Plate tectonics and mountain building

One of the greatest strengths of the modern plate tectonics theory is its ability to explain the origin of virtually all of the present and most ancient mountain belts on Earth. In other words, mountain building (orogeny) stands in strong causal interrelation with the global plate drift pattern. The motor of orogeny is subduction. To enable subduction, a basin floored by oceanic crust must be present. The process of orogeny becomes initiated by subduction of ocean floor and finds its climax in the collision of continents and island arcs. Mountain belts are elongate zones characterized by crust thickened to more than 70 km, in comparison to normal continental crust that is 30–40 km thick. The most classical *style* of orogeny involves continental collision that follows a lengthy period of subduction and completes a Wilson cycle (see below). Ensuing orogeny leads to crustal thickening, deformation, metamorphism, and uplift. This style is called the Alpine style of orogeny. In contrast, orogenic belts that face Pacific-style oceans do not culminate with continent-continent collision, but rather involve long periods of ocean-slab subduction beneath continental margins with repeated episodes of collision that involve island arcs, oceanic plateaus, and microcontinents. This orogenic style is exceptionally rich in volcanic/plutonic production and is called the Cordilleran style of orogeny.

The following discussion overviews the above orogenic styles; however, it is necessary to realize that transitions exist between each style and that many mountain belts have been generated by combinations of both of these. But first we will examine some of the variations in the types of subduction zones and active continental margins that lead to these different styles of orogeny.

Types of active continental margins within orogenic styles

The Alpine and Cordilleran orogenic styles describe long-lasting orogenic cycles that commonly involve numerous phases of orogeny. Each style consists of smaller events that are related to specific geometries of the active continental margin within the larger orogenic cycle. We refer to these as *types* of orogeny. **Figure 11.1** illustrates these types, although it must be emphasized that other types are also possible.

The *island-arc type* of orogeny (**Fig. 11.1a–c**) forms following lengthy periods of subduction activity within an ocean or along its margins; the subduction generates island arcs and backarc basins such as those presently found in the western Pacific region. Subduction in these cases is generally initiated by the high density of old oceanic lithosphere ("spontaneous subduction"; see **Fig. 7.5**). The persistent subduction that accompanies island arc evolution eventually results in closure of the ocean when the subduction velocity exceeds the unilateral spreading rate at the mid-ocean ridge for a long time. During closure, the island arc collides with the approaching passive continental margin and is thrust over it. However, the convergence process is not terminated because the motion of the plates is driven by the global plate drift pattern and oceanic realms are still present to be subducted. The subduction zone typically jumps across the accreted arc into the adjacent ocean realm and a new island arc is built (**Fig. 11.1a**). Following arc accretion and renewal of subduction, the polarity of the subduction zone may change; a new volcanic arc is then built on the collided terraine (**Fig. 11.1b**). When the new oceanic lithosphere is still hot, it may be obducted or overthrust by the plate convergence because it is too buoyant to sink by its own density ("forced subduction"). There may be several intra-oceanic subduction zones in a given region such as it is presently the case around the Philippine Sea Plate and the Molucca Islands (**Figs. 11.1c, 13.1**).

The final closure of all oceanic realms results in collision of two or more continents. The two continent margins and the intermediate island arc systems become overthrust and underthrust, folded, and metamorphosed where they were dragged or pushed into depth. During an orogenic cycle of the island-arc type, new continental crust is created from the long lasting subduction by lengthy magmatic activity. An example for this type of orogeny occurred during the Late Precambrian Panafrican orogen in northeast Africa and the Arabian Peninsula (Ch. 12). However, as was stated above, not all oceans close.

The *Andean type* of orogeny (**Fig. 11.1d**) is represented by and named after the Andes Mountains. Along this active continental margin, mountain belts were generated by long-lasting magmatic

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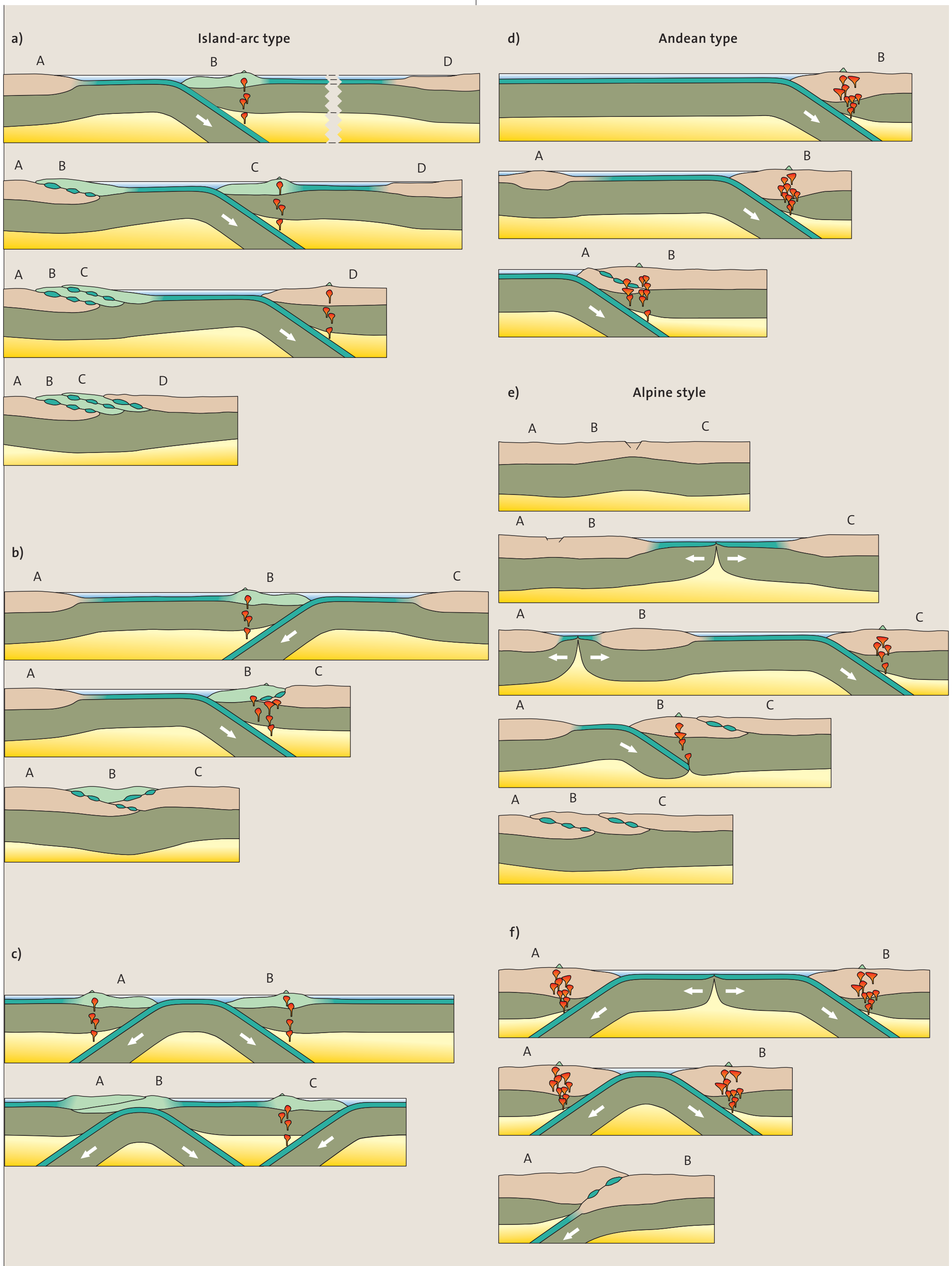


▲ **Fig. 11.1** Schematic profiles illustrating the evolution of orogens as a consequence of subduction and collision. The island-arc type (a–c) creates complex scenarios with opposing or flipping subduction zones – three possible scenarios are shown. The Andean type (d) is characterized by abundant magmatic growth (like the island-arc type), sporadic terranes may contribute to crustal thickening. The Cordilleran style is a combination of the Andean and island arc types but without subsequent continental collision. The Alpine style (e, f) represents the "normal" collisional orogen, where two continent masses collide and intermediate continent splinters (example e: Alps) may be involved. Example e shows a complete Wilson cycle, typical of Alpine-style orogens.

activity, subduction, and terrane accretion, processes also common in the geologic history of the North American Cordillera (Ch. 9). Andean orogenic style is characterized by subduction directly under the continent rather than beneath a fringing island arc system.

The *Alpine style* of orogeny (**Fig. 11.1e,f**) is described by a "Wilson cycle". Such a cycle, named after J. Tuzo Wilson, the discoverer of the transform faults, starts with the break-up of a continent and growth of an ocean. Such oceans may remain limited in size or attain the dimensions of the Atlantic

Ocean. Eventually, the ocean closes during continent-continent collision ending the cycle. The types of active margins present adjacent to the closing ocean basin can be either island arc type, Andean type, or both. Active margins may be present on either or both of the closing continental margins. The size of the ocean between the continents will determine whether crustal growth by magmatism is large, small, or insignificant. Commonly, small continental blocks separate from the passive margins to form microcontinents and eventually, accreted terranes. Such blocks complicate the subduction and collision process. There are many examples of this type of complicated Alpine-style orogeny including the Early Paleozoic Caledonides and the closing of the Iapetus Ocean; evidence for this event is now present in both Europe and North America (Ch. 12). One of the most complicated of all such orogenies involved several phases of opening and closing of the Tethys Ocean. During the Paleozoic and Mesozoic, each closing was marked by collision as numerous peri-Gondwanan microcontinent terranes were welded to Asia. Events culminated with the collision of India to form the Himalayas.

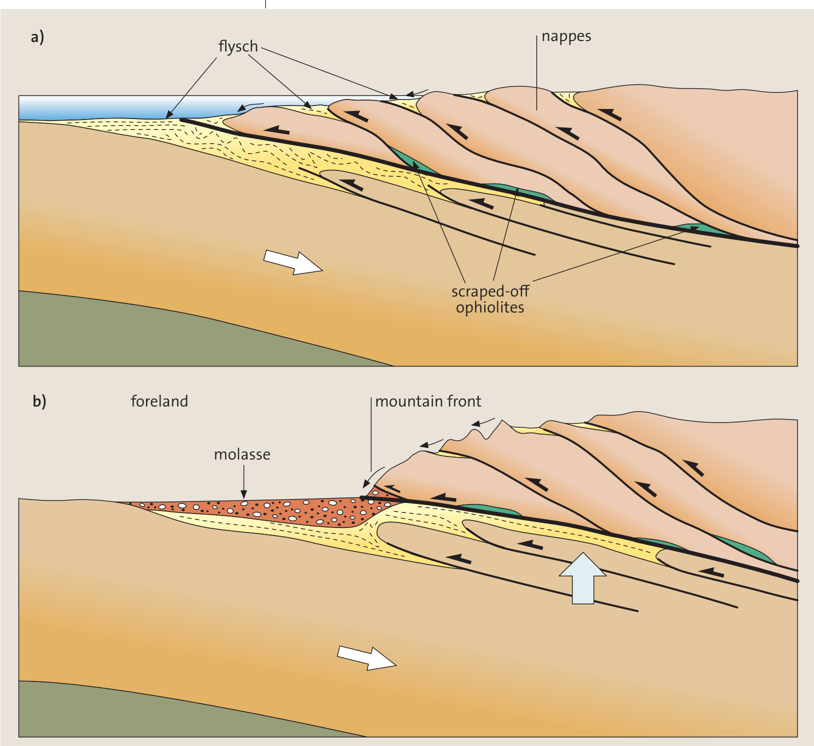
The turnabout in the Wilson cycle from divergent plate movement to continent convergence may be performed by "spontaneous subduction" of old oceanic lithosphere or, in the case of young intermediate oceanic realms, forced by the global plate drift pattern, "forced subduction". The margins of the present central Atlantic Ocean between North America and northwestern Africa consist of old, Lower to Middle Jurassic ocean floor that will be transformed in subduction zones in near geological future. Because of the symmetry of the ocean floor it may happen that subduction zones form at both margins of the ocean (**Fig. 11.1f**). The Atlantic is currently being subducted beneath the Caribbean Plate. In the Penninic Ocean of the Alps, subduction occurred also only on one side and was enforced by the plate drift pattern as Africa closed on Europe (Ch.13). An intermediate continental splinter complicated the closure history (**Fig. 11.1e**).

Cordilleran-style of orogeny involves a prolonged case of both island arc and Andean orogenic types of margins. In the Andean type, an active continental margin persists through extensive periods of time and in the island arc type, the margin involves fringing island arc complexes. The Cordilleran-style orogeny is associated with a Pacific-style of ocean. Pacific-style oceans remain as huge ocean basins over immense periods of geologic time and perhaps, unlike the Atlantic or Tethys, never close. During the Paleozoic, the giant Panthalassa Ocean existed at the location of the present-day Pacific Ocean. Most of the present Pacific Ocean is currently underlain by the Pacific Plate, an almost exclusively oceanic plate; during the Mesozoic, it was underlain by the Farallon Plate, an equally large oceanic plate. However, oceanic crust older than Jurassic is not found in the Pacific Ocean today. Long-lasting and mostly rapid subduction has been compensated by accordingly rapid spreading. Through the subduction of old, dense lithosphere, rapid subduction was stimulated that, in turn, brought about rapid spreading along the mid-ocean ridge thus preventing collisions of the surrounding continents. Only in such a rapid circuit of spreading and subduction, is it possible to form and leave open large oceanic realms; only these circumstances permit oceanic crust to persist great distances from the mid-ocean ridge; rapid spreading of oceanic crust creates plates that are not too old or dense to remain at the surface. Along the edges of large, semi-permanent (Pacific-style) oceans, both Andean and island-arc margins are present. This is the realm of the Cordillian-style orogeny. Therefore, very large volumes of ocean crust are subducted along Cordilleran margins. Orogenic belts in Eastern Australia, East Antarctica, and North and South America represent such regions. They are characterized by extremely long-lived subduction, Australia and Antarctica since the Cambrian and North and South America since the middle Paleozoic. These mountain belts are characterized by immense amounts of volcanic and plutonic rocks and classic foreland fold and thrust belts.

Within large, Pacific-style oceans, the probability is high that seamounts, older (inactivated) island arcs, oceanic plateaus, or isolated continental fragments will drift along and become accreted as terranes to the active continental margin. The consequences are crustal thickening, deformation, and metamorphism along the active continental margin. However, magma generation contributes considerably to crustal thickening – in the Andes, the crust attains a thickness of 70 km. Estimates from the Andes suggest that subduction-related magmatism caused oceanward growth of the continental crust of up to 200 km since 200 Ma (Drake et al., 1982). Numerous intrusive bodies penetrate the crust, including older plutons, and testify to the long lasting magmatic activity. The Cordilleran margin of western North America probably comprises the largest Phanerozoic plutonic and volcanic complex on Earth. In fact, the once continuous Peninsular–Sierra Nevada–Idaho–Coast plutonic complexes that now stretch from northern Mexico to northern Canada may constitute the largest batholith complex on Earth.

Comparisons between Alpine style and Cordilleran style orogenies yield important distinctions between the two. Alpine style defines a Wilson cycle and concludes with continent-continent orogeny; Cordilleran style does not. Alpine orogenies are characterized by basement-involved crustal stacking during continental collision that generates thick nappe sequences; Cordilleran orogenies tend to involve thick sedimentary sequences and sedimentary thrust sheets. Alpine style generates melting within the thickened crust including melting of sedimentary rocks to produce S-type granites; Cordilleran style generates vast primary melts and produces new (juvenile) crust and I-type granites. Alpine style produces bilateral foreland fold and thrust belts and foreland basins that form the sites for thick molasse deposits; Cordilleran style produces a single foreland fold and thrust belt behind the active arc with thrusting directed towards the continent; thrusting generates a single foreland basin, commonly called a retroarc foreland basin. Alpine orogenies tend to destroy older foreland basins by structural, metamorphic, and plutonic processes; Cordilleran orogenies commonly preserve foreland basins, sometimes with little deformation. Both styles commonly have accreted terranes, although Cordilleran orogenies tend to have more oceanic-type terranes and Alpine orogenies tend to have accreted microcontinents. Both styles can contain numerous ophiolite sequences. Classic Alpine-style orogenies include the Caledonian-Acadian (Silurian-Devonian of Europe and eastern North America that closed the Iapetus Ocean), Variscan-Alleghenian (Carboniferous-Permian of Europe and North America that closed the Rheic and related oceans), Urals (Carboniferous-Permian that sutured Europe and Asia and closed the Ural Ocean), and the greater Alpine-Caucasus-Himalayan orogeny (late Mesozoic and Cenozoic that sutured Eurasia and parts of Gondwana and closed the Tethys Ocean). Cordilleran-style orogenies include the greater Nevadan-Sevier (late Mesozoic and early Cenozoic of Western North America), Andean (Mesozoic and Cenozoic of South America), New England (Permo-Triassic of Eastern Australia), and numerous orogenic events that affected parts of Antarctica, South America, South Africa, and SE Australia throughout the Phanerozoic.

▼ **Fig. 11.2** Collision of two continents and evolution from **a)** the flysch stage (flysch sedimentation in deep troughs in front of the upper plate or of individual nappes) to **b)** the molasse stage (filling up of the foreland trough with the debris derived from the ascending mountain range).



Continent-continent collision

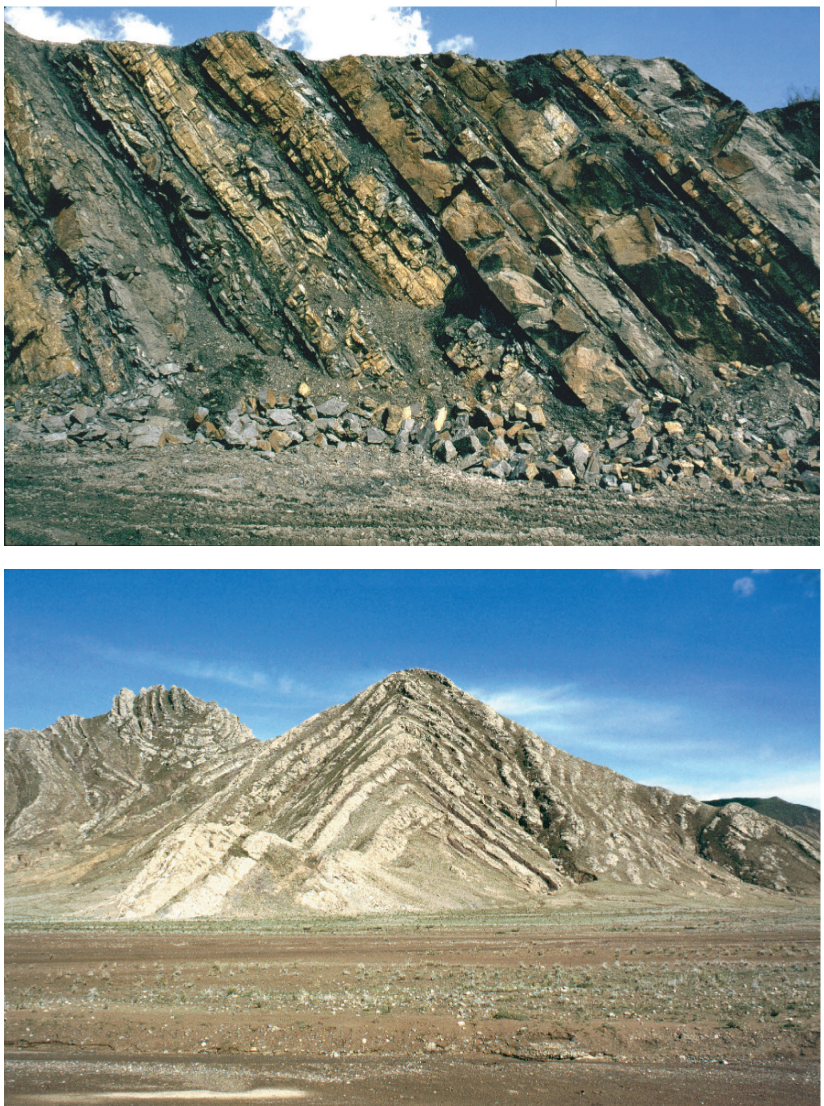
Alpine-style orogeny culminates in continent-continent collision. The collision of continents causes considerable deformation and large-scale overthrusting on the order of a 100 km or more. The condition of the subduction process persists during the early stages of collision. In front of the overriding plate, a deep trench or trough exists that becomes filled with large volumes of turbidites (flysch sediments) derived from the raising nappe pile in the hinterland (**Figs. 11.2a; 11.3**, upper). During this time, the trough is underlain by continental crust of the subducting plate and the flysch sequence becomes deposited on top of the shelf sequence of the passive continental margin of that plate. Flysch generally forms the youngest sedimentary deposits of such a sequence because the sediment pile becomes overthrust beneath the margin of the upper plate in the following stage. Similar but smaller-scale conditions exist in front of individual nappes within the plates (**Fig. 11.2**).

During the stages of plate collision, the axis of the sedimentary trough migrates towards the subducted plate as the nappes advance (**Fig. 11.2**). Initially the top of the sedimentary prism resides below sea level, even at near-abyssal depths. During this time, the trough is filled with deep-water flysch, commonly turbidites, that is derived from the uplands on the upper plate (**Figs. 11.2a; 11.3a**). The flysch accumulates in the main trough between the two plates and also in smaller basins formed between individual thrust sheets along the front of the upper plate. As the main flysch trough migrates over the subducted plate and as the mountain front on the overriding plate rises, sediment input rate increases and the basin fills with sediment. Filling of the trough is enhanced because the thrusting process slows down due to high frictional forces along the thrust planes and consequently less sediment is removed from the trench by underthrusting. Eventually, the top of the sediment interface becomes subaerial and continental depositional systems replace and overlie the marine deposits (**Figs. 11.2b; 11.3** lower). The resulting coarse-grained sediments are dominantly gravels deposited in shoreline and fluvial depositional systems and alternate with fine-grained sediments deposited in lakes or stillwater reaches of rivers. These deposits are called molasse. In humid settings, coal may be intercalated as in the case of the Pennsylvanian molasse deposits of the Appalachian basins of the east-central United States. In arid settings, continental redbeds are formed as in the case of the Devonian Old Red Sandstone of Northwest Europe during the Caledonian orogeny. For etymology of the terms "flysch" and "molasse", see p. 104.

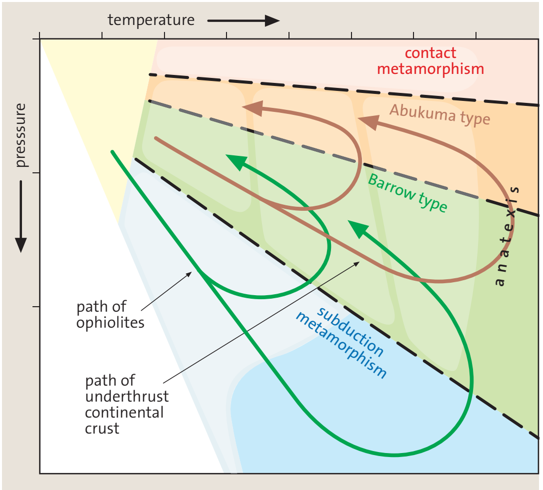
The burial of continental domains during collision and overthrusting not only causes deformation, but also regional metamorphism across large parts of the emerging orogen. The most important factors of metamorphism are (1) the lithostatic pressure, which is determined by the depth of burial and acts with the same magnitude in all directions, (2) the temperature, (3) the directed pressure, which creates a tectonic stress that acts at higher magnitude in the direction of tectonic compression, and (4) the fluid phase, mainly water. During metamorphism the mineral assemblage (paragenesis) of a rock is adapted to the prevailing pressure and temperature conditions by mineral reactions. These two factors are mainly responsible for the paragenesis of the metamorphic rock. Stress, in contrast, causes deformation, rotation, and recrystallization of minerals under preferred orientation – it is therefore responsible for the cleavage, a typical texture of most metamorphic rocks. The fluid phase considerably accelerates mineral reactions and recrystallization – in a "dry" environment that lacks a fluid phase, reactions remain largely incomplete.

During continent collision, large volumes of cool rock from the upper part of the subducting continental margin are brought to increasing depth. Therefore, rocks involed in an orogeny usually initiate metamorphism with conditions of a high pressure/temperature (P/T) ratio as is typical of high-pressure or Barrow-type metamorphism (see **Fig. 7.24**). In an advanced stage of burial, the P/T ratio gradually lowers due to thermal adjustment: the rocks are heated to the ambient temperatures of the given depth after burial has been completed and exhumation started. Therefore, the P/T conditions usually describe a loop (**Fig. 11.4**). During burial, the temperature increase lags considerably behind the normal gradient of 30° C/km because rocks exhibit low thermal conduction; the extreme case is that of high-pressure metamorphism in subduction zones. During exhumation of rocks the temperature gradient increases – the heated rocks are lifted to shallower depth and subsequently cool under delayed conditions. Therefore, as compared to burial, the P/T ratio is considerably lower.

Continent collision commonly leads to anatexis, the partial melting of crustal rocks. The presence of "wet" metamorphosed mudstone and sandstone favors the formation of melts that initiate around 650 °C. Anatexis generates large bodies of magmatic rock, chiefly granites to granodiorites, that are called batholiths; they are surrounded by zones of high-grade metamorphism. The temperatures of the melts usually only slightly surpass the melting temperature of the rock; therefore, the magma bodies cannot ascend larger distances but rather solidify near their place of formation. These rocks



▲ **Fig. 11.3** Flysch sequence in the Eastern Alps (Rhenodanubian Flysch) and molasse sequence in Tibet (Qiuwu Molasse, see **Fig. 7.18**). In the deep-marine flysch beds, layers of sandstone, and mudstone alternate; each layer is deposited by a turbidity current and shows graded bedding (becoming finer grained towards its top). The folded terrestrial molasse beds in Tibet show alternating conglomerate-sandstone (light layers) and mudstone (dark layers).



▶ **Fig. 11.4** Typical pressure-temperature loops from rocks in a collisional orogen. Oceanic crust or continent splinters can be deeply subducted and experience subduction metamorphism; during their ascent they will be overprinted in amphibolite or greenschist facies (green paths). Other parts of continental crust experience pressure-emphasized regional metamorphism (Barrow-type) or even anatexis (partial melting) during burial and reach fields of the Abukuma-type regional metamorphism during ascent (brown paths). During their ascent the rocks are much hotter than during descent at the same depth, because both heating and cooling are slow processes. Compare **Figure 7.24**.

typically classify as S-type granites, formed from sedimentary protoliths (Ch. 7).

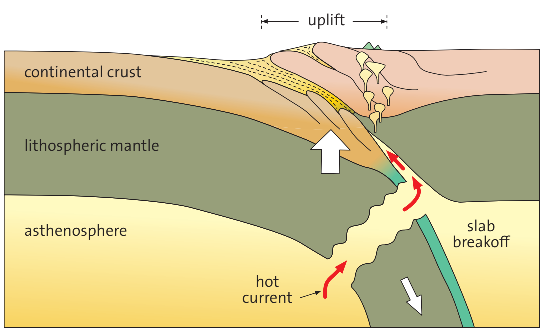
Alpine-type continent-continent collisions are characterized by distinct features (see also above). These include *(a)* large nappes that express large-scale stacking of continental crust; *(b)* broad belts of deformation and regional metamorphism in which the metamorphic history traces a typical pressure-temperature loop; *(c)* the occurrence of ophiolitic sutures that mark the seam of collision and represent remnants of the ocean floor, commonly transformed by high-pressure metamorphic conditions; *(d)* island-arc magmatism. The latter may be of minor importance due to limited duration of subduction of small ocean basins, in which cases the growth of continental crust is also rather limited.

Uplift, erosion, and elevation of mountains

Crustal stacking during continent collision typically generates crustal thicknesses between 50 and 70 km. As a consequence, the newly formed orogen experiences isostatic uplift that is responsible for the morphological feature called a mountain range. However, in many orogens, including the Alps and the Himalayas, uplift of the surface to high elevations was not the immediate consequence of crustal thickening but occurred after a delay of several million years.

During and following the collision process, the subducting continental margin is pulled downward by the attached oceanic slab. The dense oceanic slab forms a counterweight against the thickened and buoyant continental crust. Continued slab pull causes ongoing thrusting of the colliding continent margins and leads to the creation of new nappes and greater thrusting distances; however, new, high mountains do not develop yet. With time and continued subduction, the resistance against compression in the collision zone becomes strong enough so that the dense, heavy oceanic lithosphere in the subduction zone breaks off; this results because rocks possess low tensile strength. The detached slab freely sinks into the mantle (**Fig. 11.5**). This process is called slab breakoff. Slab breakoff causes the thickened continental crust to lose its counterweight and commence isostatic uplift. By analogy, a small boat being pulled down by a heavy anchor will bob up if the anchor is cut free. Isostatic uplift allows mountain ranges to reach great heights.

Climate plays a key role for shaping the mountain range. In arid regions like the Altiplano of the Andes or the Tibetan Plateau of the Himalayan region, denudation by erosion is very slow. Therefore, uplift of the crustal stack generally equals uplift of the surface. Elevated surfaces in arid climate commonly show surprisingly low relief (**Fig. 11.8**, upper). In humid regions erosion rates are accelerated and relief is increased. This is especially apparent where glaciation is present, as ice is a very efficient agent of erosion. Glacial erosion creates deeply incised valleys with strong relief. Although erosion reduces the mean elevation of a mountain range, summit areas, which generally exhibit lower than average erosion rates, commonly increase in elevation due to uplift. The affect of climate applies well to the Himalayas where the southern monsoon generates enormous amounts of precipitation. Here, the erosion rates are considerably higher than in the much dryer Tibetan Plateau that lies in a rain shadow behind the mountain front. As a result, the



◀ **Fig. 11.5** Cross section showing effects of slab breakoff. Breakoff of the subducted, dense oceanic lithosphere after collision is triggered by the buoyancy of the lower-plate continental crust and the decelerating plate convergence (caused by increasing frictional forces). The slab breakoff allows hot asthenospheric mantle to ascend into the newly created space and to cause partial melting. The melts can rise and penetrate the crust of the upper plate. In addition, the loss of the heavy counterweight enables rapid uplift of the mountain range.

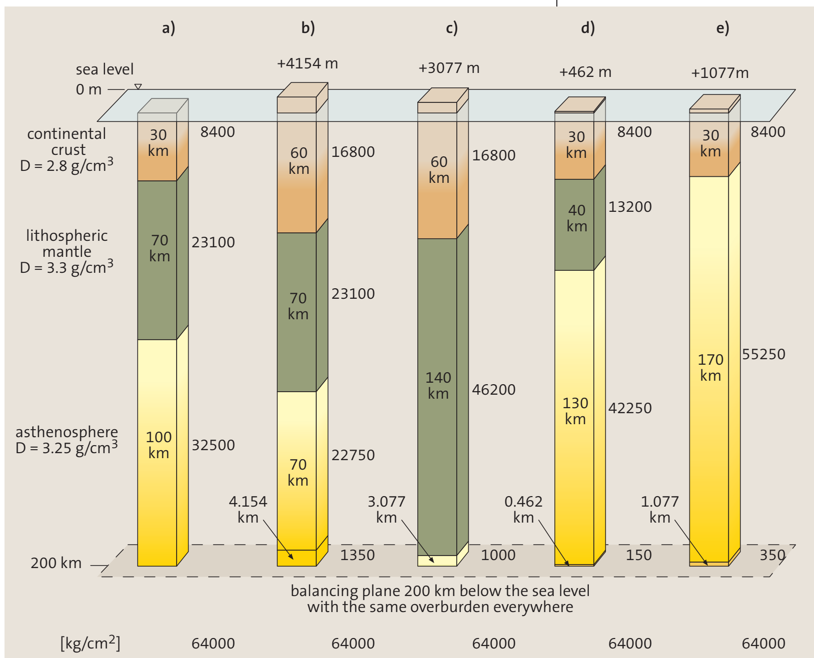
Himalayas show lower mean elevation but higher summits than the Tibetan Plateau (**Fig. 11.8**, lower; see also below). Mountain topography is complex and reflects many factors including rock type, structure, isostatic uplift, and climate.

According to the principle of isostasy, a body floating in a denser medium will stand higher in proportion to its thickness. For example, a thick iceberg will attain higher elevation than a thin one. With respect to continental crust, this means that thick crust is characterized by high mountains or plateaus and thin crust is characterized by lowlands or continental shelves below sea level. If isostatic equilibrium is attained, the weight of any rock column of a given area that rests on a theoretical horizontal plane at depth is equal (**Fig. 11.6**). Because the configuration of the Earth below the lithosphere is rather uniform, a plane of equilibrium can be defined at or slightly below the base of the lithosphere. Examples, as illustrated in **Figure 11.6**, show the isostatic influence of crustal thickening during an orogenic cylce and changes in the thickness of the lithospheric mantle due to syn-orogenic thickening or post-orogenic delamination.

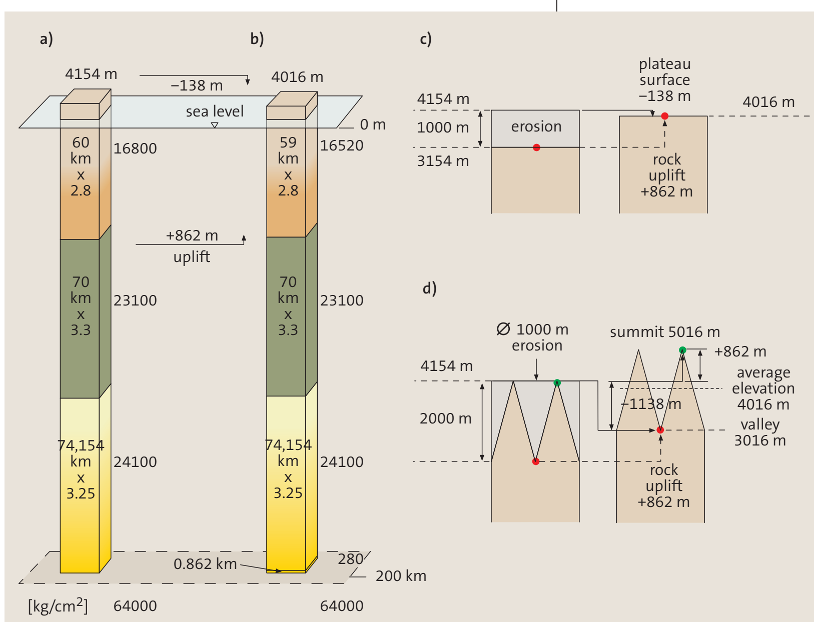
The models presented in **Figure 11.6** assume the following: (1) Continental crust has an average density of 2.8 g/cm^3; the rocks of the upper and middle crust generally show slightly lower density but the more basic rocks in the deep crust have higher specific weight. (2) The peridotites of the lithospheric mantle have a density of 3.3 g/cm^3. (3) Peridotites of the asthenosphere, which is hotter and contains minor amounts of melt, have a density of 3.25 g/cm^3. Although the actual densities may slightly deviate from these values, this would not change the results significantly.

Column "a" assumes an initial continental lithosphere with 30 km of continental crust and 70 km of lithospheric mantle. Therefore, a weight of 64,000 kg per square centimeter would rest on a theoretical plane at 200 km depth (**Fig. 11.6a**). Such a rock column would be at sea level, actually slightly above (which is neglected here for simplicity).

Column "b" shows a crustal thickness that was doubled during continent-continent collision to 60 km; the weight of the rock column is only

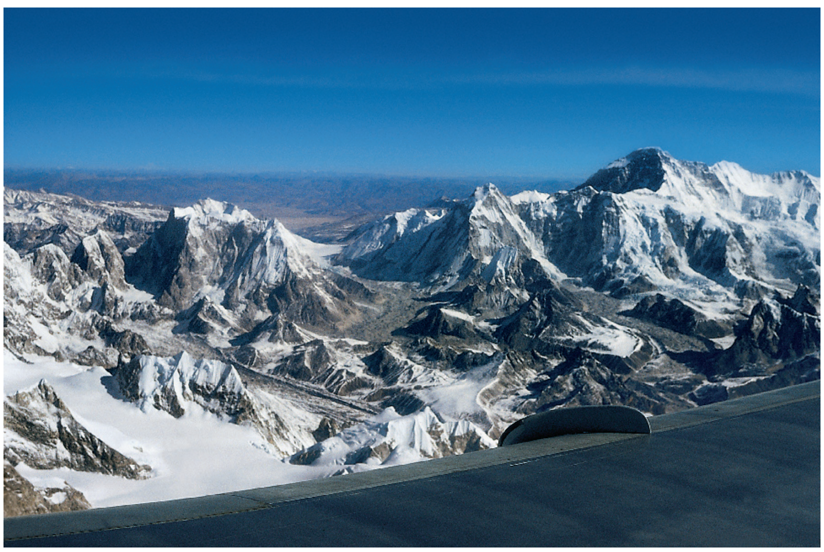


▲ **Fig. 11.6** Effects of changes in thicknesses of continental crust and lithospheric mantle with respect to surface elevation. A balancing plane is assumed at 200 km depth on which the pressure per unit of area (here taken as kg/cm^2) must be equal, when equilibrium is attained. Thickening of the relatively light crust and thinning of the dense lithospheric mantle both cause uplift of the rock column. **a)** Initial conditions with a 100 km-thick lithospheric plate containing continental crust. **b)** Doubling of the crust during orogeny. **c)** Doubling of both crust and lithospheric mantle (possible under mountain ranges). **d)** Delamination of a part of the lithospheric mantle as compared to the initial condition. **e)** Complete delamination of the lithospheric mantle.



▶ **Fig. 11.7** Effects of erosion and relief on mountain uplift and topography. Erosion reduces the mean elevation but the uplift of peaks follows the pace of valley incision. **a)** Conditions shown in **Fig. 11.6b**. **b), c)** Erosion of 1000 m causes reduction of surface elevation by 138 m und rock uplift of 862 m. **d)** When erosion concentrates in valleys and spares the peaks, the latter will rise 862 m, whereas the mean elevation will be reduced by 138 m. The volume of erosion in c and d is the same.

▼ **Fig. 11.8** Photographs contrasting Tibetan Plateau with Himalayas. The Tibetan Plateau (above) forms a flat depression with limited relief, only interrupted by individual mountain ranges (NW Tibet, photo taken on ground at 5100 m). The Himalayas (below, aerial photgraph) achieved their high relief due to the intense monsoonal precipitation (Cho Oyu, 8200 m, to the right; Tibetan Plateau in the background).



62,650 kg/cm^2 at 200 km depth because thick and buoyant continental crust replaces more dense asthenoshpere. This mass deficiency is balanced by buoyancy of the rock column that is provided by lateral inflow of asthenosperic material at depth. The rock column regains isostatic equilibrium, when it rises 4154 m because the weight at 200 km depth is again 64,000 kg/cm^2 (**Fig. 11.6b**). Therefore, the now 204.154 km-high rock column would stand 4154 m above the original surface elevation, sea level. Over a large area, this scenario could result in a high plateau, depending on the nature of climate and erosion as discussed above.

Column "c" illustrates both thickened crust and lithospheric mantle, each doubled from column "a". Such a geometry can be generated during an orogenic cycle – although in many orogens the thickness of the lithospheric mantle is not well constrained; it may attain a thickness of 140 km or more. In this case, uplift of the rock column to 3077 m would be sufficient to establish isostatic equilibrium (**Fig. 11.6c**). The lithospheric mantle, slightly denser than the asthenospheric mantle replaced by it, compensates for a smaller part of the mass deficiency caused by the thickened crust.

Columns "d" and "e" demonstrate the effects of delamination of the lithosphere or thermal transformation of lithosphere to asthenosphere. Both processes will cause uplift. Either part of or the whole lithospheric mantle may delaminate and sink freely into the deeper upper mantle. Equally plausible is a temperature increase that transforms the whole or the lower part of the lithospheric mantle into asthenospheric mantle as both consist of peridotite. In these cases, dense lower lithosphere becomes replaced by slightly less dense asthenospheric material. Compared to the original scenario of column "a", in "d", a 30 km-thick crust is maintained but lithosphere delamination or thermal transformation results in the lower 30 km of the lithosphere being replaced by asthenosphere and an uplift of the rock column of 462 m results (**Fig. 11.6d**). The uplift effect is only one ninth of that in the case of doubling crustal thickness (column "b"). If the whole lithospheric mantle is replaced by asthenosphere, as shown in column "e", uplift will be 1077 m (**Fig. 11.6e**).

Each case shown in **Figure 11.6** results in mass deficiency that causes surficial uplift. However, uplift is a slow process and will partly be compensated by erosion. Due to the very low erosion rates in arid climates, surface uplift will nearly equal the uplift of the rock column (rock uplift). This is valid for the Altiplano or the Tibetan Plateau. In the Tibetan Plateau, there is some precipitation and erosion, but drainage is internal as no river breaks through the margins of the plateau. Therefore, the mass balance remains constant. The mean elevation of 5000 m corresponds to a crustal thickness of 60–70 km and a relatively thin lithospheric mantle, as has been demonstrated beneath parts of the plateau. The thin lithospheric mantle probably formed by thermal transformation of lithospheric mantle to asthenospheric mantle due to hot convection streams.

When isostatic equilibrium is attained, uplift will cease. Erosion reduces the thickness of the relatively light crust by removal of material at the surface, again leading to mass deficiency in the rock column. To balance this loss of mass, the rock column will again be uplifted, but the new surface elevation, or mean elevation in case of a relief, will be lower than before because the eroded crust is compensated by dense asthenosphere at depth. Erosion of 1000 m of crust results in a decrease in surface elevation of 138 m and rock uplift of 862 m as shown in **Figure 11.7a–c**. The formula is

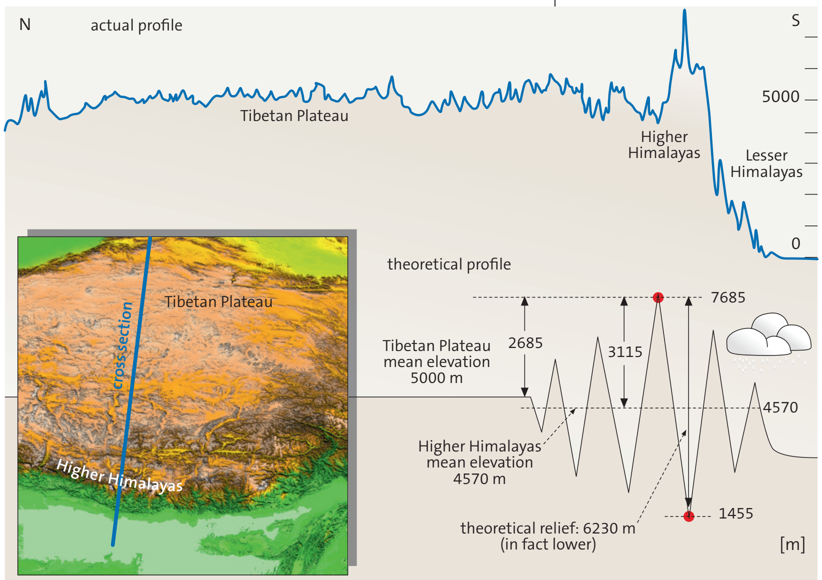
rock uplift (862 m) = erosion (1000 m) + surface uplift (–138 m),

whereby surface uplift is negative. The relation between 138 m decrease in surface elevation and 1000 m erosion (i. e., 13.8 %) remains constant, if the same densities for the rock units are assumed (see example of Himalayas below, **Fig. 11.9**).

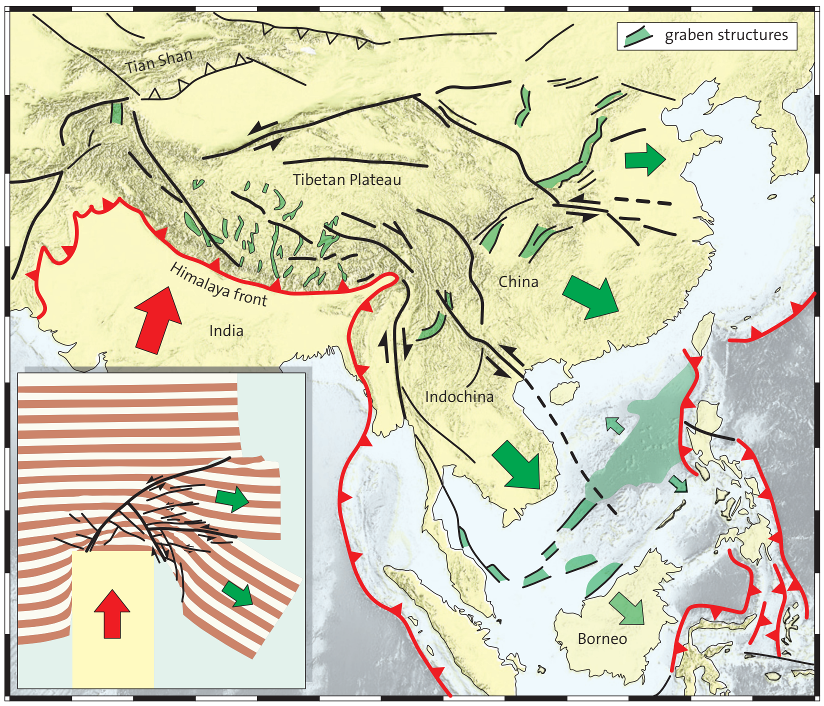
However, erosion does not result in a planar surface, but rather causes valley incision and creates a relief that can be considerable, depending on uplift and elevation. This means that valleys, places where erosion is greatest, become deeply incised whereas summit areas, where erosion is low, will rise relative to their initial position. When 2000 m-deep valleys form, the summits do not experience erosion, and erosion averages 1000 m as in the above example: rock uplift will be again 862 m accompanied by a decrease in *mean* elevation of 138 m. To summarize, the valley floors will eventually become 1138 m deeper than the surface in its original position, whereas the peaks will rise 862 m (**Fig. 11.7d**). For simplicity, idealized V-shape valleys have been ilustrated, which although not realistic, leads to an acceptable conclusion. When the valleys are more U-shaped, the valley floors become less deeply incised. Also, it was assumed that erosion is absent on the peaks, also not completely true in most cases. Therefore the actual resulting relief will be less than in the example shown. Nevertheless, the example illustrates how erosion and relief exert decisive influence on the evolution of the geomorphology of a mountain range. As erosion lowers mean elevation, summits rise and considerable relief is generated.

In the case in the Himalayas, the heavy monsoon rains, increased during the rise of the Himalayan mountain range, can explain how a landscape typical of the Tibetan Plateau was transformed into the present High Himalayas along its southern margin (**Fig. 11.8**). Analysis of a digital elevation model, a digitized topographic map, shows that the mean elevation along the shown profile (**Fig. 11.9**) is 5000 m in the Tibetan Plateau and 4570 m in the High Himalayas; these values will be slightly different in other profiles.

To explain the lower mean elevation of the High Himalayas (–430 m) relative to the Tibetan Plateau, an average of 3115 m must have been eroded (**Fig. 11.9**). Isostatic balancing generates a rock uplift of 2685 m (2685 = 3115–430; the relation of these figures is the same as in the example of **Fig. 11.7d**, each multiplied by 3.115). Assuming idealized V-shaped valleys and no erosion at the peaks,



▲ **Fig. 11.9** Topographic profile across the Tibetan Plateau and the Himalayas, and theoretical elevation profile illustrating the relief formation in the Himalayas by monsoonal precipitation. The theoretical profile follows the scheme shown in **Fig. 11.7d**. The evaluation of the digital elevation model (map) shows that the mean elevation in the High Himalayas is 430 m lower than in the Tibetan Plateau. Evaluation by B. Székely.



▲ **Fig. 11.10** Escape of crustal blocks in southeastern Asia (Tapponnier et al., 1986). The escape motion towards the Pacific region is caused by continued northward migration of India. It is accompanied by crustal extension as revealed by graben structures and metamorphic domes (shown in green). The SE motion of Southeast Asia is taken up in the adjacent subduction zones. Similar structures to the NE are greatly reduced in size because there is no room for "escape". The box shows an analog experiment with plasticine where the pushing indenter and the escaping wedges duplicate the structures in southeastern Asia.

the peaks would attain an elevation of 7685 m, the valley floors 1455 m. The relief would amount to 6230 m. According to the simplifyed assumptions, the actual relief is expected to be slightly less. Because the Tibetan Plateau also bears some mountain ranges exceeding 6000 m in elevation, even higher peaks have to be expected in the Himalayas.

Despite the simplification of this example, the predicted results in the Higher Himalayas are perfectly met. The 7000 to 8000 m-high peaks are located close to valleys that are incised to less than 2000 m above sea level. Therefore, the widely accepted hypothesis that the morphological difference between Tibetan Plateau and High Himalayas is largely due to the erosive processes caused by the powerful monsoon rains, is strongly supported by the above calculations.

Of course, crustal thickening, uplift and erosion are processes that overlap and act simultaneously. The results are rising mountain ranges with high elevation and relief. Only in places where erosion is limited, will high plateaus form.

Collapse and crustal escape

Rapid exhumation of deeply buried rocks is not necessarily coupled with strong surface uplift as is the case in the Himalayas. Burial during collision causes heating and softening of rocks. Rocks rich in quartz, one of the most common minerals in continental crust, become ductile at ca. 300 °C and 10 km depth; therefore, they will react to stress by plastic deformation, not by fracturing. They will lose much of their strength and undergo plastic flow at depths of more than 10 km without becoming melted. This process is called thermal weakening. Through this process, the heated crustal stack will respond to the gravitational instability caused by orogenic thickening and react by lateral flow that is driven by gravity forces. The result is horizontal crustal extension and is called "gravitational collapse". It enables deeply buried crustal material to approach a near-surface position very rapidly; such is the case in metamorphic domes (see **Fig. 3.19**). The overlying rocks of the upper crust are not removed by erosion, but rather become considerably thinned and extended tectonically resulting in rapid exhumation of the deeper rock units; this is the process of tectonic erosion or denudation (see **Fig. 13.9**). Because collapse is always a consequence of crustal thickening, younger extensional structures overprint older compressional ones and may partly or even completely obliterate them. This explains an apparent paradox of many orogens – extensional structures dominate compressional ones. Such is the case in the Basin and Range Province of western North America.

Another consequence of collision in many orogens is the lateral escape of crustal blocks. This feature was first described in southeastern Asia. Here, the eastward motion of crustal wedges along subvertical faults compensates for the ongoing post-collisional compression in the Himalayas. A prerequisite for this process – as is also the case for the gravitational collapse – is the availability of a "free space" towards which the blocks can move. East of the Himalayas and the Tibetan Plateau, crustal blocks greater than one million square kilometers in area escape towards the east where they are able to move freely against the convergent plate margins in the western Pacific region (**Fig. 11.10**). Similar blocks are also known from Asia Minor (**Fig. 8.10**) and the Eastern Alps (**Fig. 13.9**). In the Eastern Alps, the combination of gravitational collapse and lateral escape led to a complex process described as "lateral tectonic extrusion" (Ch. 13; Ratschbacher et al., 1991b).

Thick continental crust becomes gravitationally unstable because of the heating and softening of the rocks as described above. Therefore, crustal thickening to greater than 70 or 80 km cannot occur because the rocks become too weak to support the crustal stack. Ongoing thickening by compression is prevented by collapse or escape. Collapse is the more efficient process insofar as it leads to crustal thinning (horizontal stretching compensates for vertical shortening), whereas during escape the crustal thickness remains constant (horizontal stretching compensates for perpendicular horizontal compression). Incidentally, an initial stage of gravitational collapse is also occurring in Tibet; this is in addition to the escape motions of the crustal blocks. Numerous young, active graben structures (**Fig. 11.10**) and several metamorphic domes indicate west-east extension under vertical shortening. Due to the size and elevation of the Tibetan Plateau, a dramatic change can be expected in near geologic future; it would not be unexpected if the region evolved to a state similar to that of the North American Basin and Range Province.