3

Landforms and tectonics of plate margins

3.1 Convergent plate margins: general characteristics

The fundamental types of landform created at convergent plate margins are orogens and island arcs. Two distinct mechanisms are responsible for their development into significant morphological features. One involves collision and occurs where the motion of two plates towards each other brings into contact two continents, a continent and an island arc or two island arcs. The other mechanism involves the effects of heating associated with plate subduction. Such consumption of lithosphere can take the form of either subduction of oceanic lithosphere beneath a plate capped by continental crust or the subduction of oceanic lithosphere beneath a second oceanic plate.

The subduction of oceanic lithosphere is a steady-state process in that it can continue until a continent or island arc arrives at the subduction zone. The west coast of South America provides an example of oceanic lithosphere subduction (Nazca Plate) beneath continental lithosphere. The distinction between steady-state subduction and plate collision at convergent boundaries is due to the resistance of continental crust and island arcs to subduction, although it is possible that the latter can be subducted. The collision of two continents leads to the eventual cessation of subduction, whereas other types of collision may cause a change in the location of the subduction zone. The main types of convergent plate margin and their possible modes of development are summarized in Table 3.1. They are classified into steady-state and collision margins and then further subdivided according to the nature of the crust on the interacting plates, the mode of plate interaction and the type of orogen or island arc formed. Also listed are possible courses of subsequent development for each margin subtype.

3.1.1 Steady-state margins

Figure 3.1 illustrates the two types of **steady-state margin**. Where oceanic lithosphere is subducted beneath another oceanic part of a plate the associated volcanic activity and other thermal effects produce an **intra-oceanic island arc.**

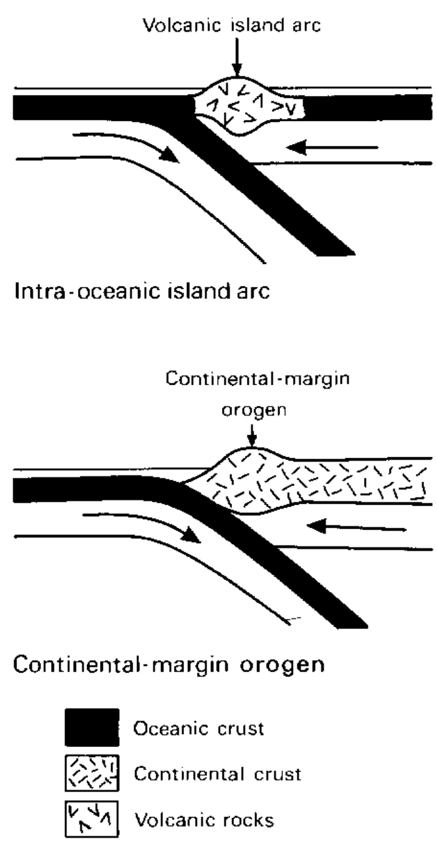

Fig. 3.1 Major elements of the two types of steady-state margin.

Table 3.1 Classification of convergent plate margins

MARGIN TYPE	STEADY STATE		COLLISION			
TYPE OF CRUST ON INTERACTING PLATES	Oceanic/oceanic (though may contain sliver of *continental* type crust)	Oceanic/continental	Continental/ continental	Oceanic (island arc type)/ continental		Oceanic (island arc type)/ *oceanic (island* arc type)
NATURE OF INTERACTION	Oceanic → oceanic	*Oceanic → continent*	*Continent* → continent	Island arc → *continent*	Continent → island arc	Island arc → island arc
TYPE OF OROGEN FORMED	Intra-oceanic island arc	Continental-margin orogen	Inter-continental collision orogen	*Modified* continental-margin orogen	Modified 'passive' continental-margin orogen (subduction halted)	Compound intra-oceanic island arc orogen
POSSIBLE EVOLUTION	Island arc → continent collision OR continent → island arc collision	Continent → continent collision OR island arc → continent collision	No further development as convergent boundary since subduction halted	Back-arc subduction halted but forearc subduction may continue	Subduction halted OR polarity reversal with new subduction zone in back arc	Subduction of 'underthrusting arc' continues OR polarity reversal of subduction behind 'overriding arc'

These are particularly extensively developed in the western Pacific Ocean. Their typical arcuate plan form is probably due to the relationship between the angle of subduction and the Earth's curvature. Some idea of the effect can be gained by depressing the surface of a table tennis ball with the thumb. Subduction of oceanic lithosphere beneath a plate carrying continental crust gives rise to a **continental-margin orogen.** The Andes provide an excellent example and this type of orogen is sometimes referred to as an Andean type or Cordilleran type. However, such an orogen may also form a **continental-margin island arc** if the continental crust behind the arc is below sea level. This is the case with the Sumatra–Java part of the Sunda Arc in the East Indies.

3.1.2 Collision margins

A **collision margin** can take a number of forms, depending on the characteristics of the interacting plate boundaries. The first type involves the collision of two continents which gives rise to an **intercontinental collision orogen** (Fig. 3.2), the most notable example of which are the Himalayas. This type must begin with a continental-margin orogen on the overriding plate and be transformed into an intercontinental collision orogen when continental crust on the underthrusting plate reaches the subduction zone. The changes experienced by the leading edge of the continent on the subducted plate are dramatic since there is a conversion from a 'passive' continental margin (see Chapter 4) to an active plate boundary. Eventually the subduction zone is converted into a **suture zone** as the two continental blocks are welded together.

A second type of collision involves the migration of an intra-oceanic island arc towards a subduction zone bounded by continental crust (Fig. 3.3). Unless the island arc is subducted it will be accreted to the continental-margin orogen previously formed along the edge of the continent. Subduction along the continental margin would eventually cease, but it would probably continue along the outer edge of the island arc as oceanic lithosphere would still be available for subduction there. The resulting orogen would be a modified continental margin type.

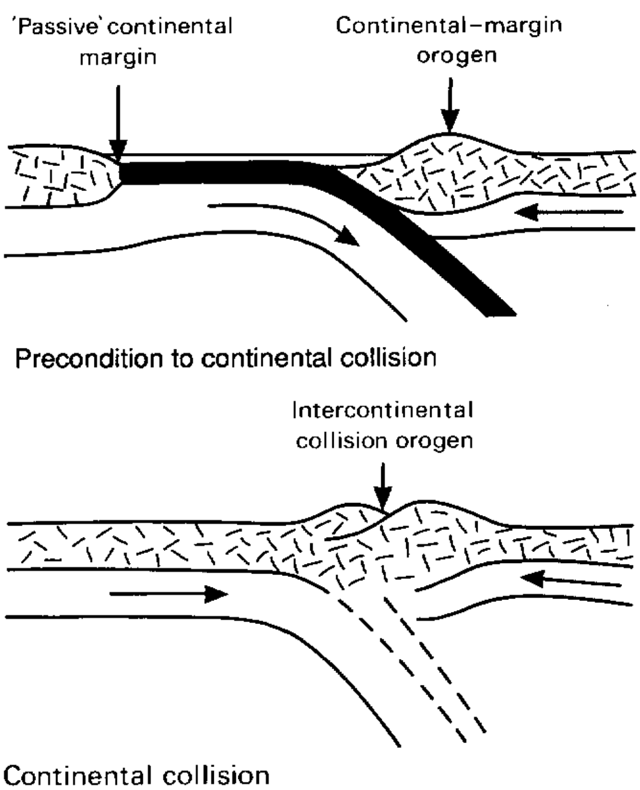

Fig. 3.2 Development of an intercontinental collision orogen (key as for Figure 3.1.)

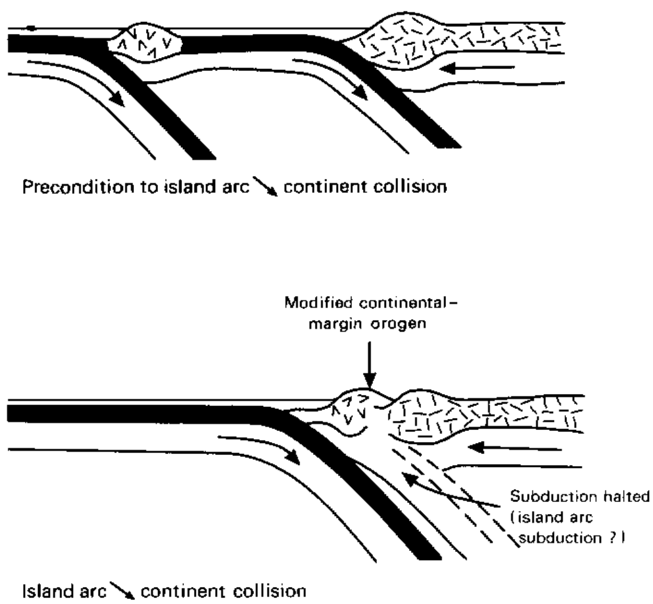

Figure 3.3 Possible consequence of an intra-oceanic island arc moving towards a subduction zone adjacent to a continent (key as for Figure 3.1).

Where a continent moves towards a subduction zone associated with an intra-oceanic island arc the consequences are rather different because of the resistance of continental crust to significant subduction. The interaction can be accommodated in two ways (Fig. 3.4). A possible example of such a continent – island arc collision is provided by northern New Guinea. The collision of intra-oceanic arcs can also take two forms (Fig. 3.5). Some of these plate boundary configurations and stages of development are largely conjectural since clear examples do not exist at the present day. There is, for example, uncertainty as to whether island arcs can be subducted and whether the direction of subduction can be reversed, as illustrated in Figure 3.4.

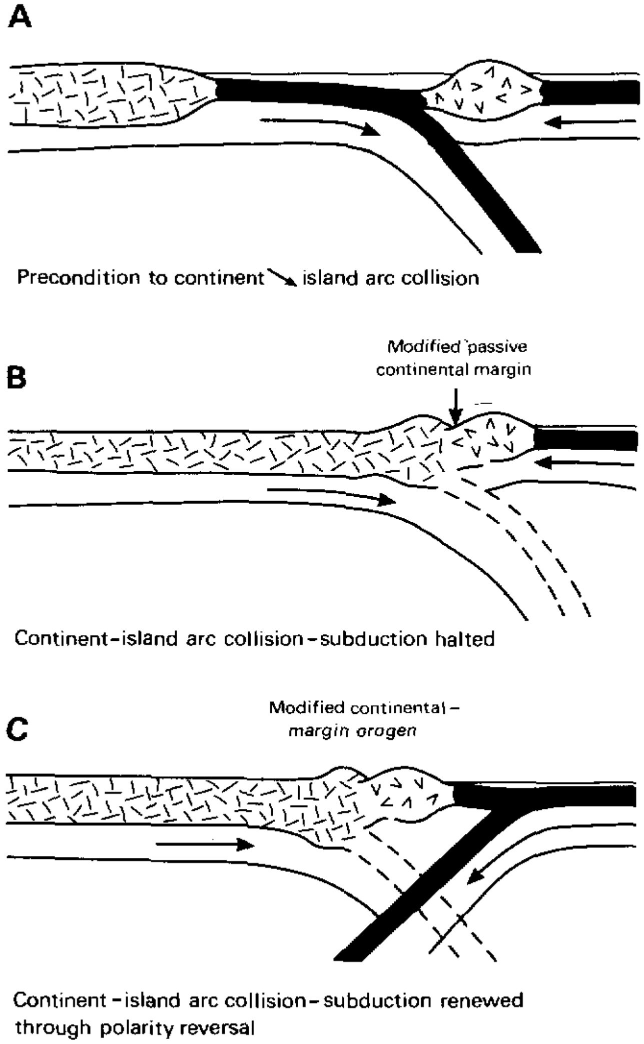

Fig. 3.4 *Possible development of a convergent margin where continental crust reaches a subduction zone adjacent to an intra-oceanic island arc. One possibility is that the collision between the island arc (on the overriding plate) and the continental margin (on he subducting plate) will eventually stop subduction (B). An alternative mode of development which maintains the relative opposing motion between the two plates is a reversal of the polarity of subduction, a process sometimes called 'flipping' (C). The original subduction zone is closed off but a new one forms behind the advancing island arc. Both processes would give rise to a modified continental-margin orogen. In (B) a 'passive' continental margin would gradually be created since subduction would eventually stop. In (C) an active modified continental-margin orogen would be formed with an accreted island arc along its edge and a new subduction zone beyond the island arc (key as for Figure 3.1).*

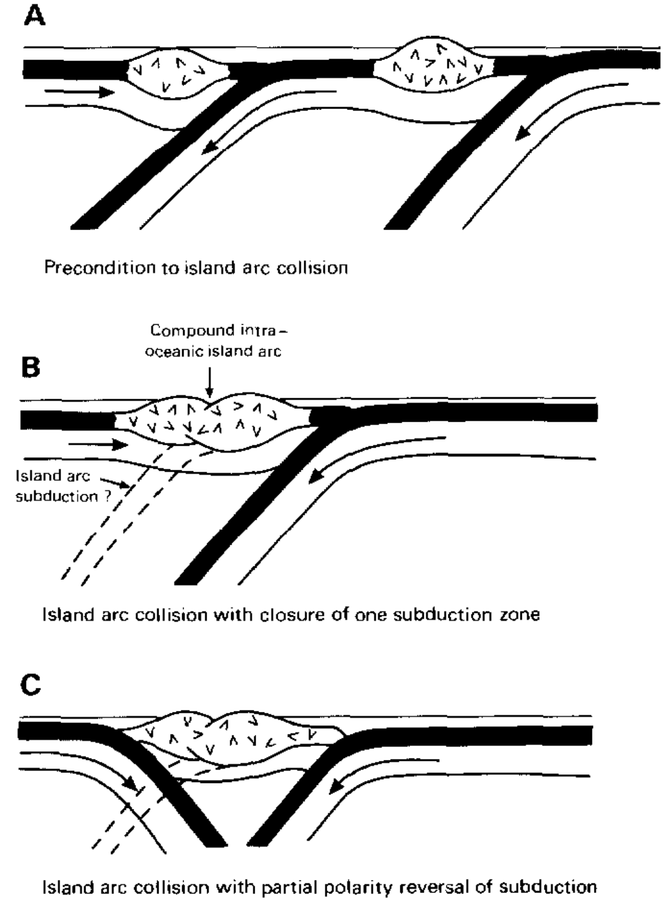

Fig. 3.5 *Possible evolutionary sequences following the collision of two intra-oceanic island arcs. When the island arcs collide, assuming that one is not subducted below the other, the subduction zone at the collision site will be closed off (B). Part of the original relative motion between the two island arcs could be maintained by continued subduction adjacent to one of them. However, another possibility is a partial polarity reversal of the system by the establishment of a new subduction zone behind the island arc on the overriding plate (C) (key as for Figure 3.1).*

3.2 Intra - oceanic island arcs

There are about 20 intra-oceanic island arcs located along subduction zones at the present day, the great majority being found in the western Pacific Ocean (Fig. 3.6). Such volcanic island arcs must be distinguished from other volcanic chains, such as the Hawaiian Islands, which are located far from plate boundaries and have a different origin (see Section 4.2.1).

3.2.1 General characteristics and formation

At a subduction zone the downgoing oceanic lithosphere finds its topographic expression in the trench formed on the frontal side of the island arc (Fig. 3.7). Much of the veneer of sediments resting on the oceanic crust of the downgoing plate will not be subducted since it is not firmly attached to the underlying crust and because it is not sufficiently dense to sink of its own accord into the asthenosphere. Consequently it either accumulates in the trench or is scraped off the downgoing plate and forms wedges of sediment on the inner (arc side) of the trench.

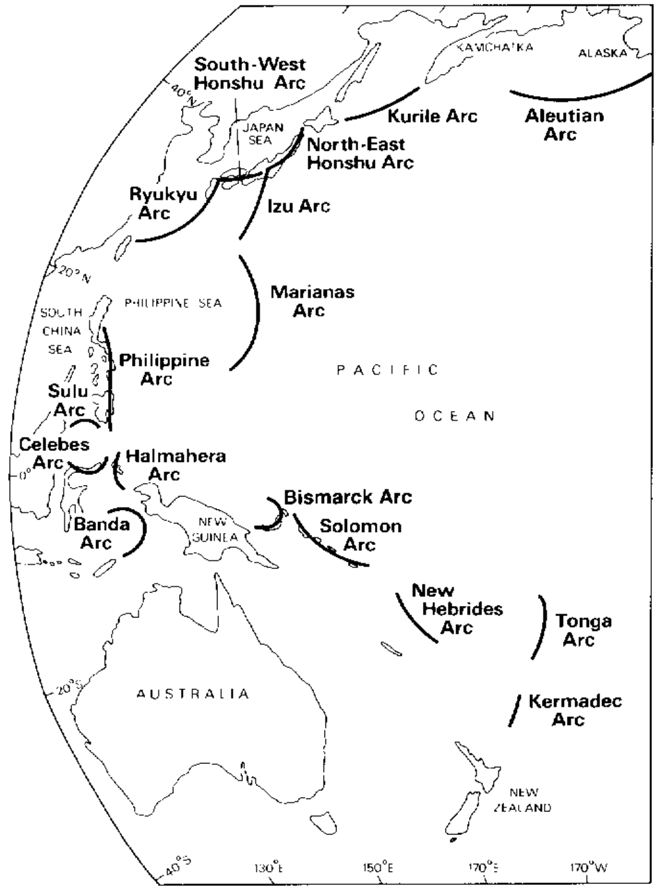

Fig. 3.6 *Location of intra-oceanic island arcs in the western Pacific Ocean. Other intra-oceanic arcs are the Lesser Antilles in the Caribbean, the South Sandwich Islands (Scotia Arc) in the South Atlantic and the Andaman and Nicobar Islands between Burma and Sumatra (see Figure 3.9).*

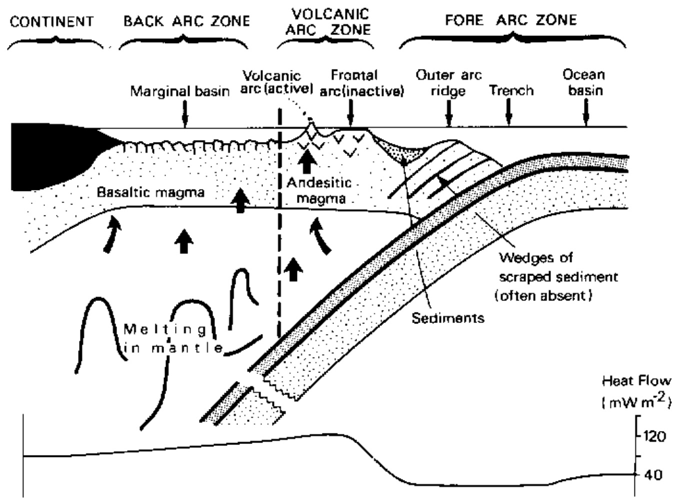

Fig. 3.7 *Schematic representation of the processes operating during the formation of intra-oceanic island arcs and the major morphological elements common to most arcs.*

As the plate descends frictional heat is generated between its upper surface and the surrounding mantle. Heating probably begins at quite shallow depths but because the downgoing plate is cold virtually all of this heat is initially absorbed. This is why low heat flows are recorded at the surface immediately behind the trench (Fig. 3.7). At a depth of about 100 km melting begins to occur along the edge of the subducted lithosphere and heat begins to be transferred upwards by rising pockets of magma. The dramatic surface expression of this marked increase in heat flow is the formation of volcanoes along the arc although relief is also generated by the emplacement of large igneous intrusions (see Section 5.1) in the underlying crust. The most common form of volcanic activity in this zone is the eruption of andesitic lavas, although there is a broad compositional range from basalt to rhyolite.

Fairly consistent patterns are evident in the location of the volcanic activity with respect to the fringing trench. The volcanoes tend to be located between 80 and 150 km vertically above the Wadati–Benioff zone marking the position of the downgoing lithosphere. This is related to the angle of subduction and the depth at which melting of the subducted lithosphere becomes significant. With typical angles of subduction of 30–50° this means that there is a gap of some 75–175 km between the trench and the arc of volcanoes. The active volcanic belt may consist of a single chain of volcanoes or form a broad belt up to 150 km in width. At any one time, however, volcanic activity is generally confined to a zone no more than 50 km across.

3.2.2 Morphological and structural elements

If we look at intra-oceanic island arcs in more detail it is possible to identify several major morphological and structural elements that are common to many (Fig. 3.7). The fundamental division is into the **forearc**, **volcanic-arc** and **back-arc** zones. Beginning in the forearc zone and moving towards the volcanic arc the first feature encountered is the trench which rises to the **outer-arc ridge**. This is located over the subduction zone and consists largely of thrust sheets of sediment and slivers of oceanic crust, known as **ophiolite,** emplaced by the subducted plate. The outer-arc ridge is usually a submarine feature but in a few cases, such as Barbados in the West Indies and Middleton Island in the Aleutian Arc in Alaska, parts of it are exposed above sea level. Between the outer-arc ridge and the volcanic arc a **forearc basin** may be formed. The volcanic-arc zone itself may consist of two elements: the active volcanic arc and an inactive **frontal arc** located on its forearc side and composed of older volcanic rocks. The whole entity is sometimes referred to as a **magmatic arc.** In most island arcs only a relatively small proportion of the individual volcanoes actually rise above sea level.

A complicated pattern of basins and ridges is located behind the volcanic arc. This back-arc zone may simply consist of a large marginal basin separating the intra-oceanic island arc from an adjacent continent, or it may be a more complex feature containing one or more **remnant arcs,** separated from the volcanic arc and each other by **interarc basins**. All these structures are submarine.

A number of processes have been proposed to explain the formation of back-arc basins but no one mechanism seems to be applicable to all cases. One favoured process is known as **back-arc spreading** and this has been applied to several such basins. Japan, for example, is composed of relatively thick continental-type crust, some of it quite old (Early Palaeozoic), rather than the more usual modified oceanic crust of most other intra-oceanic island arcs. This is difficult to explain unless Japan has been detached from the eastern margin of the Eurasian continent and has migrated eastwards. The process thought to be responsible is broadly analogous to sea-floor spreading along mid-oceanic ridges. Indeed, linear magnetic anomalies have been identified behind a number of island arcs; those behind the Scotia Arc in the southern Atlantic Ocean, for instance, indicate that spreading has occurred here over the past 7–8 Ma. Back-arc spreading may be initiated within the volcanic arc because the high heat flux from below weakens the lithosphere. If the lithosphere of the back-arc zone has a component of motion away from the volcanic arc relative to the underlying asthenosphere the arc may be split apart. This eventually leads to the creation of a remnant arc separated by an interarc basin from an active volcanic arc.

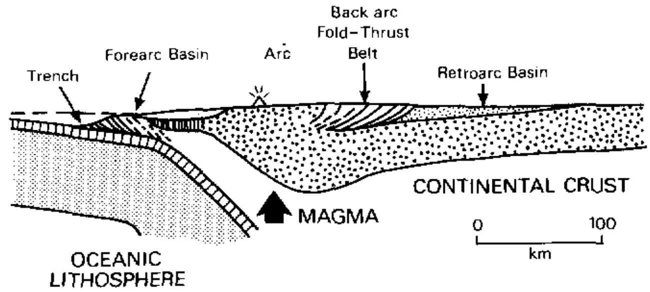

Fig. 3.8 Main structural and morphological features of a continental-margin orogen. The type illustrated is the 'contracted' form (see text). (After W. R. Dickinson (1977) in: M. Talwani and W. C. Pitman III (eds) Island Arcs, Deep Sea Trenches and Back-Arc Basins. *Copyright by the American Geophysical Union, Washington, DC, Fig. 5, p. 38.)*

3.3 Continental-margin orogens and continental-margin island arcs

Continental-margin orogens and continental-margin island arcs are broadly similar to intra-oceanic island arcs in their tectonic setting. The differences between them arise from the presence of continental crust on the margin of the overriding plate. This influences not only the nature of volcanic activity and landforms produced but also the processes tending to initiate back-arc spreading behind the volcanic arc. *Continental-margin orogens are found* along the western edge of South America (the Andes) and along part of the west coast of North America (the Cascade Mountains of Oregon and Washington, USA.). The western part of the Sunda Arc in the East Indies is an example of a continental-margin island arc since its back-arc zone comprises drowned continental crust (the Sunda Shelf).

As with an intra-oceanic island arc the oceanward side of a continental-margin orogen is marked by an oceanic trench associated with the downgoing oceanic lithosphere (Fig. 3.8). An outer-arc ridge, formed of wedges of deformed oceanic sediment and slivers of oceanic crust in addition to sediment originating from the adjacent continent, may undergo extensive vertical development and in some cases rise above sea level. The Mentawai Islands off the west coast of Sumatra represent the highest parts of the outer-arc ridge of the Sunda Arc and this structure can be traced northwards as far as the Indoburman Ranges in Burma (Fig. 3.9). Parts of the coastal ranges of Mexico, Peru and Chile similarly represent outer-arc ridges. Volcanicity and magma emplacement, largely in

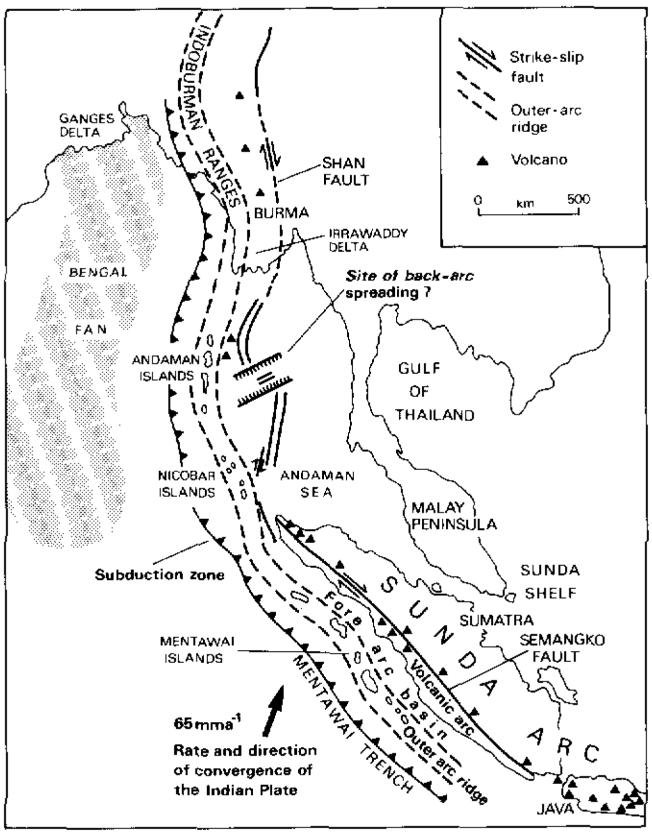

Fig. 3.9 Map of the major structural and morphological elements of the eastern margin of the Indian Ocean (based on various sources).

the form of large granite intrusions, characterize the volcanic arc. This is usually located between about 175 and 275 km from the trench, a rather greater distance than is typically the case with intra-oceanic island arcs.

The sequence of continental-margin orogen development begins with the subduction of oceanic lithosphere at, or close to, a continental margin. Frictional heat generated along the upper edge of the underthrusting lithospheric slab at depths exceeding 100 km leads to the upwelling of magma in an expanding dome within the growing orogenic belt which eventually rises above sea level. As the elevation of the orogen increases, the presence of hot, mobile rock within its core promotes gravitational spreading and sliding (see Section 7.3) through upwelling and lateral flowage into the more rigid outer crust. Eventually granite intrusions are emplaced at high levels within the mountain mass and volcanic activity develops. Extensive stretching of the crust above these granite intrusions produces a faulted terrain with active volcanoes. Crustal thickening, due in large measure to the vast volume of granite emplaced above the subduction zone, is considerable in some continental-margin orogens. For instance, crust up to at least 70 km thick occurs below parts of the Andes .

Some continental-margin orogens show evidence of deformation attributable to contraction in the back-arc region. They appear to be jammed against the adjacent cratonic block and are known as **contracted continental-margin orogens**. In other orogens, such as the coastal ranges of the north-west USA, this feature is absent and the term **non-contracted continental-margin orogen** is applied. In contracted continental-margin orogens the back-arc region, known as a **retroarc basin**, is separated from the rear of the volcanic arc by a system of down-thrusted and folded masses of rock (forming a **back-arc fold-thrust belt**) (Fig. 3.8). These features develop if the back-arc lithosphere behind the orogen is moving towards it relative to the underlying asthenosphere, since this convergence will probably lead to partial subduction of the back-arc lithosphere. As the lithosphere carrying the rigid cratonic crust moves towards, and underthrusts, the continental-margin orogen, the stripping of its sedimentary cover gives rise to the back-arc fold-thrust belt. This part of continental-margin orogens can therefore be viewed as a kind of subduction zone, albeit one that involves orders of magnitude less convergence that oceanic trench subduction zones.

The presence of anomalously thin crust, widespread extensional faulting and high rates of heat flow has led some researchers to interpret the Basin and Range Province of the western USA and northern Mexico as an area of back-arc spreading developed on continental crust behind the subduction zone along the Pacific coast. The region, which as a whole has been elevated by 1000–2000 m, has a fault-generated relief of up to 4000 m comprising uplifted mountain blocks and intervening basins (Fig. 3.10). Such back-

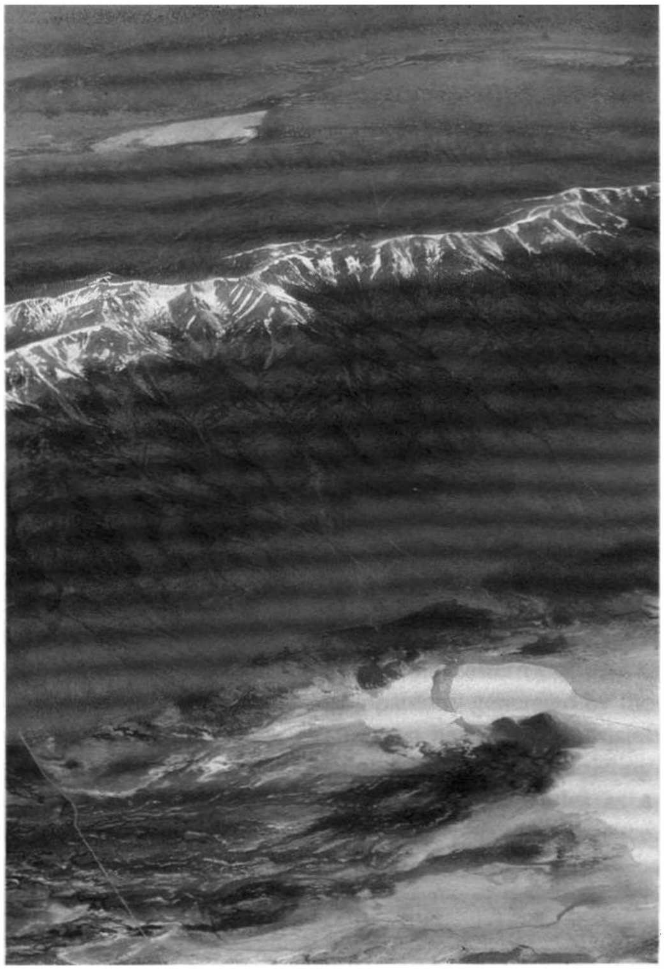

Fig. 3.10 *A snow-capped range separating sediment-filled basins in the faulted Basin and Range Province, Nevada, south-west USA.*

arc spreading is analogous to that encountered behind some intra-oceanic island arcs. However, the Pacific–North American plate boundary has undergone significant changes over the past few million years with the subduction of part of the East Pacific Rise spreading ridge and the evolution of a predominantly transform margin. It is possible that these changes have been critical in generating the considerable crustal extension (possibly up to 100 per cent) calculated for the Basin and Range Province.

3.3.1 The Andes

3.3.1.1 General characteristics

The Andes constitute by far the most spectacular present-day example of a continental-margin orogen. Stretching from the Toco Peninsula of Trinidad in the north to Tierra del Fuego in the south, they form a varied and complex mountain system some 9000 km long with elevations in excess of 5000 m. Three major divisions can be recognized (Fig. 3.11):

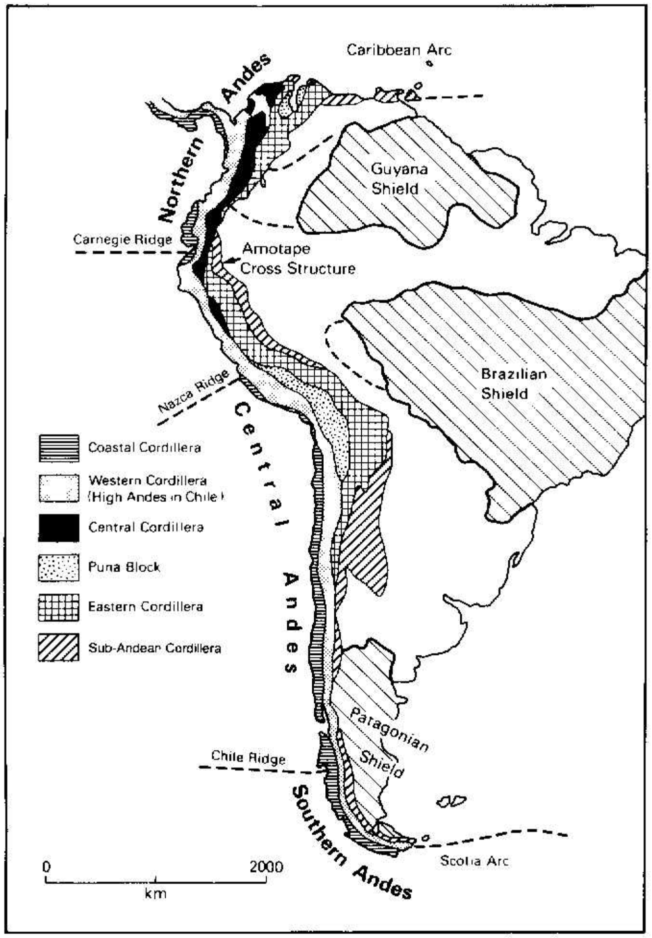

Fig. 3.11 *Major structural and morphological divisions of the Andes. (Modified from A. Gansser (1973)* Journal of the Geological Society London ***129****, Fig. 2, p. 96. Reproduced by permission of the Geological Society.)*

(1) the Southern (Patagonian) Andes, extending from Tierra del Fuego northwards to the Gulf of Penas at latitude 47 °S where the actively spreading Chile Ridge extends westwards into the Pacific; (2) the Central (Chilean–Peruvian) Andes reaching from the Gulf of Penas northwards to the Amotape cross structure at the Peru–Ecuador border where the northwestern trend of the mountain belt changes to a northeasterly orientation; and (3) the northern (Colombian–Venezuelan) Andes, extending from the Amotape cross structure northwards and then eastwards to eventually link with the Caribbean Arc.

Since the development of the plate tectonics model, the Andes have sometimes been regarded as representing a classic example of the orogenic consequences of the subduction of oceanic lithosphere beneath an overriding continental plate. This, however, is an over-simplified view as the current phase of uplift and volcanicity represents only the most recent episode in a complex history stretching back to the Mesozoic; moreover, the history of this vast mountain system differs greatly from one part to another.

In the range as a whole block faulting is significant and vertical movements predominate over thrusting and other effects involving laterally induced stress. Any plate tectonics model of the Andes must in fact account for the uplift essentially in terms of vertical tectonics. Another feature which has to be explained is the emplacement of massive granite intrusions of various ages which cover about 464 000 km^2 (about 15 per cent) of the surface of the range; this probably represents a volume in excess of $2 \times 10^6 km^3$ – the greatest mass of granite intrusions on Earth. They are concentrated in the Coastal Cordillera of the central Andes but also occur in the High Cordillera (Fig. 3.11). Volcanic activity is localized within the range; in the high plateaus of the central Andes in the border area of Chile, Argentina and Bolivia where the range reaches its greatest breadth, layers of lithified volcanic ash from 500 to 1500 m thick produced by explosive rhyolitic eruptions, cover an area in excess of 200 000 km^2. Quaternary lavas have erupted through these deposits to form over 900 volcanoes ranging in elevation from 5000 to 7000 m (Fig. 3.12).

Fig. 3.12 *The volcanic peak of Cotopaxi reaching an elevation of 5896 m in the Ecuadorian Andes. (Photo courtesy D. Munro.)*

3.3.1.2 Evolution of the central Andes

In the central Andes there has probably been more or less continuous subduction since the Mesozoic. Consumption of the Nazca Plate along the nascent Peru–Chile Trench had begun by at least the Late Triassic – Early Jurassic and marked the conversion of the previously 'passive' continental margin (see Chapter 4) of western South America to an active convergent plate boundary (Fig. 3.13 (A) and (B)). The start of subduction may have been associated with the initiation of spreading on the East Pacific Rise, but in any case it certainly preceded the splitting of South America away from Africa 135–140 Ma ago. This early phase of plate interaction was probably marked by the development of a volcanic arc off the north Chile coast and was apparently contemporaneous with minor emplacement of granite intrusions in the Eastern Cordillera.

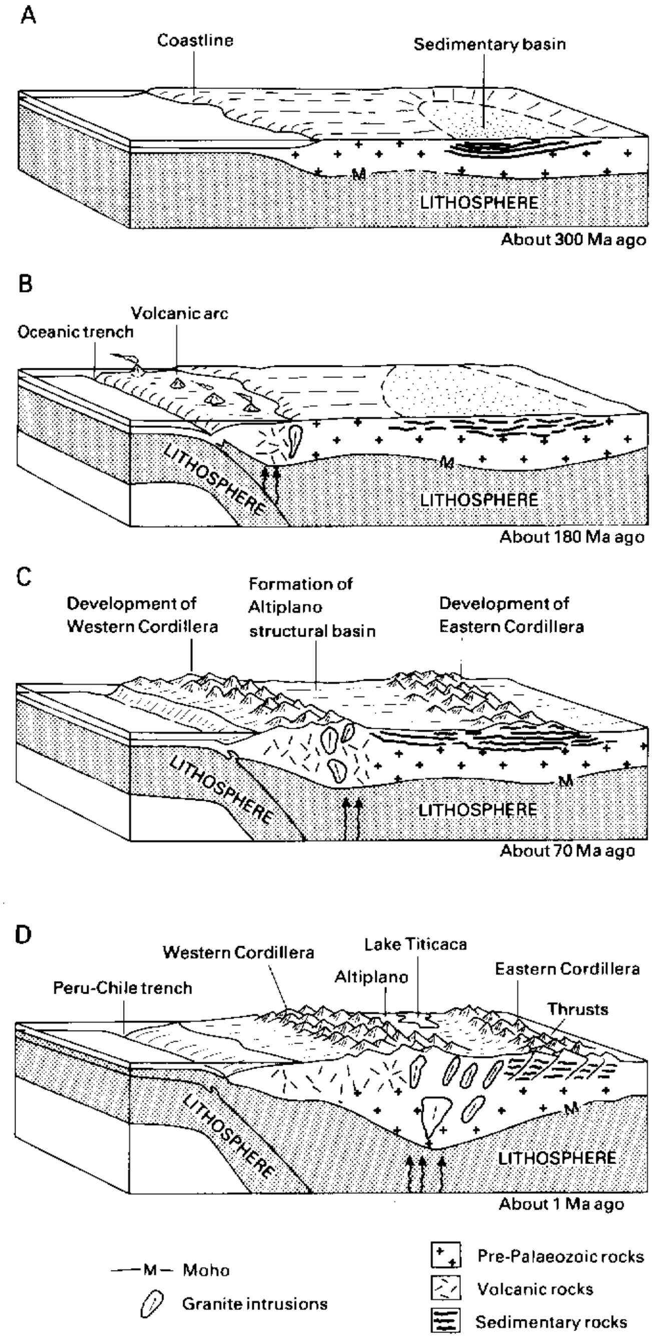

Fig. 3.13 *Model of the evolution of the central Andes. (Based partly on D. E. James, (1971)* Geological Society of America Bulletin ***82***, *Fig. 10, p. 3341.)*

A major episode of mountain building occurred in the Late Cretaceous–Early Cenozoic to the east of the earlier orogeny (Fig. 3.13(C)). This was the major period of emplacement of granite intrusions which resulted in the growth of the Western Cordillera. A further consequence of this large addition of granite to the core of the mountain mass was the transmission of compressive stresses to the east where the Eastern Cordillera were folded and uplifted. Between these two ranges a large structural basin was created which now forms the high plateau of the Altiplano. The modern phase of orogeny began in the Miocene with volcanism and the further emplacement of granite intrusions. A more active period of igneous activity in the Pliocene – Quaternary, involving massive emplacements of magma, caused intense folding and the formation of large thrust sheets in the Eastern Cordillera. A series of high, narrow ranges was formed as the Eastern Cordillera were pressed against the resistant block of the Brazilian Shield to the east (Fig. 3.13 (D)). This folded and thrusted belt probably represents the partial subduction of the western fringe of the Brazilian Shield, and the major change of the orientation of the Andes at the Amotape cross structure seems to be the result of the movement of separate basement blocks in the back-arc zone.

There appears to be a clear distinction between the causes of uplift on the western and eastern sides of the central Andes. In the Western Cordillera evidence of tensional deformation indicates that crustal stretching has been associated with the emplacement of magma and volcanic activity related to subduction zone melting. In the Eastern Cordillera, however, compressional stresses apparently generated by the growth of the volcanic arc to the west have caused crustal shortening, and uplift here is attributable to folding, thrusting and crustal thickening. Nevertheless, for the Andes as a whole, compressive stresses arising from plate convergence have not been the major cause of uplift.

3.3.2 The Sunda Arc

The islands which make up the East Indies archipelago form a region of fascinating complexity. For the present, however, we will confine our attention to the Sunda Arc comprising Sumatra, Java and the chain of smaller islands extending eastwards as far as longitude 123 °E (Fig. 3.14). To the west of this point oceanic lithosphere of the Indian Plate is currently being subducted below the Sunda Arc but to the east the Arc is in collision with the continental shelf of northern Australia (this is marked on Figure 3.14). There is thus a transition from a steady-state boundary in the west to a collision boundary in the east. Since the western part of the Sunda Arc is located on the edge of a block of continental crust (at present largely below sea level) it is a continental-margin island arc rather than an intra-oceanic island arc.

Across the Sunda Arc there is a distinct sequence of oceanic trench, outer-arc ridge, forearc basin, volcanic-arc and back-arc zone (Fig. 3.9). The volcanic arc is located about 125 km above the Wadati-Benioff zone representing the seismic activity along the downgoing Indian Plate. This maintains a shallow angle of subduction beneath the outer arc but steepens below the main volcanic arc to about 60°. Beneath Sumatra the subduction zone does not exceed a depth of 200 km, but north of Java it reaches 400 km and north of Flores, at the extreme eastern end of the zone of

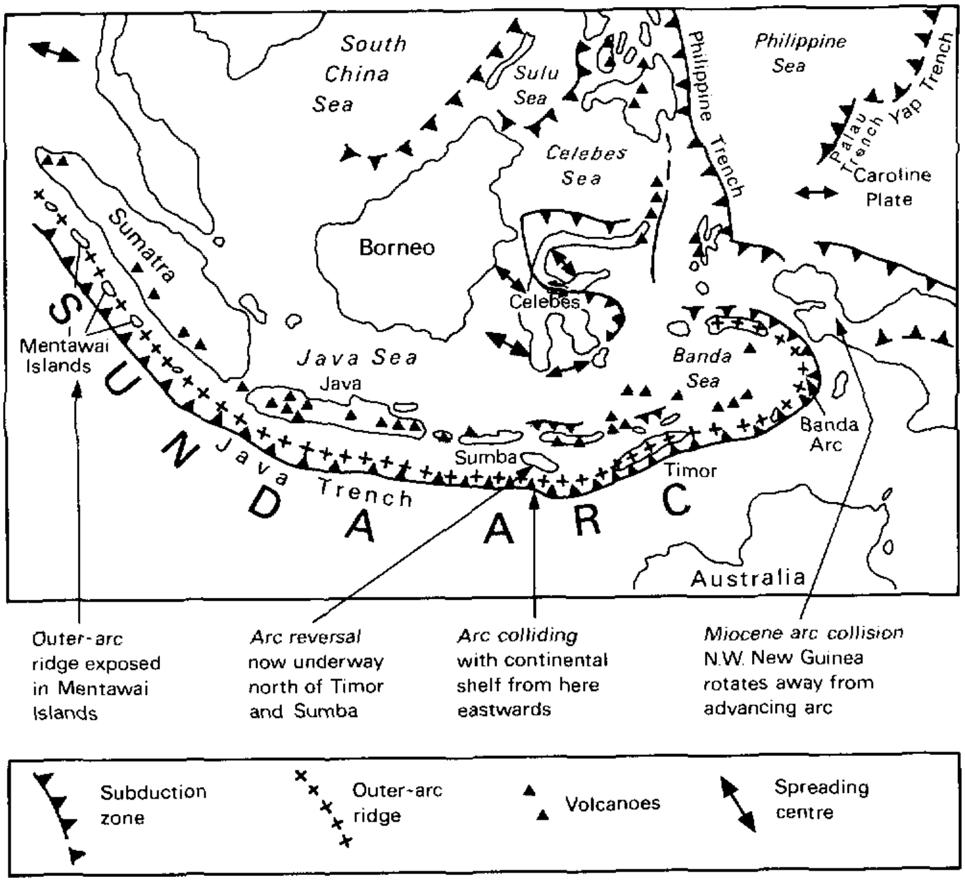

Fig. 3.14 The tectonic and morphological elements of the Sunda Arc and adjacent region. (Modified from W. Hamilton (1977) in: M. Talwani and W.C. Pitman III (eds) Island Arcs, Deep Sea Trenches and Back-Arc Basins. *Published by the American Geophysical Union, Washington, DC, Fig. 3, p. 18 and W. B. Hamilton (1988)* Geological Society of American Bulletin, ***100****, Fig. 2, p. 1511.*

steady-state subduction, it attains a depth of 700 km. Sumatra is composed of old, thick continental crust comprising volcanic rocks of Permian, Cretaceous and Cenozoic age. Recent volcanic rocks, including extensive sheets of volcanic ash, are predominantly acidic and intermediate in composition since the magmas have risen through continental crust. By contrast in Java, where the crust is relatively young and thin, the volcanic rocks are mostly intermediate in composition and the large ash sheets indicative of highly explosive eruptions are scarce (see Chapter 5).

The outer-arc ridge of the Sunda Arc is of particular morphological significance. Although it is from 1 to 3 km below sea level off Java, along most of the coast of Sumatra parts of it are exposed as the Mentawai and other islands (Figs 3.9 and 3.14). The height of the ridge is apparently related to the amount of sediment being moved into the adjacent subduction zone where lithosphere is being subducted at a rate of about 65 mm a^{-1}. Large quantities of sediment are present on the ocean floor off Sumatra because of its position with respect to the Bengal Fan which represents the vast submarine extension of the Ganges delta on the Indian subcontinent some 2000 km to the north (Fig. 3.9).

3.4 Intercontinental collision orogens

Although crustal collisions may involve intra-oceanic as well as continental-margin island arcs, it is the convergence and eventual collision of continental crust that gives rise to intercontinental collision orogens (Fig. 3.15). Such mountain belts develop when the oceanic lithosphere originally lying between two continents is eventually consumed. The previously existing subduction zone is converted into a suture zone marking where the two continents are welded together. An embayment of oceanic crust not consumed before the boundary becomes 'locked' is called a **remnant-ocean basin.** **A peripheral foreland basin** is developed on the margin of the underthrusting plate roughly parallel to the strike of the suture belt, either on a remnant-ocean basin covered by a thick layer of sediment, or on true continental crust. A **retroarc foreland basin** occurs on the overriding plate behind the volcanic arc previously developed in association with the earlier subduction of oceanic lithosphere.

3.4.1 General sequence of development

The initial impact occurs between the outer-arc ridge of the

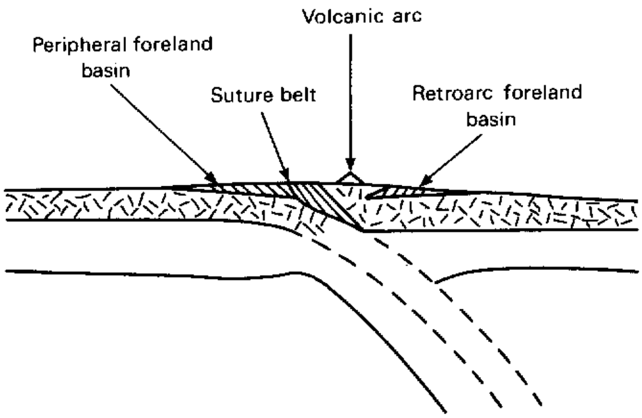

Fig. 3.15 Major structural and morphological elements of an intercontinental collision orogen.

overriding continental margin and promontories on the leading edge of the continent entering the subduction zone. The first sections of the converging continental margins to collide suffer the most intense deformation. Wedges of oceanic crust are thrust up on to the overlying sediments of the subduction zone and uplift ensues (Fig. 3.16). Mountain ranges formed at the points of initial continental collision provide the sediment which accumulates in adjacent remnant-ocean basins (Fig. 3.16(B)). On reaching the subduction zone this sediment is scraped off and stacked into thrust sheets. Final closing of a remnant-ocean basin results in the thrusting of remaining sediments deposited along the continental margin on to the overriding plate and the crustal thickening produced leads to isostatic uplift (Fig. 3.16(C)).

Large-scale crustal shortening is often present in intercontinental collision orogens and this may be accommodated by folding (Figs 3.17, 3.18), or by the limited subduction of continental crust during the initial phase of collision following the previous episode of steady-state subduction. There appear to be two possible mechanisms to account for this. Since the mantle part of the descending lithosphere is cold relative to the asthenosphere it may be sufficiently dense to pull the lower part of the crust down if the upper part is detached. Alternatively, subducted oceanic crust may be able to pull adjacent continental crust down into the asthenosphere. Depending on the assumptions applied this process could lead to the subduction of several hundreds of kilo-

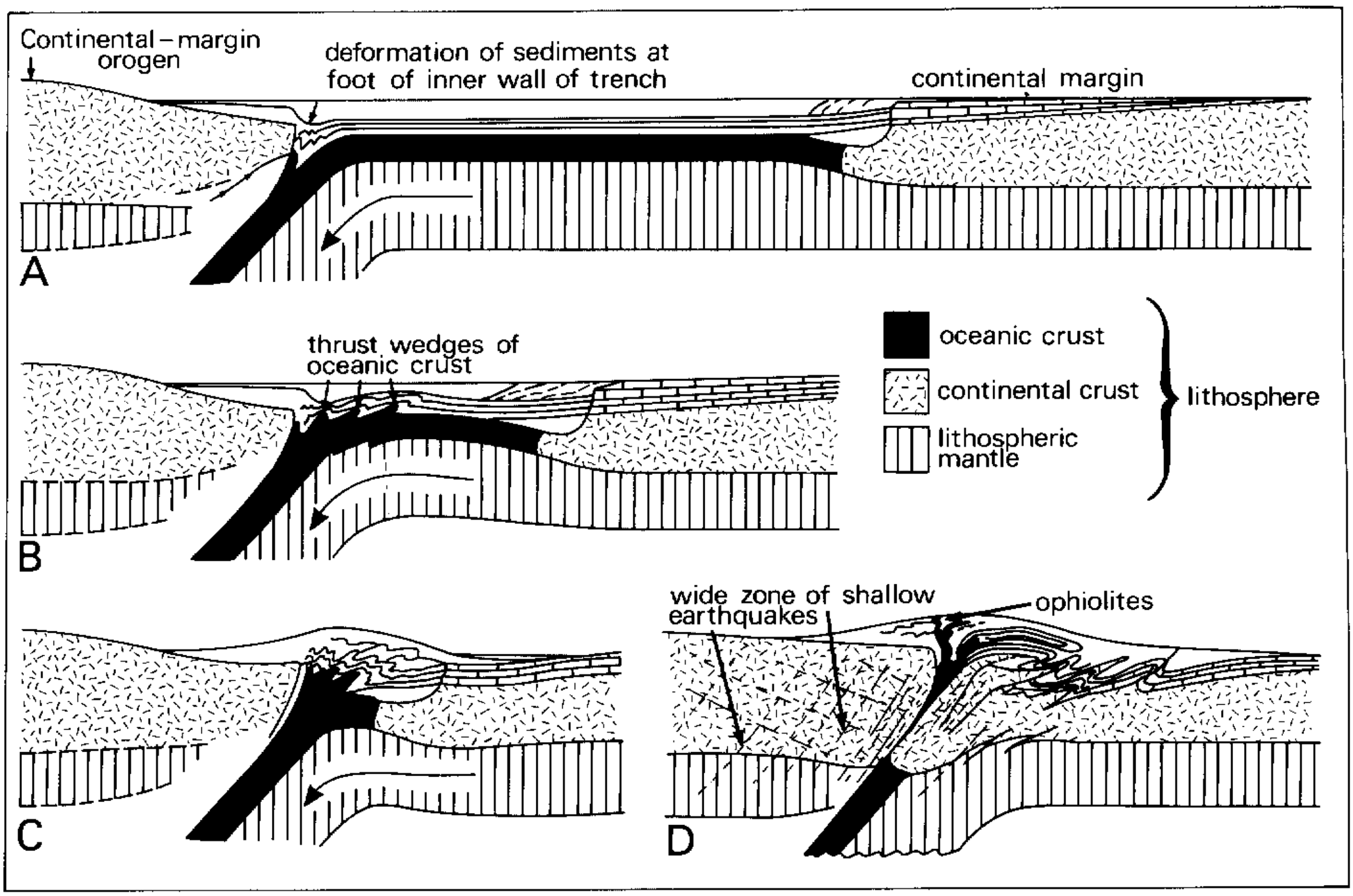

Fig. 3.16 Schematic representation of the development of an intercontinental collision orogen. (After J. F. Dewey and J. M. Bird, (1970) Journal of Geophysical Research *75, Fig. 13, p. 2642. Copyright by The American Geophysical Union.)*

metres of continental crust; more likely, however, is rather more limited subduction, but even this could involve the complete consumption of peninsulas and microcontinents.

Partial subduction of the underthrusting plate continues until frictional forces and the buoyancy of the continental crust exceed the driving force of plate motion (Fig. 3.16 (D)). Cessation of subduction is eventually followed by a halt to volcanic activity on the overriding plate. Further thrusting may develop at this stage through sliding of large rock masses under the force of gravity along the flanks of the uplifted orogen (see Section 7.3). Accumulation of sediment through the stacking of thrust sheets in the suture zone is probably the major cause of subsidence of the underlying subducted plate. Plate descent itself does not seem to be a major factor as isostatic rebound does not appear to occur after subduction has stopped. A peripheral foreland basin begins to develop as the continental crust of the underthrusting plate is bent just before entering the subduction zone.

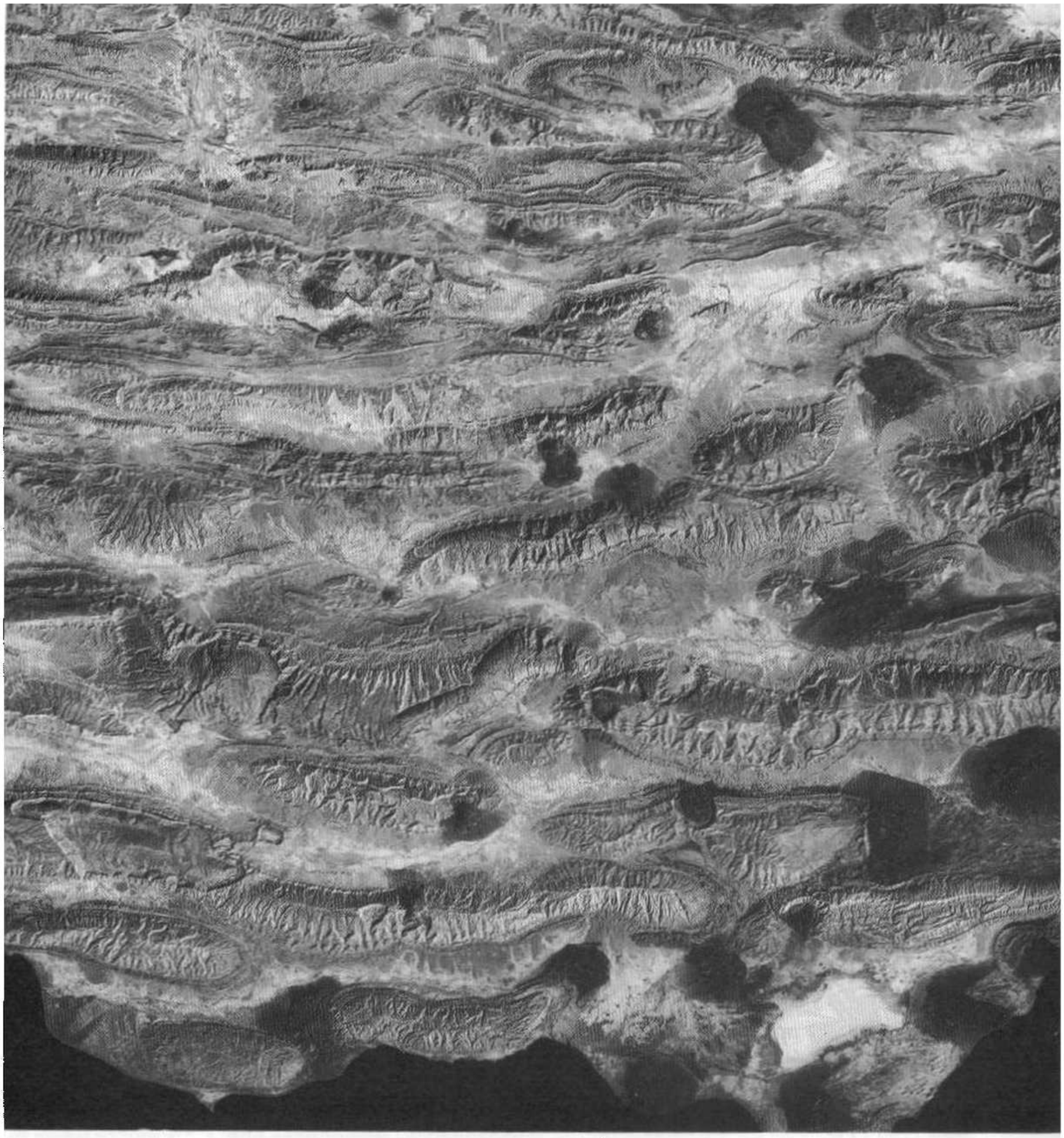

Fig. 3.17 *Landsat scene of part of the southern Zagros Mountains, Iran. Sediments of Mesozoic and Early Cenozoic age laid down on a continental shelf have been folded into a succession of plunging anticlines as a result of compression arising from intercontinental collision. The sediments contain thick salt beds which in places have intruded overlying strata as a result of their lower density and reached the surface to form salt fountains up to 1 km high (small, dark patches on image) The salt spreads under its own weight at rates of a few m a^{-1} forming salt flows. (Image courtesy N. M. Short.)*

After the phase of partial subduction and the accumulation of a considerable thickness of sediment, possibly up to 5000 m, the peripheral foreland basin may become a site of marked uplift.

Once the two continents have been welded together and active volcanism and tectonic uplift have ceased, a new plate is created in which the orogenic belt formed becomes an intra-plate feature no longer associated with an active convergent plate boundary. A number of ancient suture zones representing previous plate collisions have been recognized but only a few of these, such as the Urals and the Appalachians formed 300 to 250 Ma BP, are major relief features at the present day, most having been extensively eroded and covered by later sediments. Such ancient suture zones may be identified by various lines of evidence such as contrasting pre-collision geological histories either side of the postulated collision zone, the presence of deep ocean floor sediments and associated volcanic rocks, and preserved remnants of andesitic lavas erupted from the continental-margin orogen located on the originally overriding plate prior to collision.

In inter-continental collision orogens major sustained horizontal stresses can be generated which give rise to recumbent folds. Moreover, the back limbs of such folds may become detached along major thrust faults to produce nappes. The classic region for these kinds of structures is the European Alps. Here the convergence of the African and Eurasian Plates has led to the development of successive thrusts and nappes carrying huge masses of rock northwards towards

Fig. 3.18 Limb of the Saidmarreh anticline, southern Zagros Mountains, Iran. This plunging anticline is experiencing active uplift and although breached by a major stream has retained much of its primary form. (Photo courtesy Aerofilms Ltd.)

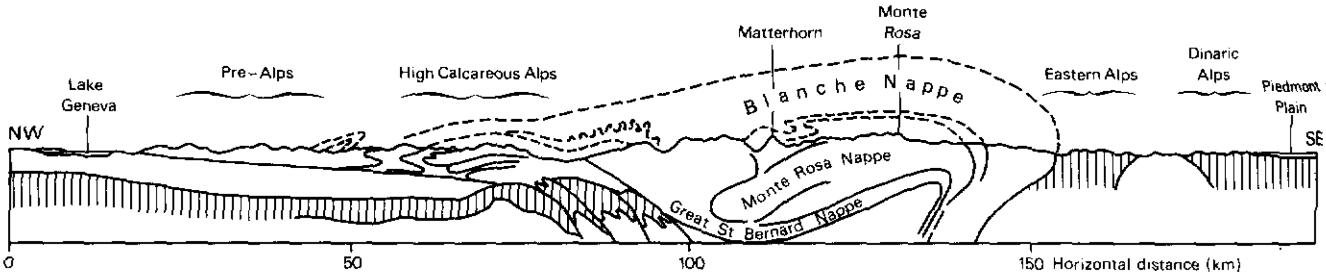

Fig. 3.19 Highly simplified cross-section from the piedmont plain of northern Italy north-westwards to Lake Geneva illustrating the relationship of major nappes to Alpine topography. (Modified from L. W. Collet (1935) The Structure of the Alps *(Edward Arnold, London) Plate XI after E. Argand, 1916,* Ecologae Geologicae Helveticae *24, 145–91).*

the European foreland (Fig. 3.19). Horizontal movements along these major thrusts can exceed 100 km and rock thicknesses of several kilometres can be involved. The thrust plane itself represents a zone of detachment (or **décollement**) along a bed of salt or other incompetent lithology. However, detailed investigations of these structures suggest that much of this large-scale lateral movement is accomplished by the spreading or sliding of rock masses under gravity (see Section 7.3).

3.4.2 Pre-collision history and configuration of converging continental margins

Although it is possible to make some generalizations about the likely sequence of development of an intercontinental collision orogen, the pre-collision history of the active margin of the overriding plate and its subduction zone has an important influence on the subsequent evolution of an orogenic belt. The margin of the overriding plate is constantly being modified by the subduction of oceanic lithosphere beneath it; volcanic material is added, sediments accumulate and the margin may be broken by strike-slip faults. The most important of these variables is probably the heating of the plate margin since this will affect its thickness and strength. Thick, strong continental-margin crust will be associated with either slow or recently initiated subduction since in neither case would there be sufficient heating for melting of the upper surface of the downgoing lithosphere. A marked contrast in behaviour would be expected during continental collision between the very thick crust of the Andes, where relatively rapid subduction has occurred throughout at least the Cenozoic, and the Alpine region where the closure of the ocean separating North Africa and southern Europe has been more gradual.

The configuration of continental margins and the angle at which they converge also affects significantly the nature of the orogenic belt that develops (Fig. 3.20). One plate may simply override the other, but another possibility is that a 'flake' of continental crust from the upper surface of the downgoing plate may be thrust over the adjacent plate for a distance of perhaps 100 km or more. Separation of the upper third of the continental crust on a plate has the effect of reducing the overall buoyancy of the subducted crust by nearly one-half, and this in turn allows more subduction than would otherwise have occurred. The eastern Alps have been suggested as an example of such 'flake tectonics' since this mechanism accounts for the existence of a 'platelet' in the Alps which is much thinner than normal and dips in a southerly direction (Fig. 3.20(B)).

Such a structure is particularly likely to form where a promontory on the advancing edge of the continent on the subducting plate collides with the opposing continent well before the rest of the continental margin (Fig. 3.20(E), (F)). Convergence associated with the continuing subduction of the remaining oceanic lithosphere along the plate margin would be maintained until the plate boundary had been fixed at a sufficient number of points to resist further subduction. Salients of oceanic crust may be left along the suture zone and parts of the Mediterranean Sea represent remnant ocean basins formed in this way.

3.4.3 The Himalayas

3.4.3.1 Morphological and structural elements

The Himalayan intercontinental collision orogen extends in a southward-bending arc, some 200–250 km across, for over 2500 km from the Indus River in the west to the Brahmaputra River in the east (Fig. 3.21). It is separated from the Trans-Himalayan zone and the Tibetan Plateau to the north by the structurally controlled valleys of the Indus and the Brahmaputra (in its upper reaches called the Tsangpo) and to the south it is bounded by the sediment-filled peripheral foreland basin of the Indo-Gangetic plain. Distinct morphotectonic zones can be recognized across the orogen along its entire length, although their detailed interpretation is disputed (Fig. 3.22). The two major ranges, the Lesser Himalayas and the Higher Himalayas, both of which exceed 8000 m in height, are bounded by three major linear structures: from south to north these are the Main Boundary

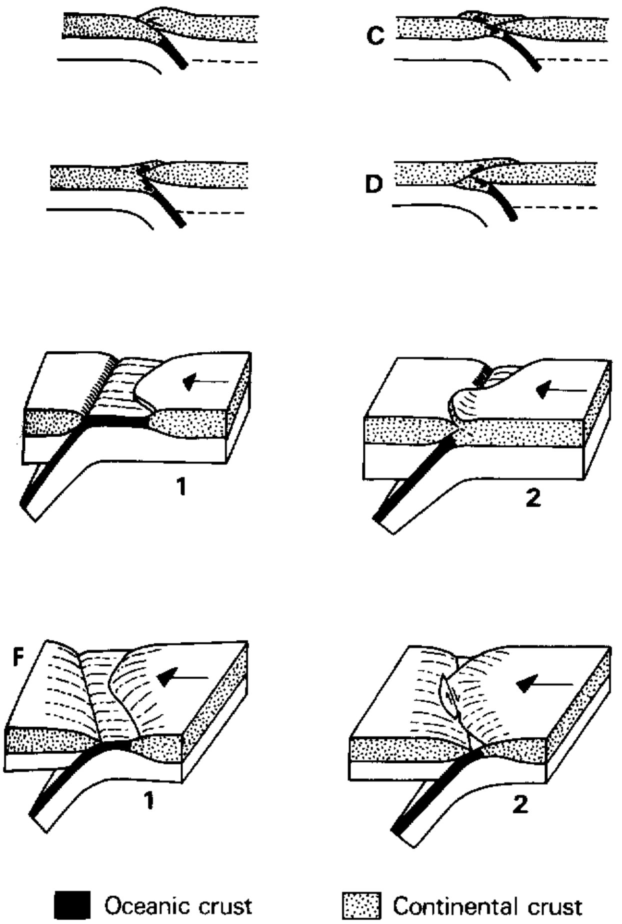

Fig. 3.20 *Various ways in which the collision between the opposing margins of two continents may be accommodated. Shown are: (A) simple overriding; (B) flake formation (e.g. eastern Alps); (C) a detached flake; (D) a deep flake with separating fractures extending below the Moho (e.g. central Alps); (E) a three-dimensional view of flake formation along an irregular continental margin; (F) flake formation associated with oblique subduction and consequential development of strike-slip faulting. (Modified from E. R. Oxburgh (1972)* Nature *239, Figs 2 and 3, p. 204 (E), and E. R. Oxburgh (1974)* Proceedings of the Geologists Association *85, Fig. 26, p. 344 (F) and Fig. 27, p. 345 (A-D).)*

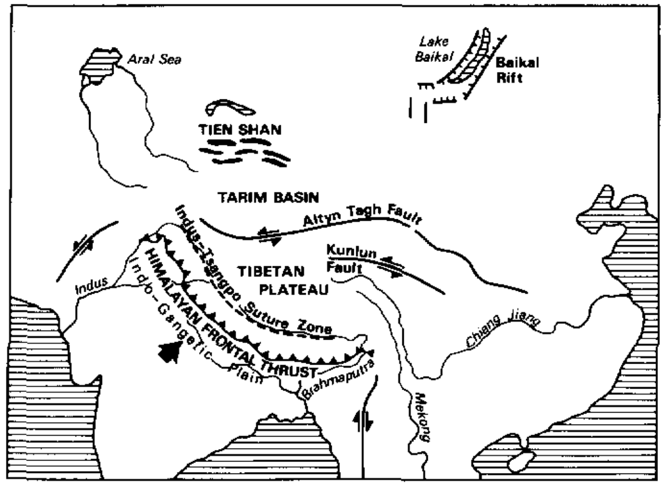

Fig. 3.21 *The Himalayas and Tibetan Plateau in relation to the major structural elements of central Asia (based on various sources).*

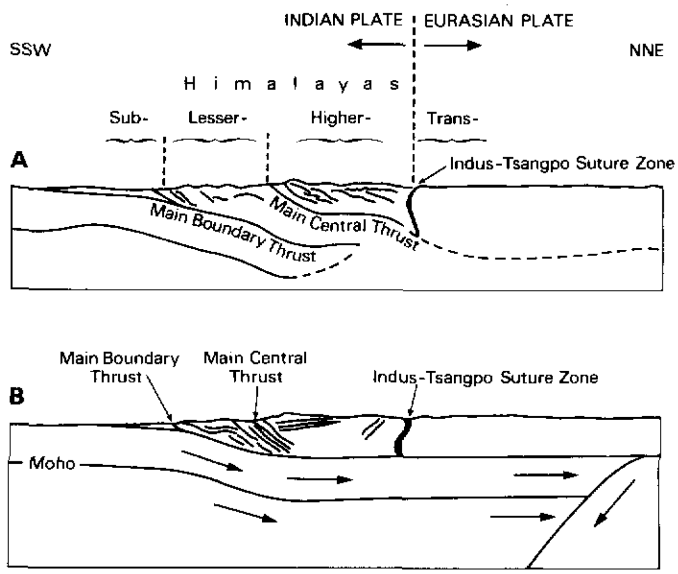

Fig. 3.22 *Two interpretations of the tectonics of the Himalayas according to Le Fort (1975) (A), and Powell and Conaghan (1973) (B). (Modified from P. Le Fort (1975)* American Journal of Science *275A, Fig. 15, p. 38).*

Thrust, the Main Central Thrust and the Indus-Tsangpo suture zone. The belt of ophiolites following the Indus and Tsangpo valleys marks the boundary between the Eurasian Plate and the Indian Plate and to the north the presence of acidic volcanic rocks indicates the subduction of oceanic lithosphere prior to the collision of the Indian and Eurasian land masses.

3.4.3.2 Models of development

From the pattern of ocean floor magnetic anomalies it is possible to trace the movement of India northwards over the past 80 Ma or so (Fig. 3.23). During the period 80 to 50 Ma BP it drifted rapidly at a rate of between 100 and 180 mm a^{-1}. Subsequently, movement has been much slower at only about 50 mm a^{-1}. This reduction in velocity seems to be related to the timing of the contact between India and Eurasia. Collision along what is now the Indus–Tsangpo suture zone seems to have occurred from about the Late Paleocene until the Early Eocene, or possibly a little later. Only limited continental subduction appears to have occurred at this stage. Significant vertical uplift probably began in the Oligocene about 35 Ma ago and has continued to the present time, but at varying rates.

Although clearly an intercontinental collision orogen, the

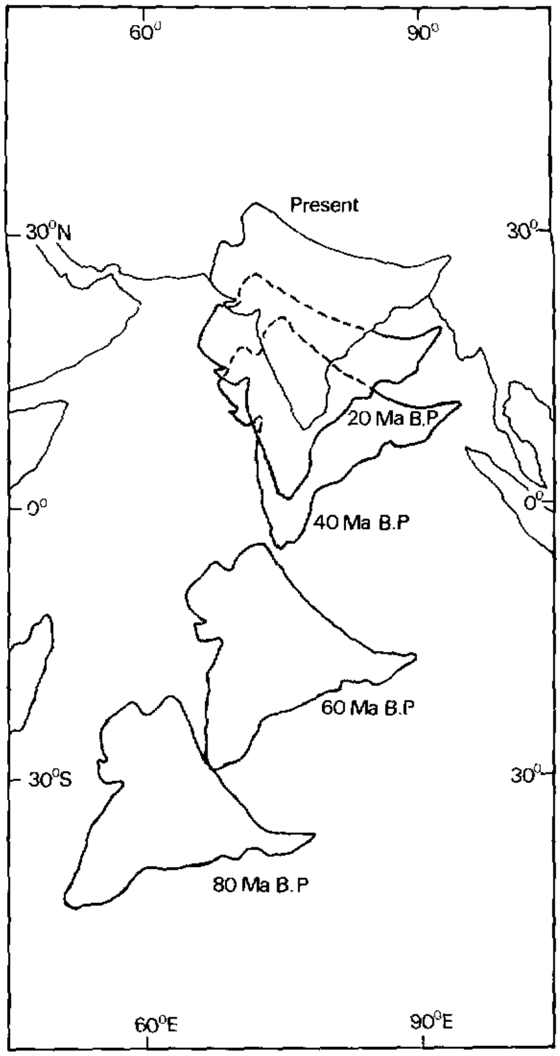

Fig. 3.23 *The northward drift of India over the past 80 Ma reconstructed from palaeomagnetic data. The position of Eurasia is assumed to be fixed in its present location and the configuration of the Indian sub-continent is drawn in its present form (although the shape of the northern boundary prior to collision is unknown). (Based on A. G. Smith, A. M. Hurley and J. C. Briden (1981)* Phanerozoic Paleocontinental World Maps. *Cambridge University Press, Cambridge Map 2, p. 9, Map 10, p. 17, Map 14, p. 21, Map 18, p. 25 and Map 22, p. 29.)*

present elevated form of the Himalayas is not a direct consequence of continental collision; rather it reflects developments that have occurred since the initial impact some 50 Ma BP. Subduction of the oceanic lithosphere of the leading edge of the Indian Plate prior to collision may have been associated with the emplacement of large granite intrusions as well as volcanic activity in southern Tibet. Little relative movement between the Indian and Eurasian Plates seems to have occurred between the Middle Eocene and Early Miocene, but isostatic readjustment of the presumably thickened crust resulted in a withdrawal of the sea on the adjacent Indian continental lowland.

The most significant development during this period was the formation of the Central Gneiss Zone, some 10–20 km broad and dipping at an angle of 30–40° to the north and bounded to the south by the Main Central Thrust (Fig. 3.22). The development of this zone appears to be related to the major underthrusting and associated metamorphism of local rocks that has apparently occurred along the Main Central Thrust some 100–200 km to the south of the Indus–Tsangpo suture zone. This underthrusting began in the Early Miocene and contributed to the doubling of crustal thickness (up to 80 km in places) and the impressive subsequent uplift of the Higher Himalayas. Detailed seismic evidence indicates that the underthrusting has been at an angle of about 15° over a distance of about 300 km; perhaps significantly this roughly corresponds to the average width of the Himalayas.

The precise relationship between the Main Central Thrust and the intense metamorphism of the Central Gneiss Zone of the Higher Himalayas remains somewhat uncertain. Some researchers have suggested that the metamorphism preceded thrusting and in fact actually promoted it by locally weakening the crust. Another interpretation is that metamorphism occurred at the same time, or post-dated thrusting, and that the frictional heating produced during thrusting contributed to the partial melting of the crust and the intrusion of granitic rocks. A further explanation involves the decoupling of the crust from the underlying lithosphere along the northern margin of the Indian Plate and the exposure of the base of the crust to hot asthenosphere. This could have led to the upwelling of magma, granite emplacement and isostatic uplift through the substitution of hot, low density mantle from the asthenosphere for the cold, dense mantle of the upper lithosphere. A further consequence of this heating could have been a weakning of the upper layers of the crust and the formation of the northward dipping Main Central Thrust along its southern edge.

The extent of subduction of continental crust below the Himalayas is also in dispute. At one extreme underthrusting is seen as proceeding for up to 1500 km at the base of the continental crust through the peeling off of the lower part of the lithosphere of the Eurasian Plate (Fig. 3.22(B)). It is more generally accepted, however, that the buoyancy of the continental crust has limited the extent of continental underthrusting and that this is why the more recent Main Boundary Thrust dipping under the Lesser Himalayas has now taken over the role played earlier by the Main Central Thrust (Fig. 3.22(A)). It seems likely that yet another northward-dipping thrust will develop even further to the south in a few million years.

Although the rate of movement between India and Eurasia fell markedly after their initial collision some 50 Ma ago, convergence of the two land masses since then may have amounted to as much as 2500 km. This represents a major problem because such an amount of continued convergence between two continents after collision is not easily accommodated within a simple plate tectonics model. The extent of crustal shortening required implies a vast degree of underthrusting by continental lithosphere, extensive crustal compression, crustal deformation over an extensive area behind the suture zone, or a combination of two or more of these effects. Critical to an evaluation of these various hypotheses is an understanding of the development of the high Tibetan Plateau lying to the north of the Himalayas.

3.4.4 The Tibetan Plateau

3.4.4.1 Morphology and structure

The Tibetan Plateau, located between the Himalayan and Karakorum ranges to the south, and the Kunlun and Altyn Tagh ranges to the north, is a roughly triangular area some 1000 km from north to south and 1700 km from east to west (Fig. 3.21). The local relief is relatively subdued but the mean elevation is some 5000 m with isolated masses of probable volcanic origin rising to between 6000 and 7000 m; the Tibetan Plateau is thus by far the highest plateau of significant extent on the Earth's surface. To the east the elevation falls and several major rivers such as the Mekong, Salween, Chiang Jiang (Yangtze) and Huang-He (Yellow) drain in an easterly and southerly direction from the Tibetan Plateau eventually reaching the Yellow and South China Seas.

The Tibetan Plateau has been an unstable area since at least the Late Palaeozoic. It seems to have formed through the successive accretion of continental and island arc fragments (see Section 3.6) from then until the middle Cretaceous and at the time of its collision with the Indian Plate it was probably an area of relatively warm and weak lithosphere. The presence of shallow marine sediments of Late Cretaceous age indicate that the region was below sea level at this time and it apparently did not emerge until the Early Cenozoic. Late Cenozoic volcanism, which is extensive over much of the area, and low seismic wave velocities together suggest that the lithosphere below the Tibetan Plateau is still unusually hot and is therefore probably weaker than in adjacent areas. The crust is now also abnormally thick (about 70 km) but gravity measurements indicate that it is in approximate isostatic equilibrium.

3.4.4.2 Models of development

A number of explanations have been put forward to account for the thick crust and high mean elevation of the Tibetan Plateau which also help to explain the great crustal shortening indicated by the post-collision convergence of the Indian and Eurasian Plates. One model proposes crustal thickening through large-scale underthrusting by the leading edge of the Indian continent (Fig. 3.22(B)), an idea first put forward as long ago as 1924 by the Swiss geologist, Emile Argand. A major problem with the extent of underthrusting required (in excess of 1000 km) is the very low angle of subduction that it implies (5° or less). This contrasts with the observed northwards dip of the Main Central Thrust at about 15°. Moreover, there is little seismic evidence for the intermediate and deep earthquakes below the Himalayas that would be expected if active subduction is, in fact, occurring.

Of the estimated post-collision convergence of up to 2500 km about 500 km could be accounted for by underthrusting beneath the Himalayas and southern Tibet, another 200–300 km by thrusting and crustal thickening in the various ranges bordering the northern perimeter of the Tibetan Plateau (such as the Pamir, Tien Shan, Altai and Nan Shan ranges) and perhaps a further 300–400 km by the crustal shortening arising from gentle folding within the Tibetan Plateau itself. This leaves around 1400 km of post-collision convergence to be accounted for, much of which might be explained by a doubling of the thickness of the Tibetan crust during the past 50 Ma from a 'normal' value of 35 km to the present 70 km. But not all the remaining convergence can be explained by this process and it has been suggested that the rest has been accommodated by the lateral movement of blocks of continental lithosphere along east–west trending

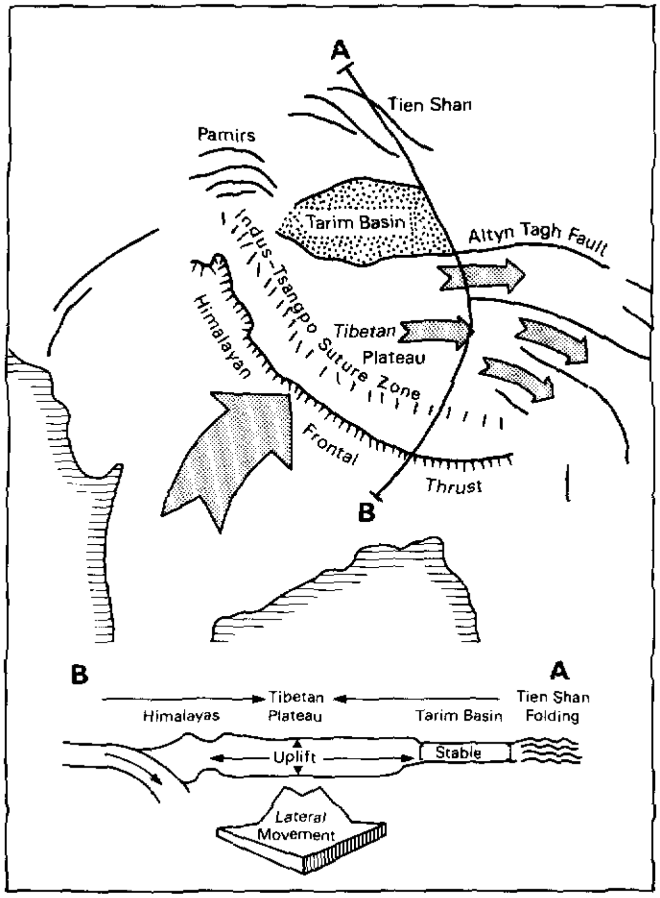

Fig. 3.24 Schematic representation of the possible effects of post-collision convergence between the Indian and Eurasian land masses based on the model of P. Molnar and P. Tapponnier.

strike-slip faults to the north of the Tibetan Plateau in Mongolia and western China (Fig. 3.24). Between 500 and 1000 km of north–south crustal shortening could be accounted for by an equivalent amount of lateral crustal movement arising from the 'ploughing' motion of the Indian Plate as it moved northwards and displaced lithospheric blocks in the Eurasian Plate. Supporting evidence for extensive east–west movement comes from the identification of a number of major faults clearly visible on satellite imagery, most notably the Altyn Tagh Fault (Fig. 3.25) which is comparable in length to the San Andreas Fault in California. The amount of displacement along these faults is unknown, but several major earthquakes associated with horizontal strike-slip movements have recently been recorded.

The hypothesized eastward movement of China towards the Pacific would be expected because the continental lithosphere of Eurasia would provide more resistance to lateral movement than the subduction zones of the western Pacific margin. Deformation of the weak Tibetan crust and flow in the upper mantle may have resulted from the north–south squeezing generated by plate convergence. The Tibetan Plateau can perhaps be viewed as the 'pressure gauge' of Asia, with the pressure applied by the Indian continent against Eurasia maintaining its considerable elevation. Moreover, the Tibetan Plateau transmits further pressure to the stable Tarim Basin to the north and this in turn pushes

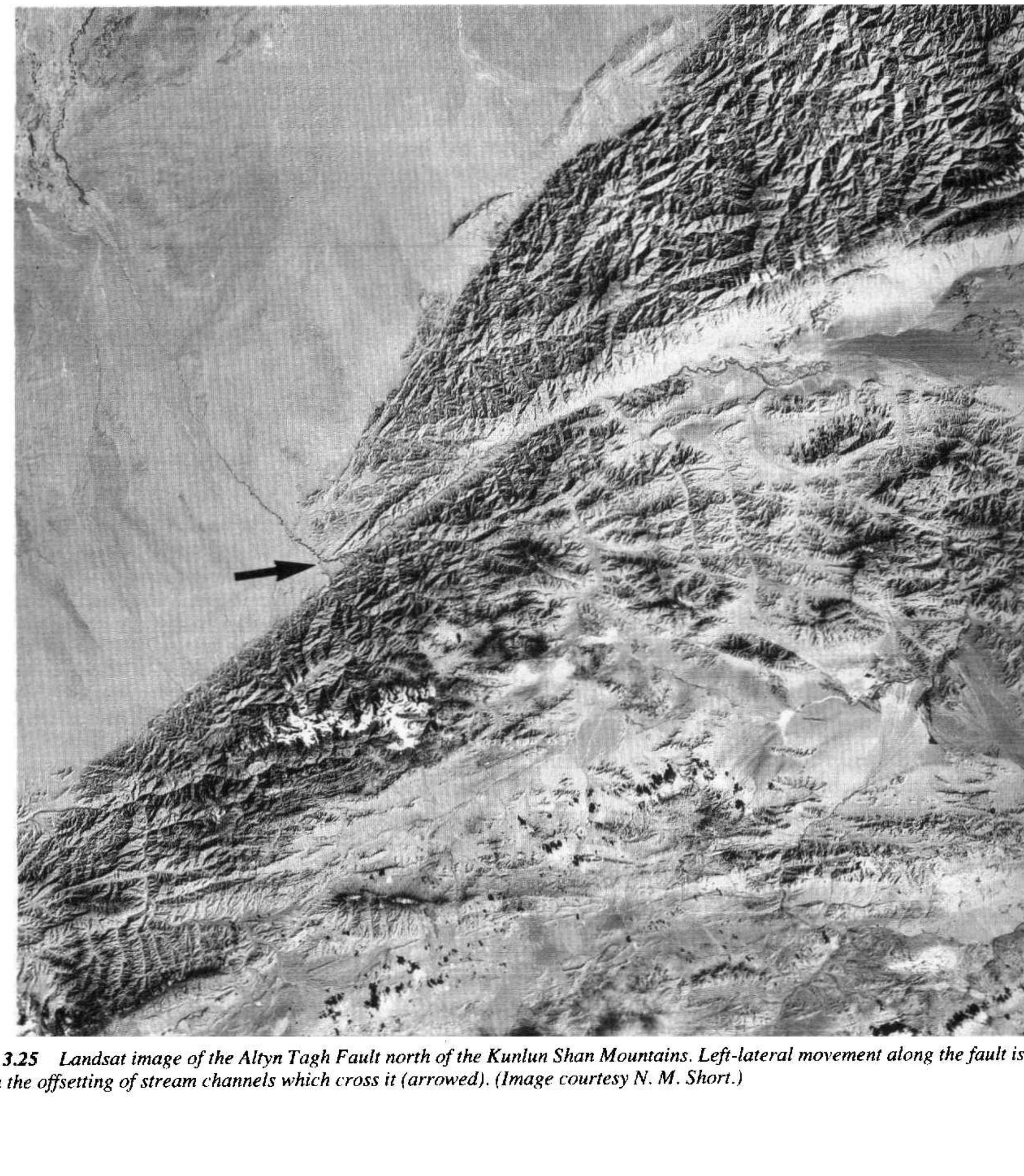

Fig. 3.25 *Landsat image of the Altyn Tagh Fault north of the Kunlun Shan Mountains. Left-lateral movement along the fault is evident from the offsetting of stream channels which cross it (arrowed). (Image courtesy N. M. Short.)*

northwards to produce thrusting and uplift in the Tien Shan range (Fig. 3.24). The association of thrust faulting on the margins of the Tibetan Plateau with normal faulting in its highest regions has been interpreted as suggesting that it has attained its maximum elevation, and that consequently it is tending to grow outwards rather than increase in altitude. This may be because lower stresses are required for the formation of new thrust faults on its margin than are needed to elevate the entire Tibetan Plateau.

Whatever explanation is applied to the post-collision convergence of the Indian and Eurasian land masses and the formation of the Tibetan Plateau, it is clear that it cannot be accommodated in any simple plate tectonics model. To consider the hypothesized eastward moving segments of lithosphere in Mongolia and China as involving 'micro-plates' is to employ the term 'plate' in quite a different sense from its original definition. The apparent extension of the effects of the collision of the Indian and Eurasian Plates to much of central and eastern Asia, and possibly as far as the Baikal Rift (Fig. 3.21), in fact indicates that the consequences of plate interaction need not necessarily be confined to plate margins.

3.5 Oblique-slip margins

So far we have confined our attention to convergent margins where two plates are in motion towards each other. We now need to consider margins where the movement is predominantly transform and plates meet along transform faults rather than at subduction zones or along collision boundaries. However, it will have become clear from the preceding discussion that purely convergent or transform movement is unlikely to occur except along short sections of a plate boundary. So-called convergent boundaries almost invariably have some element of lateral movement (see, for instance, Figure 3.20(F)), while most transform boundaries possess some convergent or divergent component. It is more realistic, therefore, to think in terms of a continuum of plate boundary types related to various degrees of convergence and divergence. Where a convergent or divergent margin has a significant transform component it is appropriate to refer to it as an **oblique-slip margin**.

Some idea of the variation that may occur along an individual plate boundary can be gained by considering the eastern margin of the Indian Ocean. Here oceanic lithosphere of the Indian Plate is being subducted beneath oceanic and continental lithosphere of the Eurasian Plate (Fig. 3.9). The Indian Plate is moving north-north-eastwards relative to the Eurasian Plate and the margin as a whole is one of low-angle oblique convergence. However, when examined in more detail it is possible to identify significant variations along the margin. For instance, although the forearc zone is predominantly one of convergence and compression, there are also transform elements, such as north of Sumatra. In the volcanic arc, both in the Sunda Arc and to the north in Burma, movement is predominantly along transform faults. These extend into the back-arc basin of the Andaman Sea where there is evidence of some divergent motion. Where the transform faults occur on land they are associated with significant topographic features such as the valley of the Irrawaddy River in Burma.

In order to emphasize the pattern of stress in oblique-slip margins, the terms **transpression** and **transtension** have been applied to, respectively, the convergent and divergent varieties. Although these two types of margin are associated with distinct types of landscape, they may occur in close proximity owing to the tendency of transform faults to contain offset segments or to have a sinuous form rather than being purely straight (Fig. 3.26). Divergent oblique-slip

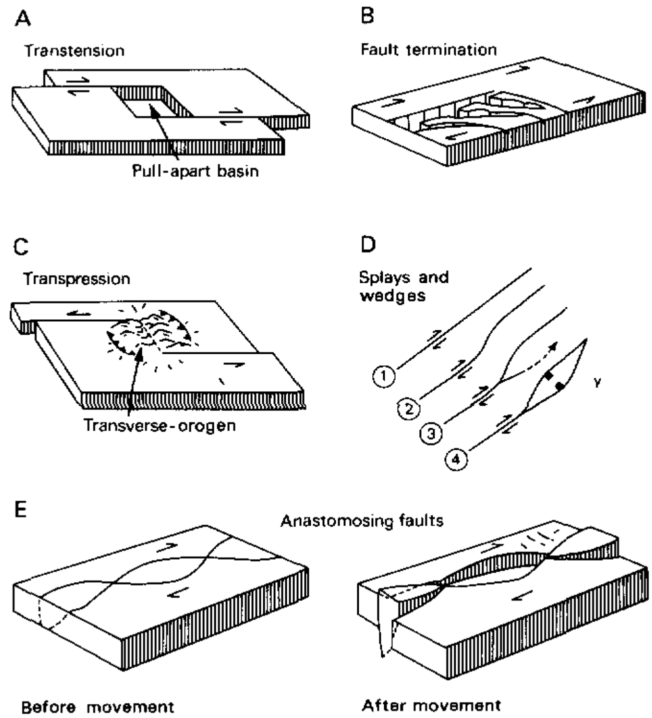

Fig. 3.26 *Fault structures and major landform features associated with oblique-slip margins: (A) pull-apart basin; (B) graben and horst structures caused by fault termination; (C) folding and faulting on a side-stepping fault and creation of transverse orogen; (D) development of fault splays and wedges through the progressive bending (2–4) of an originally straight fault (1); and (E) movement along anastomosing faults leading to upthrusting and the formation of down-sagged pods. (Based largely on J.C. Crowell (1974) in: W. R. Dickinson (ed.) Society of Economic Paleontologists and Mineralogists Special Publication 22* Tectonics and Sedimentation. *Figs. 5–7, p. 194; H. G. Reading (1980) in: P. F. Ballance and H. G. Reading (eds)* Sedimentation in Oblique-Slip Mobile Zones. *Special Publication of the International Association of Sedimentologists 4, Fig. 3, p. 12; and J. T. Kingma, (1958)* New Zealand Journal of Geology and Geophysics, *1, Fig. 1, p. 270).*

motion leads to extensional (normal) faulting and subsidence and if this is of sufficient magnitude a **pull-apart basin** may be formed. Oblique convergence, on the other hand, generates compression and results in folding, uplift and the formation of a **transverse orogen** astride the fault.

Complexity is added to this simple relationship because the predominant direction of stress along oblique-slip margins may change through time as a consequence of adjustments to the overall direction of plate movement. The San Andreas Fault System in California, for instance, changed from mainly transtensile in the Miocene to predominantly transpressive in the Pliocene. Very complex patterns of fracture are characteristic of major transform faults and these give rise to local zones of compression and extension over comparatively short distances. Because of the resulting close juxtaposition of areas of uplift and subsidence, considerable local relief can be generated.

3.5.1 The San Andreas Fault System

This situation is illustrated in classic fashion along the southern section of the San Andreas Fault System (Fig. 3.27). Overall the fault system has a length of 1200 km, but the average breadth of the fault zone is some 500 km and it forms a very diffuse boundary between the Pacific and North American Plates. It is estimated that there has been about 1000 km of lateral movement along the fault system over the past 25 Ma. Of the large number of individual faults (Figure 3.27 just shows some of the major ones) only a few are currently regarded as being active, although others have been active in the recent past.

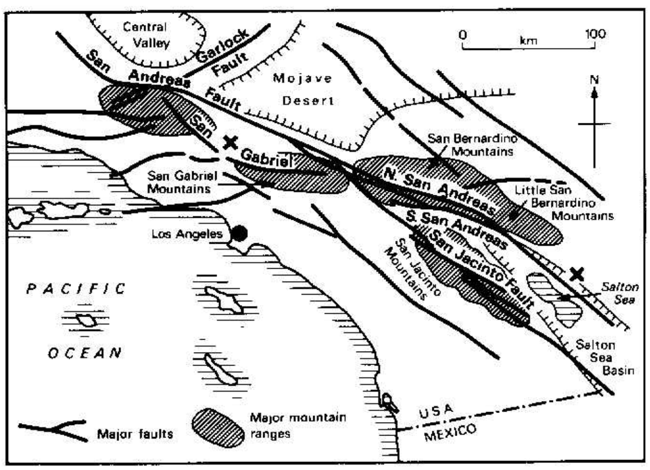

Fig. 3.27 Simplified map of the main structural and morphological features associated with the San Andreas Fault System in southern California. The points marked by an X indicate areas of similar terrane considered to have been originally adjacent but subsequently displaced laterally along the fault system. (Based on J. C. Crowell (1975) San Andreas Fault in Southern California. *California Division of Mines and Geology Special Report 118, Fig. 1, p. 11.)*

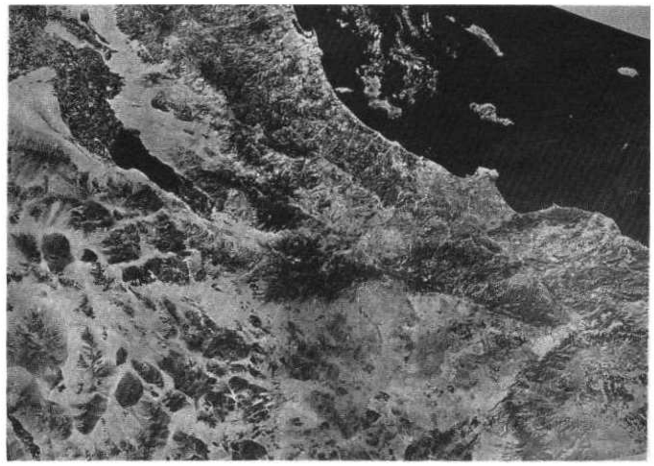

Fig. 3.28 Landsat mosaic showing the main structural and morphological features along the San Andreas Fault System in southern California. The mosaic covers approximately the same area as Figure 3.27. (Part of mosaic of ERTS-1 (Landsat) imagery prepared for NASA Goddard Space Flight Center by the USDA Soil Conservation Service.)

The individual faults which branch, join, bend and sidestep each other, give rise to numerous separate areas of uplift and subsidence (Fig. 3.28). A major subsidence feature is the Salton Sea trough, an excellent example of a pull-apart basin. Although it contains a considerable depth of sediment shed from the adjacent mountain ranges its floor still has a maximum elevation of just 74 m below sea level. By contrast an east–west trending zone of compression to the north-east of Los Angeles has led to the formation of a transverse orogen, comprising the San Gabriel and San Bernardino Mountains and collectively known as the Transverse Ranges. These attain elevations in excess of 3000 m and much of this uplift has probably occurred as recently as the Late Quaternary.

3.5.2 The Southern Alps

The Southern Alps of New Zealand provide another example of dramatic uplift and mountain construction along an oblique-slip margin. They run parallel to the Alpine Fault, a remarkably straight transform fault representing the boundary between the Indian and Pacific Plates. It is estimated that the present net movement along the fault involves an average of 40 mm a^{-1} of transform motion and about 22 mm a^{-1} of orthogonal convergence (Fig. 3.29).

Significant compression has only developed comparatively recently (the Alpine Fault was transtensile until the Pliocene) but has generated rapid uplift with the highest peaks in the Southern Alps now exceeding an altitude of 3000 m. The compression has caused thrusting on an eastward-dipping plane along which the continental crust of the Pacific Plate is being heaved up (Fig. 3.30). The western slopes of the Southern Alps are, in effect, an enormous self-perpetuating fault scarp; as rapidly as material is eroded it is replaced by

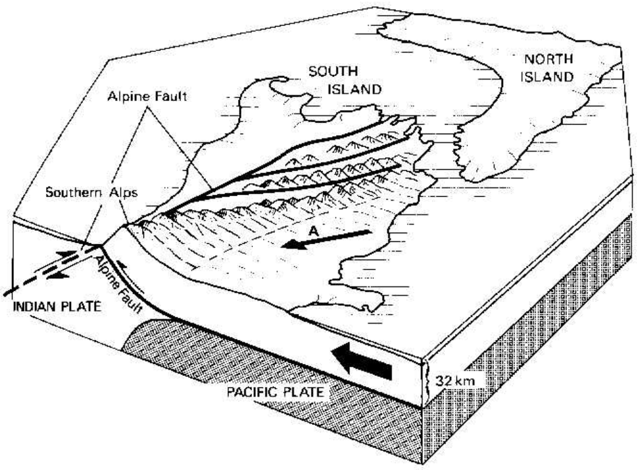

Fig. 3.29 *Schematic representation of the formation of the Southern Alps through the oblique convergence of the Pacific Plate along the Alpine Fault. The approximate direction of movement of the Pacific Plate is indicated by the arrow labelled A. (Based partly on J. Adams (1985) in: M. Morisawa and J. T. Hack (eds)* Tectonic Geomorphology. *Allen and Unwin, Boston, Fig. 5, p. 120.)*

further uplift and crustal shortening against the Alpine Fault (see Section 18.2.4).

3.6 Displaced terranes

As our knowledge of both ancient and modern mountain belts has developed, it has become increasingly evident that not all the complexities observed can simply be explained by the collision of large continental blocks. Many mountain belts situated along present or past plate boundaries contain numerous individual slivers of crust some of which, on the basis of palaeomagnetic and structural evidence, together with their assemblage of rock types, appear to have travelled great distances before arriving in their present positions. Other crustal fragments, while appearing to be of more local origin, have, none the less, experienced significant horizontal displacement and rotation with respect to the plate of which they form a part. These anomalous crustal fragments have been given various names–suspect terrane, allochthonous terrane–but we will use the term **displaced terrane** (Note that in this book the spelling 'terrane' is used to refer to structural units whereas 'terrain' is applied in a topographic sense.)

Several recent studies have indicated the significant role played by displaced terranes in the morphology and structure of both intercontinental collision orogens and continental-margin orogens. For instance, it has been estimated that over 70 per cent of the North American Cordillera is made up of displaced terranes (Fig. 3.31). Most of these seem to have travelled thousands of kilometres and have been accreted to the North American cratonic margin during the Mesozoic and Cenozoic. The terranes are of several types and include slivers of oceanic crust, fragments of intra-oceanic island

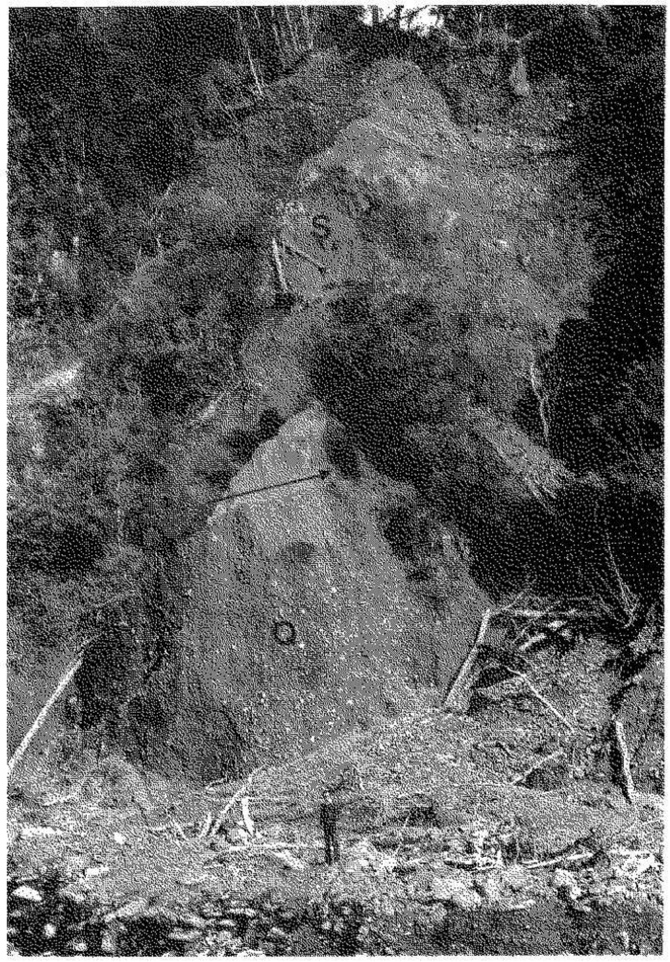

Fig. 3.30 *Schist (S) thrust over Quaternary glacial deposits (Q) on the western flank of the Southern Alps, New Zealand. Figure at the bottom of the photograph indicates scale. Fault plane is indicated by arrow.*

arcs and slices detached from unknown continental margins.

Numerous displaced terranes have also been identified in the Alpine–Himalayan chain (Fig. 3.32). Most of these are thought to have rifted away from the margins of Gondwana and moved northward before colliding with the southern margin of the Eurasian land mass and causing extensive deformation and thrusting. Indeed, it has been suggested that much of the structural and morphological complexity of the Alps may be due to the accretion or partial subduction of such crustal fragments rather than the effects of simple collision of two large land masses. Similarly, the morphology of the Himalayan orogen, in particular that of the Tibetan Plateau, may be partly attributable to the accretion of small crustal fragments prior to the main collision between India and Eurasia.

The composition of displaced terranes ranges from that of typical oceanic crust to significantly less dense granitic rock with clear continental affinities. It has been suggested that such terranes are equivalent to oceanic plateaus, which

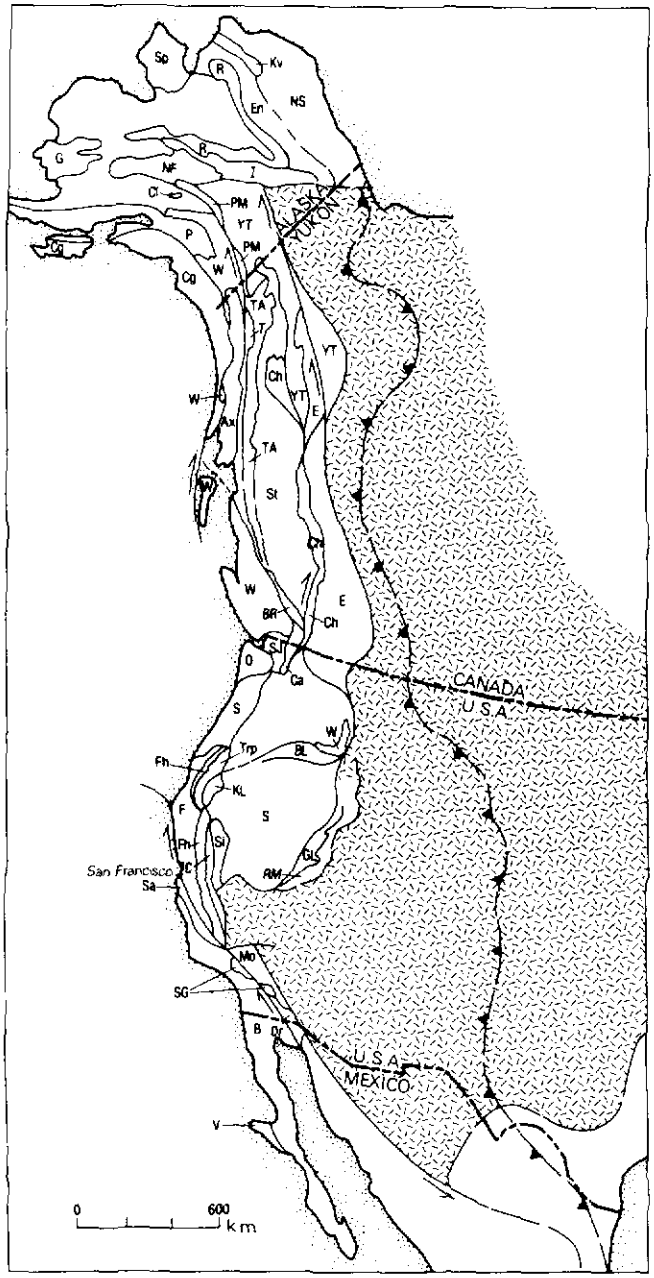

Fig. 3.31 *Map of displaced (suspect) terranes in western North America. The labelled blocks represent individual terranes which differ significantly in their geological characteristics. A proportion have clear continental affinities and of these some apparently originated in other continents at more southerly latitudes. Other terranes have rock associations indicative of an oceanic origin and represent island arcs and slivers of oceanic crust. The shaded area indicates the extent of North American cratonic basement and the arrowed line marks the eastern limit of Mesozoic–Cenozoic deformation in the North American Cordillera (From P.J. Coney et al. 1980* Nature **288**, *Fig. 1, p. 330.)*

are to be found in the present-day ocean basins and which often rise several kilometres above the adjacent ocean floor (Fig. 3.32). Many of these plateaus have anomalous crustal thicknesses of between 20 and 40 km and an upper layer 10 to 15 km thick with P-wave velocities in the range 6.0–6.3 km s^{-1}. These values are in the range of granitic rocks in the continental crust and so it is inferred that many oceanic plateaus (such as the Seychelles Bank) are drowned continental fragments originating from the edges of ancient land masses and destined to be swept towards a subduction zone in the future. As these plateaus are more buoyant than normal oceanic crust they would tend to resist subduction and therefore be accreted to continents lying along plate margins.

The implications of displaced terranes for the investigation of the relationships between tectonics and landform development have yet to be assessed. Nevertheless, the juxtaposition of terranes with often dramatically different lithologies and structural characteristics is likely to be a significant factor in the long-term development of landscapes.

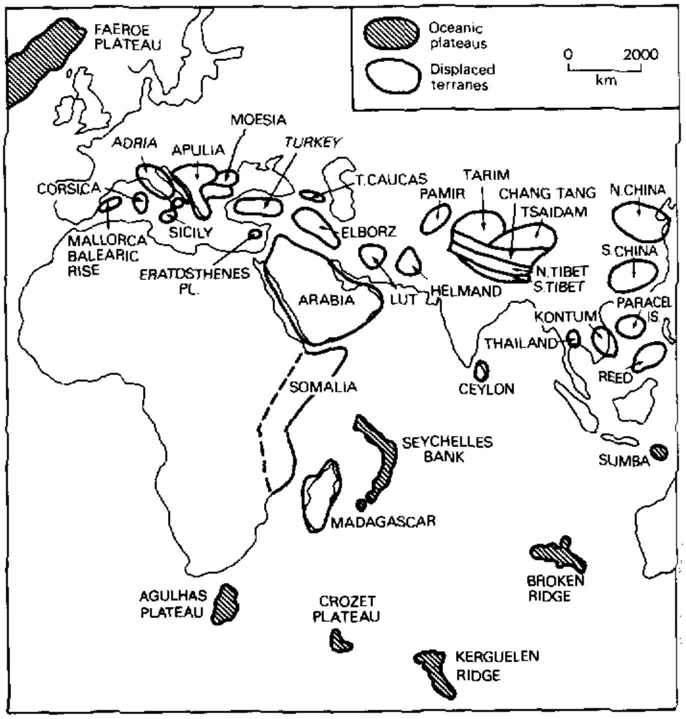

Fig. 3.32 *Map of the probable displaced terranes of the Alpine–Himalayan chain. Several major oceanic plateaus in the Indian Ocean are also shown. (Modified from A. Nur and Z. Ben-Avraham (1982)* Journal of Geophysical Research *87, Fig. 1, p. 3649. Copyright by the American Geophysical Union.)*

3.7 Mesoscale and microscale landforms associated with faulting

Plate margins are invariably regions of active faulting, and microscale and mesoscale landforms directly related to crustal movements along individual faults often provide important evidence of the nature of tectonic activity at a broader scale. Although it must be emphasized that fault-related landforms can be important in areas remote from plate boundaries it will be useful to consider them briefly here. Faulting takes place through sudden, rapid movements over distances ranging from a few centimetres to several metres, each individual movement being associated with an earthquake. Here we will focus on the primary tectonic landforms created by faulting and leave the question of the mode of subsequent landscape development of faulted terrains until Chapter 16.

Both normal and reverse faulting can create a **fault scarp**, the initial angle of which will reflect the dip of the fault plane. But weathering and slumping will very rapidly destroy the original form and reduce the slope to an angle of stability of typically between 20 and 40°. Fault scarps will only be formed where a fault breaks the surface; they can either die out laterally or merge into a monocline (Fig. 3.33(A)). Where faults die out laterally and are replaced by further offset faults an ***en echelon*** **scarp** can be formed (Fig. 3.33(B)). Vertical movement along a fault may lead to the impoundment of streams if they are flowing towards the uplifted block, whereas if they are flowing towards the downthrown side they will rapidly cut down through the fault scarp. This leads to the formation of gullies separated along the scarp by characteristic triangular facets (Fig. 3.33 (B)). Eventually the scarp will retreat from the original position of faulting and it may be buried by sediment. A scarp formed

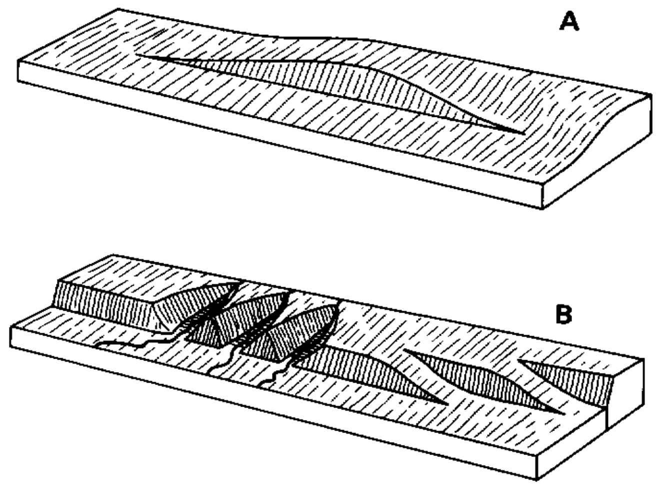

Fig. 3.33 Block diagram showing various forms of fault scarp and their associated erosional features: (A) a fault scarp dying out laterally but merging into a monocline to the right; (B) en echelon *scarps to the right and the development of gullies and triangular facets to the left.*

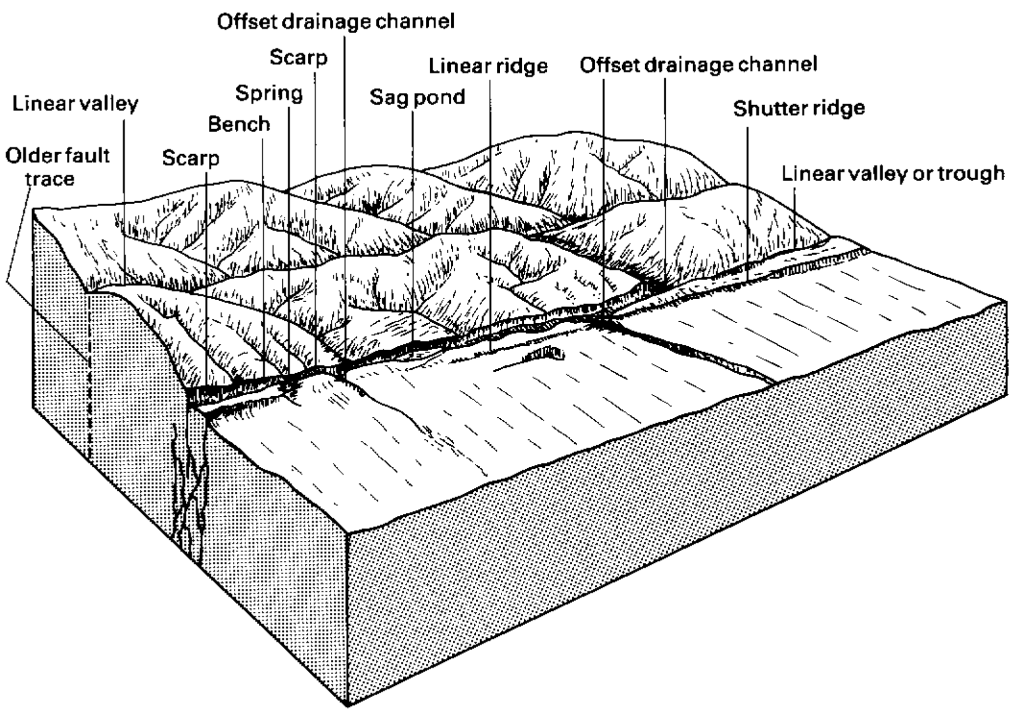

Fig. 3.34 Typical micro- and mesoscale landforms associated with major strike-slip faults. (After R. L. Wesson et al. *(1975)* United States Geological Survey Professional Paper *914A, Fig. 11, p. A21.)*

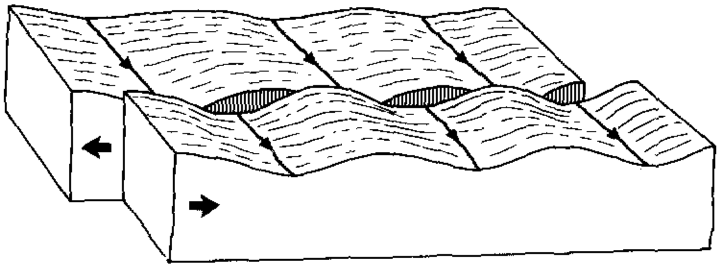

Fig. 3.35 Block diagram showing a shutter ridge formed along a left-lateral strike-slip fault.

at the site of a fault through the differential erosion of two rock types juxtaposed along it or through the exposure by erosion of a previously buried fault scarp is termed a **fault-line scarp**. It is important to distinguish such erosional features from true tectonic scarps since only the latter indicate recent or current fault activity.

Strike-slip faulting gives rise to an intriguing range of landforms at the micro- and mesoscale in addition to the major topographic features associated with oblique-slip margins (Fig. 3.34). Where there is no element of convergent or divergent movement along a strike-slip fault it may generate no significant landforms. If, however, faulting occurs across a gullied terrain the resulting offsetting of truncated gullies produces what are known as **shutter ridges** (Fig. 3.35). Localized zones of compressional and tensional stress are common along strike-slip faults and they give rise to a number of distinctive landforms. Divergence across the fault leads to subsidence and the formation of shallow, elongated troughs or sags, typically a few tens of metres across and a few hundred metres long, which, many contain a **sag pond**. In the zones of convergence, on the other hand, compression generates positive topographic features through the formation of ridges and linear and *en echelon* scarplets. Strike-slip faulting also usually leads to the displacement of stream channels (Fig. 3.25) and this feature is often useful in determining the magnitude and frequency of individual movements along the fault (see Section 16.3.2).

Further reading

The literature on the structure and tectonic development of plate margins is vast, but unfortunately very little of it takes a specifically geomorphic perspective. An excellent summary is provided by Miall (1984, pp. 384–440) which, although focusing on sedimentological aspects, also has much relevant material on the morphology of plate margins. Dickinson and Seely (1979) give a more specific treatment of forearc regions with excellent diagrams of their morphology, structure and evolution. A good coverage of the major processes of orogenesis, containing a number of detailed case studies, can be found in Hsü (1983) and Schaer and Rodgers (1987), while Von Huene (1984) and Molnar (1986) provide useful introductory reviews, and Molnar (1988) considers the problem of reconciling the plate tectonics model with the problem of orogenesis. More details on the complexities of convergent margins are contained in articles by Cross and Pilger (1982), Dewey (1980) and Uyeda (1982) while the possibility of the reversal of subduction polarity is assessed by Johnson and Jaques (1980) and Mattson (1979). Various aspects of island arcs are discussed in the collection of papers edited by Talwani and Pitman (1977), with a more specifically geomorphic perspective being provided by Nunn (1988), Woodroffe (1988) and Yoshikawa *et al.* (1981).

Continental-margin orogens are considered in several of the references mentioned above. Useful introductions to the tectonics and structure of the Andes are provided by Gansser (1973), James (1971) and Dalziel (1986). Garner (1983) discusses the tectonic geomorphology of the Ecuadorian Andes and Jordan *et al.* (1983) consider the effects on Andean topography of variations in the angle of plate subduction. The North American Cordillera, which has a tectonic history of bewildering complexity, has not been examined in detail in this chapter but various aspects of its evolution are considered by Dickinson (1976), Eaton (1987) and Smith and Eaton (1978). The Sunda Arc is discussed by Hamilton (1979) in a comprehensive study of the whole Indonesian region, while Hamilton (1977) provides a more concise analysis.

Articles considering specific aspects of intercontinental collision include those by Burke *et al.* (1977) on suture zones, Molnar and Gray (1979) on the question of limited continental subduction and Şengör (1976) on the effects of the irregular configuration of converging continental margins. There has been a surge of field studies in the Himalayan–Tibetan region since the late 1970s and this has generated a large body of published work. General overviews include those by Allègre *et al.* (1984), Gupta and Delany (1981), Molnar (1984, 1986) and Tapponnier *et al.* (1986), and Selby (1988) provides a specifically morphological study of the Nepal Himalaya. Two contrasting models of the evolution of the Himalayas are presented by Le Fort (1975) and Powell and Conaghan (1973), while the tectonic development of the Tibetan Plateau is discussed by Bingham and Klootwijk (1980) and Ni and York (1978). The model of the Tibetan

Plateau uplift associated with the lateral displacement of lithospheric blocks is introduced by Molnar and Tapponnier (1975), but there is also a more accessible and well-illustrated version (Molnar and Tapponnier, 1977).

Oblique-slip margins are considered in detail in various papers in Ballance and Reading (1980); the introductory article by Reading (1980) is particularly useful as it presents a general review of sedimentation patterns in relation to topography. The evolution of the San Andreas Fault System is concisely covered by Crowell (1974, 1979). The Southern Alps have been the subject of a number of studies examining the relationship between tectonics and landforms; these include those by Adams (1985), Basher *et al.* (1988), Whitehouse (1988) and articles included in Soons and Selby (1982). Berryman (1988) and Kamp (1988) examine landforms associated with oblique convergence in North Island, New Zealand.

Displaced terranes are starting to generate a significant literature and this has been reviewed by Howell (1985) and Schermer *et al.* (1984). The large number of displaced terranes in the North American Cordillera are considered by Coney *et al.* (1980) and the possible role of oceanic plateaus in orogenesis is assessed by Nur and Ben-Avraham (1983). Ollier and Pain (1988) draw attention to the neglected problem of how displaced terranes relate to landscape development in the context of Papua New Guinea. Examples of specific studies of micro-and mesoscale fault-related landforms include those by Gerson *et al.* (1985), Nash (1981) and Wallace (1977), while Armijo *et al.* (1986) and Molnar *et al.* (1987) provide detailed examples of how such landforms can assist in the interpretation of major tectonic features.

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