to the mountain range (Fig. 1.20). An example of this is the development of the Indo-Gangetic plain basin adjacent to the Himalaya. This foreland basin extends much further into the surrounding craton and forms in advance of the thrust front of the orogenic belt. Close to the mountain range, the sediments are coarse-grained and deposited in a shallow water or continental environment. Farther away the sediments are fine-grained and often turbidite. The sediments thus form a wedge-shaped unit whose stratigraphy reflects the subsidence history of the basin as it grows and migrates outwards as convergence continues. The stratigraphy shows progressive overlap of sediments onto the foreland. This means that the stratigraphy is characterized by units, which thin laterally, overlap older members, or are even truncated by erosion.

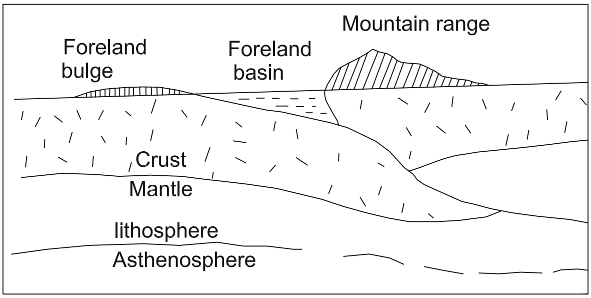

Fig. 1.20 Foreland basin formation as a flexural depression of the continental crust

1.6 Orogeny and Plate Tectonics

Careful study of fold belts has revealed that the period of orogenesis is generally quite long, often exceeding 100 million years. Orogenic belts or fold belts usually consist of roughly parallel ridges of folded and thrusted sedimentary and volcanic rocks, portions of which have been strongly metamorphosed and intruded by somewhat younger igneous bodies. Initiation of the orogenesis occurs with the development of a depositional trough or geosyncline because of rifting or ductile stretching of a pre-existing continental crust or lithosphere. The rifted basin becomes a wide depositional site by further stretching of the continental crust due to divergent subcrustal convection currents. The floor of this rifted basin sinks below sea level and is flooded. The separating margins of the rift thus become the coasts of a growing intercontinental sea. The floor of such oceans may be made of stretched continental crust or, if crustal extension was extreme, the floor may be of oceanic crust. This basin receives sediments, which are a vast accumulation of deep-water marine deposits that may exceed a few thousand meters in thickness as well as thinner shallow water deposits. The deep-water sediments deposited in eugeosyncline (seaward

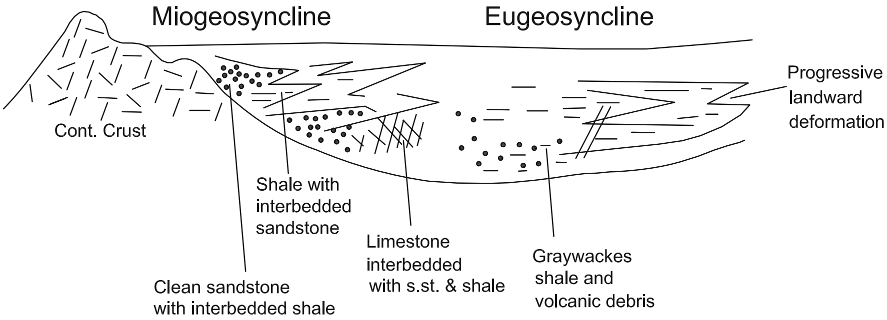

Fig. 1.21 Different sedimentary deposits in shelf, slope and deep-sea conditions within a geosyncline

of the miogeosyncline) consist primarily of graywackes, volcanic debris, and shale. The shallow water deposits of the miogeosyncline and those on the shelf region are toward the continental side. They consist of relatively clean sandstones, limestones, and shale (Fig. 1.21).

The plate tectonic cycle begins at the divergent boundaries along ocean ridges, causing the two lithospheric plates (overridden by continental plates) to move away and converge elsewhere with another lithospheric plate. With the seafloor spreading, the lithosphere is transported laterally, away from the ridge toward a convergent plate boundary, i.e. toward a trench. The migrating plates carry the sediments deposited on their margin and the deformation generally progresses in a landward direction so that the deep-water sediments are the first to be deformed. At the convergent plate boundary the oceanic lithosphere is bent and it moves down into the Earth's mantle, producing an ocean trench at the surface. The descending plate beneath the ocean trench is heated by friction and other heat sources at depth and causes partial melting of the down-going plate with some overlaid sediments. Water and other volatiles help in the melt generation and the melt rises to the surface, forming volcanic arc, as volcanoes parallel with the ocean trench. The oceanic trench becomes a trap for the ocean sediments and for additional sediments derived from the adjacent island arc.

During the plate movement, the leading edge of the overriding plate bulldozes material from the ocean floor into an accretionary wedge. Some sediments are scrapped into the wedge with time. The accretionary wedge grows with time and may include other materials such as sea mounts or islands scrapped from the sea floor.

Where two plates converge, and the leading edge of one plate is made of oceanic and the other of continental block, the intervening oceanic crust (being denser) of the oceanic plate is subducted or thrusted below the continental plate; a very small quantum of sediments is carried in the subduction. The subducted oceanic crust (with minor trapped sediments) undergoes melting at depth. The melt rises near or on the surface, resulting in the formation of volcanic arc or *magmatic arc* (commonly andesitic in composition). The rocks of the arc at depth crystallize into granitoids

(I-type). The magmatic arc is often located a few hundred kilometers seaward of the ancient coastline. The igneous rocks of the arc are eroded and deposited in the basins on either side of the arc and are called *fore-arc basin* (on the trench side) and *back-arc basin* on the other side of it.

With continued subduction the intervening oceanic crust disappears, and finally the colliding plates become continent versus continent. Because of the plate convergence, the sediments and the interbedded lava on the ocean floor are folded and even thrusted to increase in thickness and eventually affected by higher geothermal gradients. The complexly folded rocks at some stage of their deformation are recrystallized by heat brought upwards by hot fluids. These fluids are in fact liberated from subducted material or from the magmas generated below. The influx of heat is also notably due to heat production by radioactive elements in the sedimentary pile. Additionally, heat contribution is also due to transformation of mechanical work (used in rock deformation) into thermal energy. This temperature increment is due to simple compressional squeezing of rocks at depth due to weight of overlying material. A further increase in temperature can also be caused by friction, generated during lithospheric sliding into the mantle. The thickened crustal rocks undergo thermal relaxation and consequently a series of progressive metamorphic reactions take place in the rocks with increased depth or pressure and temperature. Amongst these, some critical reactions give rise to *metamorphic isograds,* which are distinguishable in the exposures by the appearance or disappearance of an index mineral or a specific mineral paragenesis. The isograds separate the metamorphic rocks into zones or facies or grades (Fig. 1.22). The series of metamorphic assemblages developed in different metamorphic zones are a reflection of T-depth profile at the time of metamorphism. The T-depth profile implies that the types of metamorphism, Barrovian or Buchan/Abukuma, are the result of different tectonic conditions. High T,

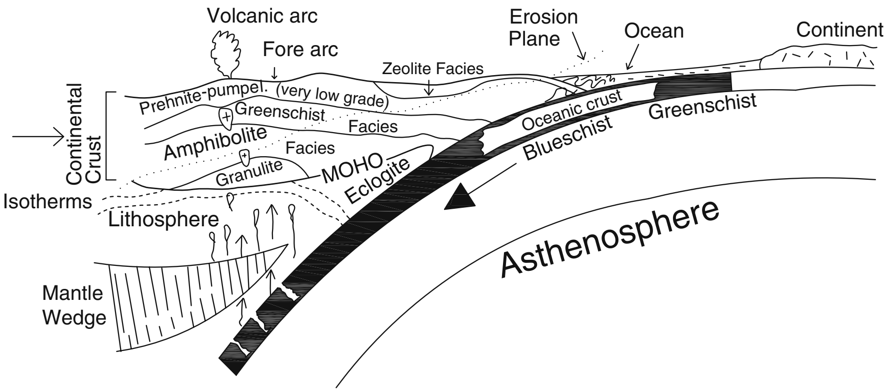

Fig. 1.22 Schematic cross section of converging ocean-continent plates, showing island arc and the distribution of metamorphic facies. High P/T facies series develops in the subduction zone while low P/T facies series is formed in the arc. *Dotted line* is trace of an erosion plane along which a progressive sequence of regional metamorphism is encountered in eroded fold belts

low P metamorphism (Buchan type) implies that temperatures are elevated above a normal geothermal gradient of 25°C/km. The condition develops in contact aureoles around shallow igneous intrusions in a volcanic arc environment. On the other hand, high P, low T metamorphism generates blueschists, which implies that the temperatures of the rocks are significantly lower than would be in a normal geotherm. This occurs in subduction zones where cold rocks are carried to large depths faster than temperatures can be re-established. Such metamorphism is rare in most orogenic belts. Where present, it is overprinted by Barrovian metamorphism. Adjacent high T-low P metamorphism (Buchan type) and blueschists (high P–low T) metamorphism constitute a paired metamorphic belt. Barrovian metamorphism indicates a normal geothermal gradient and is widespread in all orogenic belts. For this reason it is also known as classical regional metamorphism. The Alps provide a good example of the Barrovian metamorphism. In the Himalaya, however, the metamorphism is inverted such that higher grade rocks occur in higher tectonic levels. Such an occurrence could indicate a primary inversion of isotherms, as would develop at a subduction zone above a down-going slab. Alternatively, it could indicate a tectonic inversion of the isotherms following metamorphism, like that on the inverted limb of a recumbent fold. This problem will be discussed in detail at the appropriate place when the evolution of the Himalaya is considered in Chap. 3. In areas of good exposure, we can obtain an idea of the shape of isograd surfaces in 3-D. Where they are undeformed or only weakly deformed, they appear to be gently curved surfaces that intersect the Earth's surface at small angles. The interpretation of metamorphic zones in terms of plate tectonics is complex. However, the rocks in the central portion of all orogenic belts are metamorphic.

The distribution of metamorphic zones in many deformed belts is roughly symmetrical, such that the highest grade rocks constitute the central portion of the fold belt, which is a part of the mountain root at depth, and less metamorphosed or unmetamorphosed rocks occur on the flanks. In the mountain building processes by plate tectonics, the different domains have a relationship which is typical for a collision fold belt, whether the collision occurred between an oceanic plate and a continental plate or between a continent and another continent. The metamorphic isograds or facies boundaries show trends similar to that held by the subduction zone (i.e. trench). A similar orientation is also exhibited by granitic rocks of the magmatic arc (Fig. 1.23). High-grade metamorphism may be symmetrical with respect to the arc, but uneven erosion shows regionally metamorphic rocks in a progressive manner in one direction only. The sediments lying on the continental margins may be intricately folded and are often thrust upon by metamorphic rocks of the fold belt.

In the orogenic belts, roots are regions of greater crustal thickness that underlie almost all continental orogenic belts in the world. Two hypotheses have been proposed to explain the formation of mountain root (cf. Wilson and Burke, 1972). One is related to subduction processes and the other to collision processes. According to the subduction model, the mountain roots form beneath an active continental margin by the rise of mafic magmas away from the subducting slab and intrusion of the magmas into the lower crust. These magma bodies heat the crustal rocks in the vicinity and thus form the mobile core. The thickened continental crust is raised due

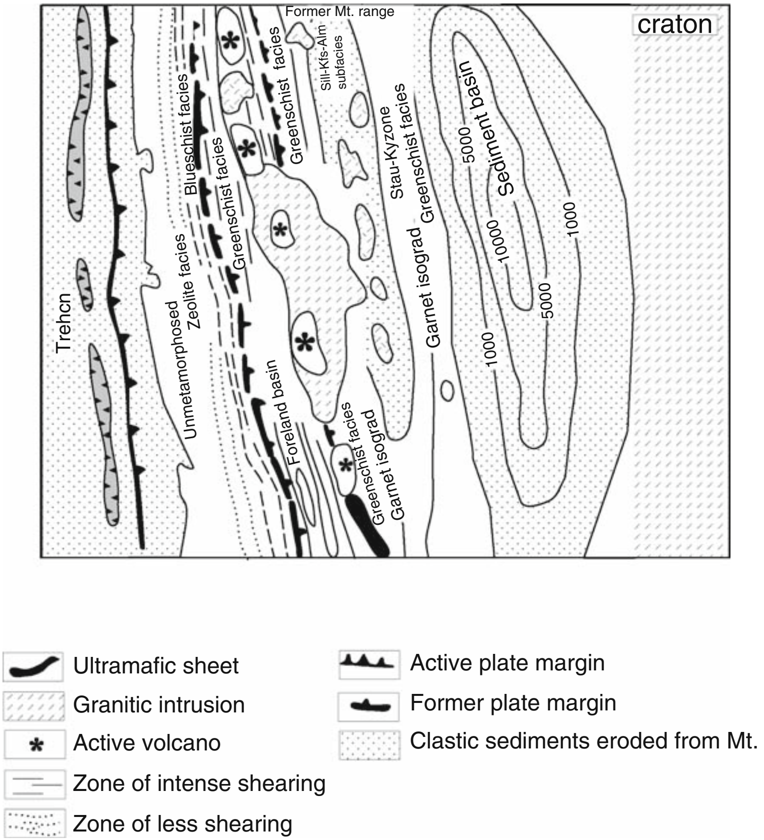

Fig. 1.23 Schematic map showing the development and location of plate tectonic elements and the distribution of metamorphic facies at a convergent continent-ocean plate boundary

to isostasy. The high topography of the orogenic belts exists because of the isostatic support given by the thickened crust "floating" in high-density mantle. Gravitational collapse of the thickened crust results in thrusting of the recrystallized rocks on the foreland. The second model explaining the formation of the mountain roots is the collisional model. According to this model, the root is formed by underthrusting of one continent by another. This causes crustal thickening and hence elevated topography of the mountain belt. The collision model seems more plausible for the formation of roots and high topograpahy of the orogenic belts. In young mountains, like the Himalaya, thickening of mountain root approximately reflects the amount of shortening that has taken place, but thickening should occur rapidly otherwise erosion and ductile collapse will hinder the formation of the root and high topography of the orogenic belt.

It is possible to map the isograds and successive structures, such as s-surfaces (S1, S2, S3 etc.), lineations (L1, L2, L3 etc.) and fold axes (F1, F2, F3 etc.), and relate them temporally with the metamorphic events of minerals (syn-, late-, or post-tectonic with respect to deformation phases D1, D2, D3 etc.). This may help in unraveling the geohistory of the fold belt. At certain locales, the rocks subjected to regional metamorphism also undergo anatexis (partial melting). This process gives rise to migmatites, granulites, and granitic plutons. The granitic plutons (S-type) are the products of accumulation of anatectic melt that intrudes the overlying rocks. The isostasy and the buoyancy of the anatectic magma and the heated metamorphic rocks are exhumed to form lofty mountain ranges or fold belts. During uplift of the orogen, some parts of the subducted oceanic crust may also be obducted as high-pressure rocks called *blueschists* and eclogites. The orogenic uplift is often accompanied by the formation of down-sliding nappes. With the elevation of orogenic belt, unconsolidated marine sediments begin to slide downwards along gentle inclines and fall off as *turbidite currents* into the marginal deeps where they often build thick chaotic *flysch* deposits. The weathered debris transported by streams is deposited as *molasses* into a marginal deep. It must also be noted that the sediments deposited on continental shelf are also complexly folded and may even be displaced inland along low-angle thrust faults and the sedimentary portion of the mountain chain.

1.7 The Wilson Cycle

As discussed in the previous section, the orogenic belts are characteristically formed of thick shallow-water sediments deposited on continental crust and of deep marine sediments that are subjected to deformation and metamorphism during plate tectonic operations. It is postulated that Proterozoic fold belts also originated by the opening and closing of the geosynclinal basin. The periodicity of ocean formation and closure is known as the *Wilson Cycle*, named after J. Tuzo Wilson, by Dewey and Burke (1973) in recognition to his contributions to the theory of plate tectonics. The Wilson cycle gave us a way of understanding mountain building processes. Therefore, we must have a good knowledge of this concept which is a necessary background for understanding the evolution of fold belts of younger and older ages. The Wilson cycle suggests a pattern for the developments of orogenic belts and the cycle is assumed to have operated through much of geological time. The Wilson cycle has three stages. (1) Separation of the continents and development of an ocean basin (geosyncline) with accumulation in separate areas of thick deposits of shallow marine sediments (miogeosynclinal) and deep-water marine sediments (eugeosynclinal); the latter with mafic igneous rocks of basaltic to andesitic compositions. (2) Movement of the continents towards each other, causing deformation of the eugeosynclinal and miogeosynclinal rocks together with emplacement of ophiolites from depth. (3) Continental collision attendant at some stage with intrusion of granite batholiths in the core region. The generation of most magma can be related to one of the three stages in the Wilson cycle. During separation of the continents, the principal igneous activity includes emplacements of tholeiitic lava flows, dykes and sills as well as alkaline intrusions—kimberlites and carbonatites—on the continents along a relatively wide zone parallel to the line of separation of the continents. Extrusion of tholeiitic basalts, and sometimes alkaline volcanics, occurs in the ocean basins. Again, extrusion of andesitic lavas occurs in island arcs along the leading edge of the continent adjacent to subduction zones. During the period when the continents move toward each other, the subduction occurs on either one or on both continental margins that are approaching. There is little magmatic activity along the continental margins devoid of subduction. Along the leading edge of the continents there are both extrusions and intrusions of granitic composition. When continents collide, the principal igneous activity is the intrusion of granitic rocks along the fold belts formed by collision. Sometimes extrusion of rhyolites also occurs along the belt. Following the collision of the continents, this process is repeated because the continents are not consumed due to buoyancy.

The Wilson cycle can include many events leading to orogeny, and one example of the events, adapted from Dewey and Bird (1970), is described below (Fig. 1.24).

The beginning of the Wilson cycle starts with the initial rupture of a continent and development of an accreting plate margin (Fig. 1.24a). As separation of continents proceeds, there develops a rift valley (Fig. 1.24b) and small ocean (Fig. 1.24c). These are the early stages of opening of the ocean. The African Rifts and Red Sea are the present day examples. With the separation of the continents, continent-derived sediments accumulate at the continental shelves at the ocean-continent interfaces. The Atlantic-type ocean is fully developed with a central ridge (Fig. 1.24d). It may be noted that the two plates are drifting away from the ocean ridge but they are not get separated because the upwelling magma is added to the trailing edge of the plate. In the following stage, the oceanic lithosphere uncouples and descends along a trench into the mesosphere. Figure 1.24e represents a consumption stage of oceanic plate margin and is the initial stage of ocean closing in the Wilson cycle. If the uncoupling takes place at a considerable distance oceanward from the continental margin, an island arc, like Japan, forms. If the uncoupling and subduction of the oceanic lithosphere or oceanic plate occurs adjacent to the continental margin, it develops Andean-type orogen because of tectonic and volcanic processes (Fig. 1.24f). As the ocean continues to contract, the opposing continental margin ultimately arrives into the consumptive regime. This is the early stage of continent-continent collision (Fig. 1.24 g). Because of buoyancy constraints of the continental lithosphere, plate consumption ceases in the later stages of continent-continent collision (Fig. 1.24 h). The lithospheric plate detaches beneath the Himalayan-type orogen, and the final-stage of the Wilson cycle is marked now by the fully evolved mountain belt as the place of suture between the newly joined continental masses.

Strictly speaking, the Wilson cycle is not a cycle because there is no evidence of exactly the same sequence of events repeating itself in the same region. In addition, whenever two continents are moving apart, they are moving toward each other on the opposite side of the Earth or toward another continent. The movements of the continents in the Wilson cycle are taken relative to a Pangaea-like landmass that appears to have formed repeatedly in geologic time because of the joining of the continents. Moreover, there are instances where the orogenesis involves only

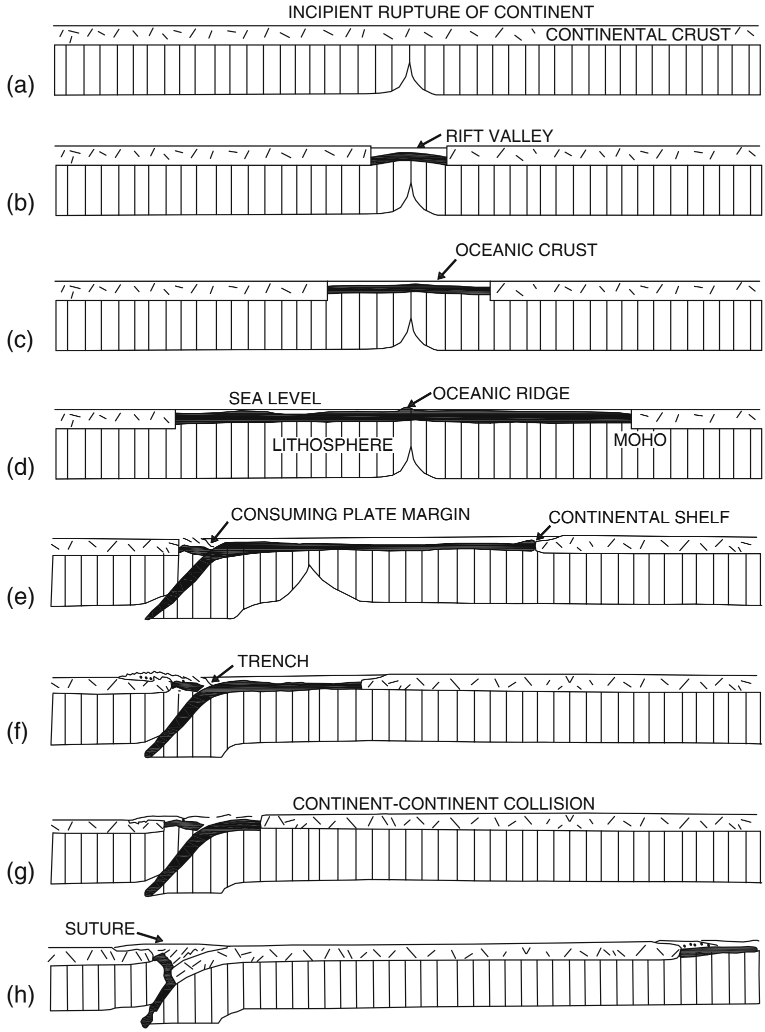

Fig. 1.24 Schematic illustration of the Wilson cycle (after Dewey and Bird, 1970)

intracratonic rifting due to ductile stretching and thinning, and the depositional basin having sialic/continental basement, and not oceanic crust. The basin becomes wider due to subcrustal convection currents and the ongoing sedimentation is intervened by volcanic lava flows, resulting from decompression melting at depth. Accumulation of the supracrustal material results in sinking of the basin. With deepening of the basin, decay of the underlying plume with time, and a change in the vertical component of stress, the diverging continental blocks are likely to reverse their movement direction to cause shortening of the crustal segment (Sharma, 2003). Consequently, the deformed and metamorphosed basinal rocks would give rise to fold belt or orogenic belt along which the crustal blocks are welded or sutured. Structures in these accreted crustal blocks are virtually unaffected (albeit overprinted) by the younger mobile belt. In contrast, the continental blocks welded together by plate tectonic processes show markedly different stratigraphy or tectonic history and have a discontinuity in the orientation or style of structures.

1.8 Suspect Terrains

During the evolution of orogenic belts, the plate tectonics motions not only result into collision of the continent(s) or arc(s) but at times some orogenic belts include terrain(s) that have distinct geology and structures, without showing any relationship to the subduction zone. That is, the relationship to the main crustal blocks is suspect. These terrains are called *Suspect* or Exotic *terrains*. They are allochthonous in the term of tectonics, and they have accreted to the continent over a considerable period, after travelling great distances. The process of amalgamation or welding has been referred to as accretionary tectonics or mosaic tectonics.

A terrain is an area surrounded by sutures and characterized by rocks having a stratigraphy, petrology, or palaeo-latitude that is different from that of the neighbouring terrains or continents. In a collisional fold belt, a terrain is a remnant of crust that has had a history different from that of the subducted oceanic crust or that of the main crustal block. Those who doubt a suspect terrain must work out detailed geological histories of the terrains, especially on the time of docking of two terrains. A suspect terrain is a mappable unit and Jones et al. (1983) have given several criteria to distinguish the identity of separate terrains. These are:

1. The stratigraphy and sedimentary history
2. Petrogenetic affinity and magmatic history
3. The nature, history and style of deformation
4. Palaeontology and palaeo-environments
5. Palaeopole position and palaeodiclination.

Based on these criteria, Jones et al. (loc. cit.) recognized four major types of suspect terrains in western North America. These are:

I. Stratigraphic terrains which are characterized by distinct stratigraphies.
II. Disrupted terrains that contain a heterogeneous assembly of flysch, serpentinites, shallow water limestones, and graywackes with occasional exotic blocks of blueschists.
III. Metamorphic terrains, in which a metamorphic overprint has destroyed the original stratigraphy.
IV. Composite terrains, which contain two or more such terrains that had amalgamated before accretion to the continent.

The chronological sequence of accretion of terrains to the continent can be known from geological events that indicate accretion and link adjacent terrains. These include the deposition of sediments across terrain boundaries; the appearance of sediments derived from an adjacent terrain; and "stitching" together of terrains by plutonic activity. Again, the Apparent Polar wandering paths (APWP) of two separate terrains should be different until their accretion, after which they should exhibit a single APW path.

Considering the diversity of terrains, especially the terrain of fold belt itself and the flanking continental blocks, it seems that we may become over suspicious on the problem of suspect terrains. However, the recognition of suspect terrains in fold belts requires a major modification of the Wilson cycle and of our ideas about how the orogenic belts develop. Moores and Twiss (1995) state that "the motion of major crustal blocks in a collision may represent only a part of the activity that constructs an orogenic belt, and many other relative plate motions may be represented by the numerous sutures among different exotic terranes".

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