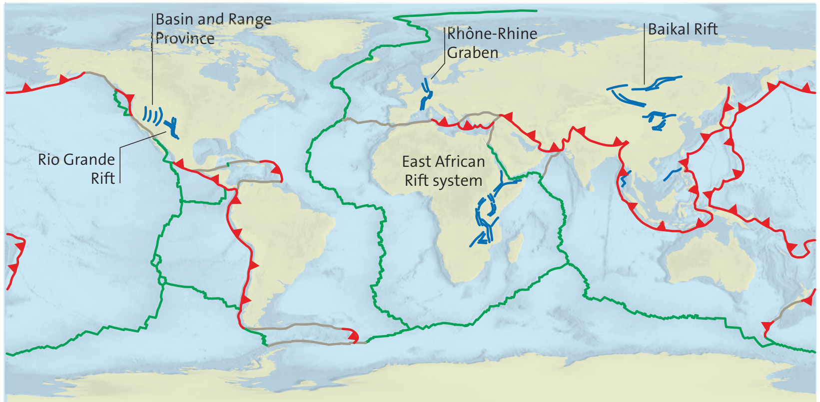
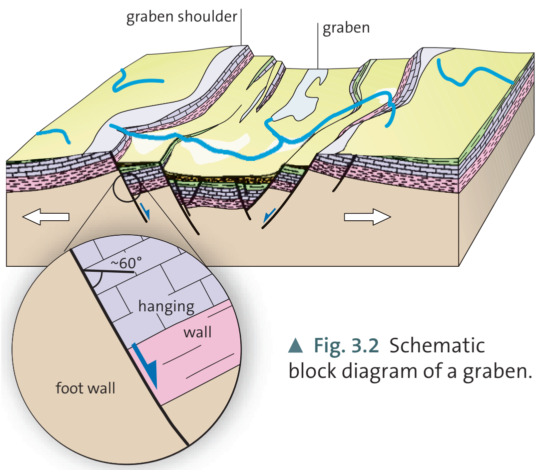
Continental graben structures

A continental graben structure or rift is a narrow, elongated, fault-bounded structure in the Earth's crust (**Fig. 3.1**). Grabens consist of a central axial depression flanked by steep walls and elevated shoulders that plunge steeply into the rift axis and slope gradually towards the exterior (**Fig. 3.2**). The most famous example is the East African rift system. Rift systems may be cut and apparently offset by transform faults; examples include the Upper Rhine Graben in Central Europe and its southern continuation in the Bresse and Rhône grabens (see below). Graben structures occur in regions where the crust and lithospheric mantle are extended and thinned (**Fig. 3.3**). Broad regions of extension are typically expressed by numerous grabens and intervening higher horst blocks such as the Basin and Range Province in western North America. Graben systems also occur in oceanic crust along mid-ocean ridge systems and will be discussed in Chapter 5.

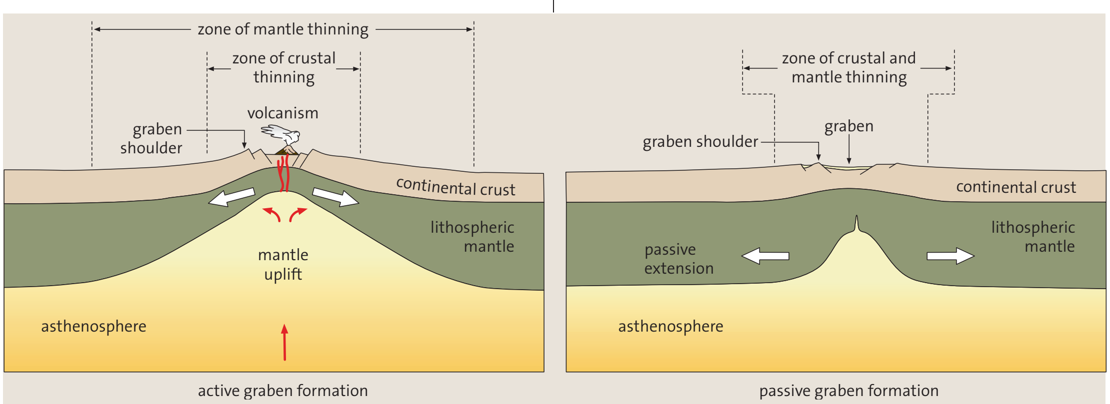
The amount of extension across a graben varies considerably ranging between approximately 5 km across the Upper Rhine Graben to 50 km across the Rio Grande Rift in New Mexico. The brittle extension, generated by fracturing associated with earthquake activity in the upper crust, extends downward to a depth of approximately 15 km. At greater depths, ductile flow occurs without fracturing the rocks; rather, deformation takes place as solid-state plastic flow. Graben subsidence is accommodated along normal faults that dip towards the central graben axis at angles of 60 to 65° (**Fig. 3.2**): the hanging wall, the block located above any point of the fault plane, moves downwards with respect to the foot wall and causes the subsidence of the graben. Normal faulting is linked to horizontal extension orthogonal to the graben axis.



▲ **Fig. 3.1** Tectonic map of Earth showing large, young graben systems.



▲ **Fig. 3.2** Schematic block diagram of a graben.



◀ **Fig. 3.3** Main characteristics of active and passive graben systems (Condie, 1997).

W. Frisch et al., *Plate Tectonics*, DOI 10.1007/978-3-540-76504-2_3,

In continental settings, as the lithosphere extends, the asthenosphere tends to rise (**Fig. 3.3**) and heat-flow rate increases; as a consequence, melting in the uppermost asthenosphere or overlying lithospheric mantle may occur. The melts penetrate the crust and feed volcanoes at the surface or form magma chambers at depth. Because the magmas are derived directly from the mantle, they are basaltic in composition, hence the close association of basaltic volcanism and graben rifting. However, when magmas are trapped at depth and accumulate in magma chambers, they potentially undergo additional processes that result in change of magma composition. Assimilation of adjacent continental crust and magmatic differentiation by removal of mafic minerals that have high melting points and sink to the bottom of the magma chamber produce intermediate to granitic melts. These various magmatic processes explain why many rift areas are associated with volcanism and plutonism of various compositions.

Active and passive graben structures

Based on the relations between topographic expression and method of formation, Condie (1997) defined two classes of grabens, active and passive (**Fig. 3.3**). *Active grabens* are generated by upwelling of the asthenosphere, commonly over hotspots; the overlying mantle lithosphere and crust respond to this process and both are thinned as a result. The mantle lithosphere and lower crust deform plastically and the upper crust is faulted to form the graben structure; both are thinned and basaltic volcanism is generated. Extension of crust at an active graben structure is much wider in the deeper ductilely reacting part of the crust and the lithospheric mantle than in the brittle upper crust (**Fig. 3.3**; Thompson and Gibson, 1994). The wide zone of the asthenospheric doming causes the bulge of the Earth's surface at active rifts to also be broad, commonly several hundred kilometers wide.

At *passive graben structures*, extensional forces are the primary cause. Initially, the extension is limited to the narrow zone of the rift, both in the deeper crust and in the lithospheric mantle (**Fig. 3.3**). This process can result in the complete tearing off of the lithospheric mantle which then leads to asthenospheric material rising to the base of the crust (Turcotte and Emerman, 1983). The surficial bulge is thus restricted to the narrow graben zone and thermal uplift of the rift shoulders is reduced as is basaltic magmatism. However, extension of the lithosphere may also lead to a wider updoming of the asthenosphere and the lithosphere above. Thus a passive rift may change into an active rift system and the passive stage is no longer detectable. Although most present graben systems seem to be active rifts, it is assumed that both processes, updoming and crustal extension, act together. The primary reason for the graben formation is thus an academic chicken-and-egg question. Nevertheless, strong updoming of the asthenosphere causes strong magmatic activity and wide graben shoulders.

Symmetric and asymmetric crustal extension

Crustal extension is believed to occur in two different modes symmetric (McKenzie, 1978) and asymmetric (Wernicke, 1981); models have been proposed for each. The *symmetric model* is based on many present graben systems (**Fig. 3.4a**). It assumes symmetric, brittle extension of the crust along normal faults in the upper 10 to 15 km, and ductile deformation at depth. Both the crust and lithosphere thin accordingly. Crustal thinning and brittle deformation cause the surface of the Earth to subside and generate the graben morphology .

If a 30 km-wide strip of 30 km-thick crust is stretched by 5 km, the resulting stretched crustal section has thickness of only 25.7 km. Such a situation is approximated in the southern Upper Rhine Graben. Assuming an original thickness of 100 km for the total lithosphere, the initial 30 km-wide tract is reduced in thickness to 85.7 km following stretching. However, the ascent of hot asthenosphere causes the lower part of the lithospheric mantle to be transformed into asthenosphere. The lithosphere-asthenosphere boundary is defined by thermal and state properties of approximately 1300 °C; it is not a material-based boundary. Therefore, lithospheric mantle can be transformed into asthenosphere by an increase in temperature and vice versa.

The original bulge of the surface, caused by a hot, relatively light bulge of asthenospheric mantle material, leads to erosion at the graben shoulders, a process that also results in a reduction of thickness of the crust. Thermal subsidence is developed after the heat source disappears and the mantle bulge cools and increases in density. The area of subsidence broadens because the area of mantle uplift is generally two to three times wider than the graben structure. Therefore, old inactive graben structures may have morphologically unobtrusive shoulders.

The *asymmetric model* of graben formation was initially developed for the Basin and Range Province but also applies to some rifts associated with the formation of passive continental margins. Asymmetric grabens are characterized by a gently dipping master fault, termed a detachment fault, that cuts at low angles through the crust from one flank of the graben down to the base of the lithosphere (**Fig. 3.4b**). The overriding upper plate of the detachment is characterized by steeply inclined normal faults that form in response to the extreme amount of brittle crustal extension, in some cases greater than 200 %.

In asymmetric grabens, the crust of the upper plate is extended and thinned at a different location from that of the lithospheric mantle, the lower plate of the detachment (**Fig. 3.4b**). This asymmetry gives rise to the following morphology: above the area of crustal thinning, the surface subsides because light crustal material is replaced by denser mantle material; above the area of lithospheric mantle thinning, the surface bulges because lithospheric mantle is replaced by slightly less dense, hotter asthenospheric material. In some cases, the bulge brings lower crustal material rapidly to the surface.

The asymmetric model has not been identified in a simple graben system (Roberts and Yielding, 1994), but rather has been used to explain wide areas of extension such as the Basin and Range Province in the western USA. Such asymmetric areas of extension are connected to the uplift of metamorphic domes that will be discussed below.

In asymmetric rift zones, the zones of crustal and lithospheric-mantle thinning overlap but are off set whereas in symmetric rift zones they coincide. However, subsidence by crustal thinning is much greater because the difference in density between crust and mantle is much greater than that between lithospheric and asthenospheric mantle.

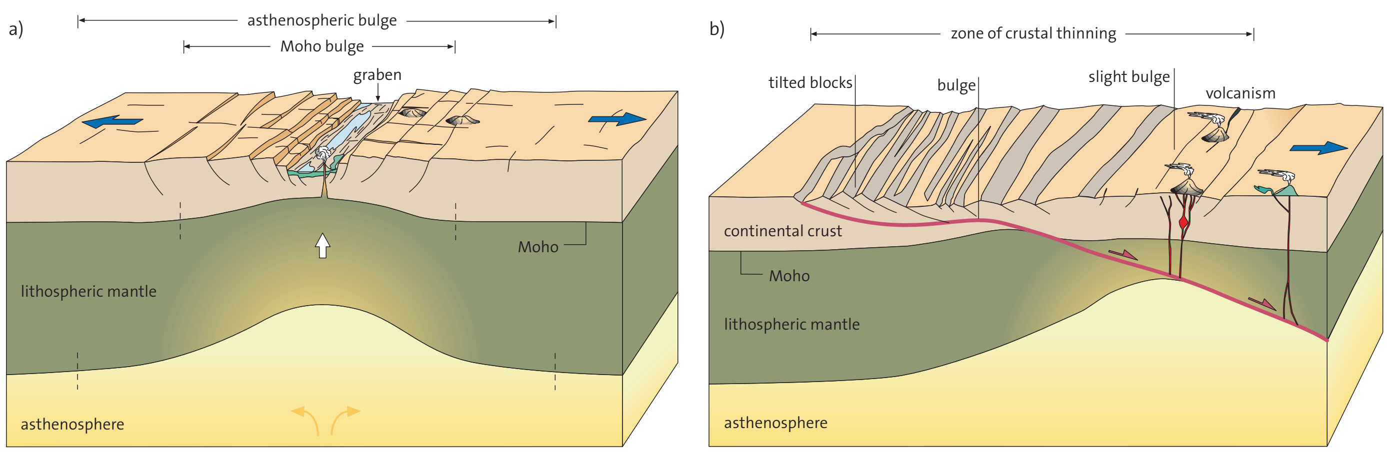
Sediments and ore deposits in graben structures

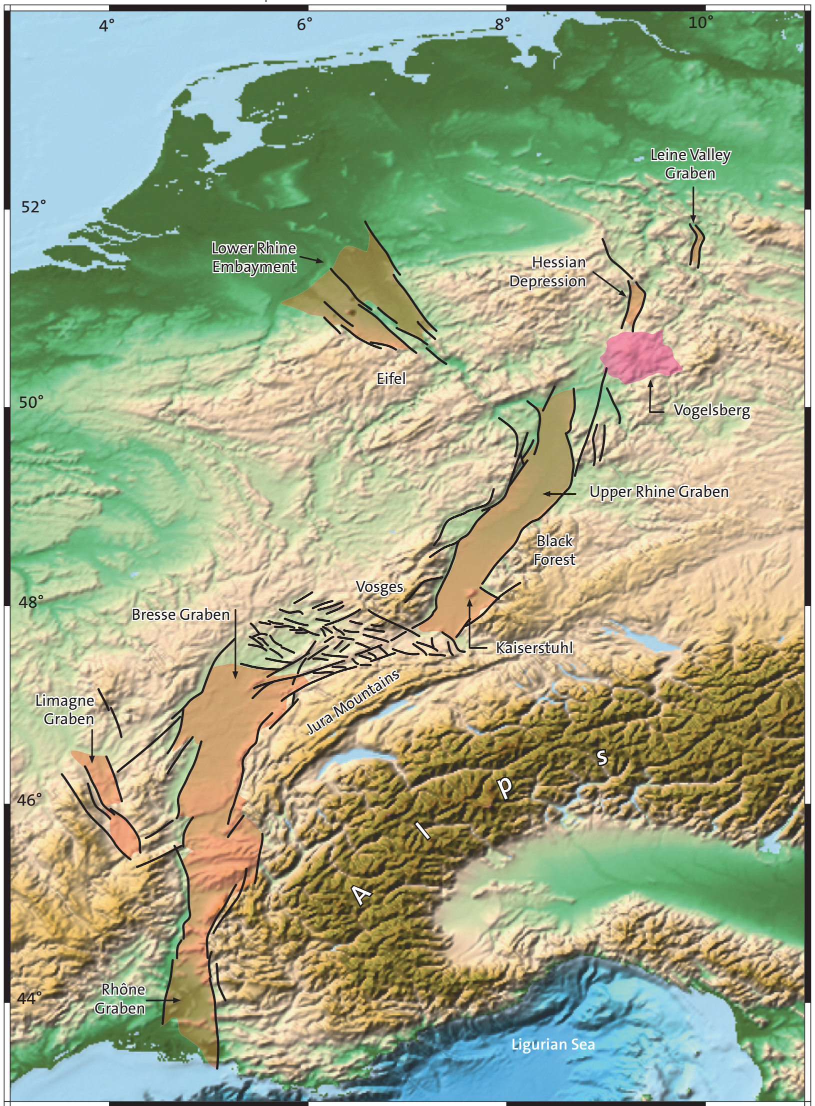
Typically sediments in graben structures are characterized by immature terrestrial deposits that are deposited by rivers that source the steep flank of the graben shoulder. Immature sediments are characterized by an abundance of mineral grains and rock fragments. Normally such minerals are readily weathered prior to reaching the site of deposition but because of the steep topography and short transport distance, they survive the sedimentary cycle in grabens. Many fluviatile sediments in graben structures are mostly composed of conglomerates rich in rock fragments and arkoses (sandstones that contain abundant feldspar) and have a relatively low percentage of quartz. Lacustrine deposits are rich in clays and, under arid or semi-arid conditions, saline sediments. Saline lakes, for instance, occur in the East African graben system.

Graben structures may also come under marine influence. Marine sediments in graben structures are mostly mudstone, marl (limy mud) or limestone. Strong evaporation in arid climates where partly or completely isolated basins fill with seawater leads to concentration of salt in the water followed by precipitation of salt. In the Upper Rhine Graben, marine ingressions are indicated by salt deposits. If a graben evolves into a narrow ocean like the Red Sea, saline deposits are typically preserved at the base of the marine sedimentary sequence.

Petroleum and natural gas are important deposits in some continental rift systems. Restricted water circulation in a narrow graben sea may lead to benthonic anoxic conditions where free oxygen is sequestered by the bottom fauna. Lack of oxygen in the lower part of the water column leads to oxygen-poor sediment which in turn prevents decomposition of organic matter. This generates an enrichment of organic material in the sediment and results in characteristic dark gray or black colors. Basin subsidence lowers organic-rich sediment into the so-called petroleum window, a temperature range between approximately 80 and 170 °C. Here petroleum forms by complicated reactions involving the organic matter. At temperatures over approximately 150 °C, gas deposits are formed. Oil shales of Messel that originated in a maar funnel (a volcanic penetration tube) within the subsiding Upper Rhine Graben provide a good example of sapropelic (organic-rich) sediments formed in an isolated basin. The incredible fossils preserved at Messel construe a unique deposit of global importance (UNESCO World Heritage Site) concerning the life of the Eocene.

▼ **Fig. 3.4** a) Symmetric and b) asymmetric model for the evolution of a graben system. The asymmetric model also explains the early-stage evolution of metamorphic core complexes. The "Moho" (Mohorovičić discontinuity) is the boundary between the crust and mantle.





▲ **Fig. 3.5** Map of the European graben system showing the relations between the Rhône Graben, Bresse Graben, Upper Rhine Graben, and Lower Rhine Embayment.

Volcanism in graben structures

Magmatic rocks that form in graben structures are typically alkaline – they have an excess of alkalis (Na_2O, K_2O) compared to the content of silica (SiO_2) or alumina (Al_2O_3); alkaline rocks with deficiency in silica are also termed "undersaturated in silica". Alkaline magmas primarily develop from lithospheric mantle that undergoes a small amount (mostly less than 10 %; Wilson, 1989) of partial melting (see Ch. 6, 7). However, tholeiitic magmas, which reflect a higher portion of partial melting (mostly more than 15 %), are also common in graben systems. They accompany a rapid rate of extension of the lithosphere, especially where associated with hot spots. Rapid extension increases the rising and melting of hotter asthenospheric mantle rocks (Ch. 6). Mid-ocean ridge tholeiitic basalts are formed in areas of rapid extension (Ch. 5), another indication that tholeiitic basalts are more important in areas that undergo strong extension of the lithosphere accompanied by increased upwelling and melting of asthenospheric material.

In graben systems such as the East African Graben or the Rio Grande Rift, the alkalinity of melts increases outward from the graben axis towards the rift shoulder. This indicates that melt formation is highest below the graben axis where tholeiitic magmatism is favored. Other graben systems such as the Cenozoic Kenya Graben and the Permian Oslo Graben show a decrease of alkalinity with time that indicates increasing rates of extension and melt formation during the evolution of the graben system (Condie, 1997). The East African graben system shows a shift from alkaline basalts and differentiates in the south at Tanzania and Kenya to tholeiitic basalts to the north in Ethiopia. This pattern parallels the increasing rate of extension from south to north (see below).

Volcanism in graben systems can be bimodal. In the northern Rio Grande Rift, tholeiitic basalts (basic, SiO_2 content of about 50 weight-%) occur beside rhyolites (acidic, SiO_2 content of about 70 weight-%). Intermediate rocks with SiO_2 contents in between are missing, however. This is not explainable through simple differentiation of an original basaltic magma. The East African Rift is dominated by alkali basalts (SiO_2 content less than 50 %), phonolites (about 55 % SiO_2, but very high content of alkalis: Na_2O along with K_2O about 12–14 %), and trachytes (about 65 % SiO_2 alkalis about 10-12 %). Phonolites develop by differentiation from alkali basalts that are substantially undersaturated in silica; however, trachytes develop from less strongly undersaturated alkali basalts. Carbonatites also occur in rift systems. These are carbonate rocks that are derived directly from the Earth's mantle and are composed of calcite or dolomite with an accompanying extremely low content of SiO_2 of only a few percent.

The generation of basaltic magmas occurs in the mantle whereas acidic magmas are generated in continental crust or by mantle melts which are, in most cases, strongly influenced by continental crust. Acidic magmas occur in graben regions of greater crustal extension and higher, continuous magmatic activity. Igneous rocks include basalts of slightly alkaline or tholeiitic composition, and significant volumes of acidic volcanic rocks; absence of intermediate rocks indicates a clear bimodality (Barberi et al., 1982). Following intrusion into the continental crust, the primary basaltic mantle melts generate acidic crustal melts with their enormous heat. This explains the bimodal distribution of magmatism. In contrast, graben systems that display little crustal extension are characterized by low magma production, interrupted volcanic activity, and strongly silica-undersaturated alkali basalts; intermediate and acidic rocks are rare.

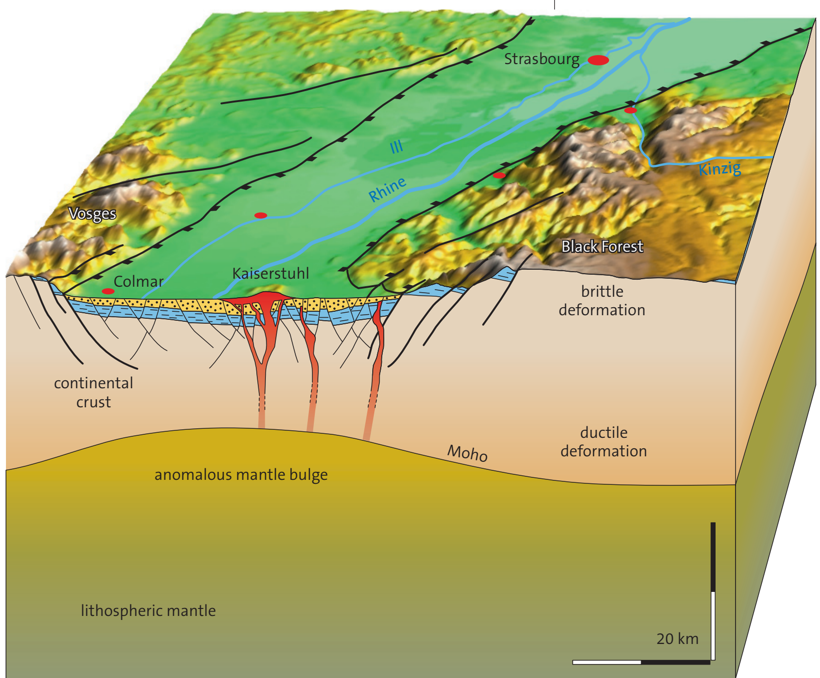
The Upper Rhine Graben in Germany

Although the Upper Rhine Graben in Germany is not one of the largest nor most active graben systems, it is along with the East African graben system a type locality for the study of graben systems. The term "graben" (*German* ditch) was used by miners for blocks that were dropped down at faults (Pfannenstiel, 1969) and was introduced into the geologic literature by Jordan (1803). Élie de Beaumont (1841) was the first geologist to describe the Rhine Graben. He understood that the facing Vosges and the Black Forest regions were broad, plateau-like uplifts separated from each other by the Upper Rhine River plain (**Fig. 3.5**). He further noted that the Rhine plain was down-dropped and bounded by parallel faults that had dip directions towards each other. This is the classic geometry of a graben system. The acceptance of the term "graben" in scientific literature was solidified by the classic work "Das Antlitz der Erde" ("The Face of the Earth") by Eduard Suess (1885–1909).

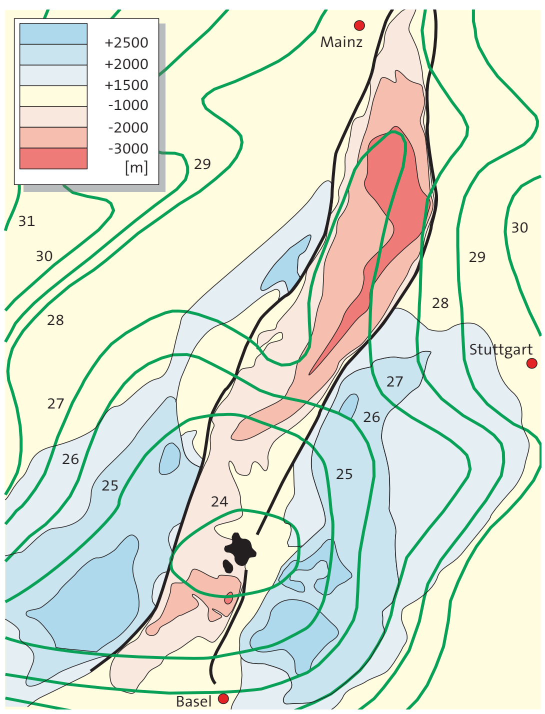
The Upper Rhine Graben extends more than 300 km from Basel (Switzerland) to Frankfurt (Germany) and forms a part of a larger fracture system that runs from the mouth of the Rhône River to the North Sea (**Fig. 3.5**). The bordering faults have dip angles that range from 55 and 85° near the surface; however, a majority of the faults dip between 60 and 65°. The faults at the flanks are parallel and all faults dip towards the center of the graben and decrease in inclination at depth. The graben has a fairly constant width of approximately 36 km with a crustal extension of approximately 5 km (Illies, 1974a). Crustal thinning is 6–7 km maximum and the continental crust in the southern part of the graben is thinned to 24 km (**Figs. 3.6, 3.7**). The graben parallels the axis of an elongated, stretched bulge that is mirrored in the graben shoulders on both sides, the Vosges to the west and the Black Forest to the east. Regionally, the graben shoulders are tilted 2–4° away from the graben.

The presently active earthquake foci occur mostly at a depth of less than 15 km. This indicates that brittle faults disappear at depth, and the crustal rocks below are deformed ductilely and not fractured. Ductile deformation of rocks rich in quartz (most of the rocks of the continental upper and middle crust are rich in quartz) initiates at temperatures of ca. 300 °C because quartz reacts from stress with plastic deformation at these temperatures. Seismic data indicate that the lower crust, dominated by rocks poor in quartz or without quartz, also reacts ductilely because of the higher temperatures; a pervasive horizontal lamination is interpreted to be the result of plastic flow (Illies, 1974a).

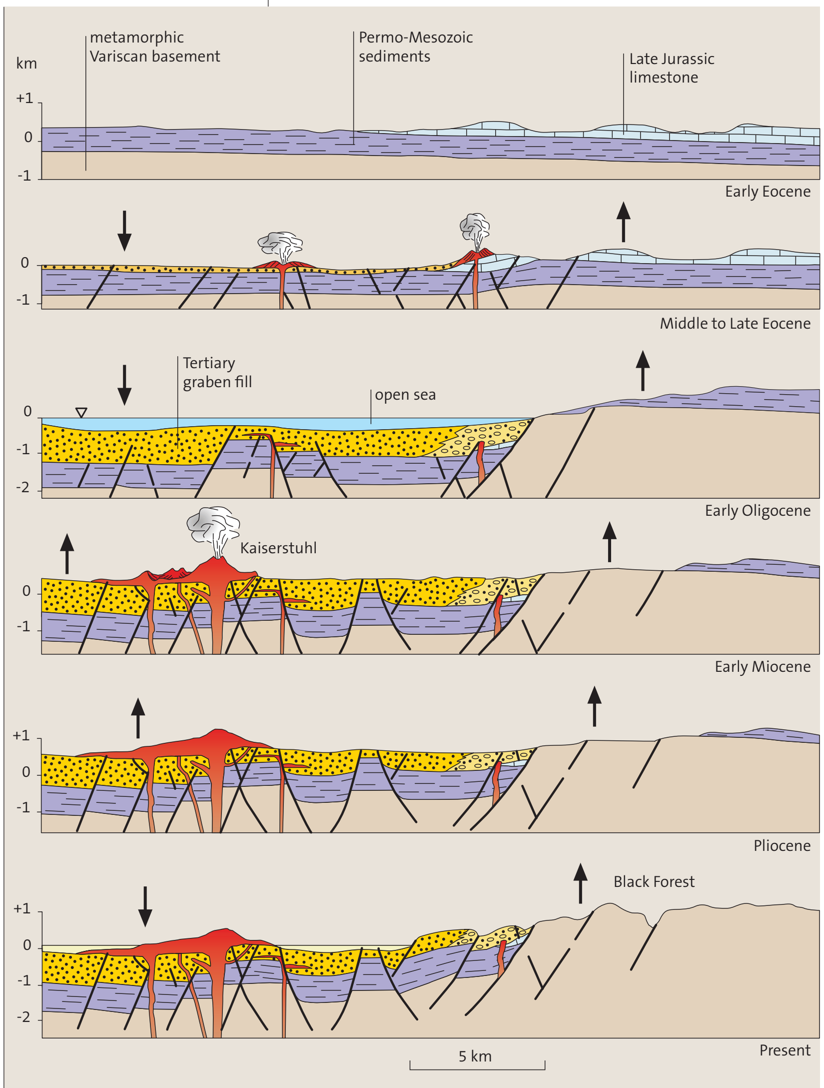
Seismic and gravity data indicate that in the Earth's mantle directly below the base of the crust, an anomaly of rocks with relatively low density



▲ **Fig. 3.6** Block diagram of the Upper Rhine Graben. Note that the upper crust is characterized by normal faults whereas the deeper crust is ductilely extended (by plastic, fractureless deformation). The Kaiserstuhl is a Miocene volcano.



◀ **Fig. 3.7** Map showing topographic and structural features of the Upper Rhine Graben as well as the amounts of uplift of the graben shoulder and subsidence in the inner part of the graben. Colored areas indicate the present level of the Early Tertiary erosional surface relative to sea level (extrapolated into the air in the blue areas). Green lines indicate the level of the crustal base (Moho) in kilometers below the sea level.



◀ Fig. 3.8 A series of cross sections showing the evolution of the southern Upper Rhine Graben in the area of the Kaiserstuhl volcano (Schreiner, 1977).

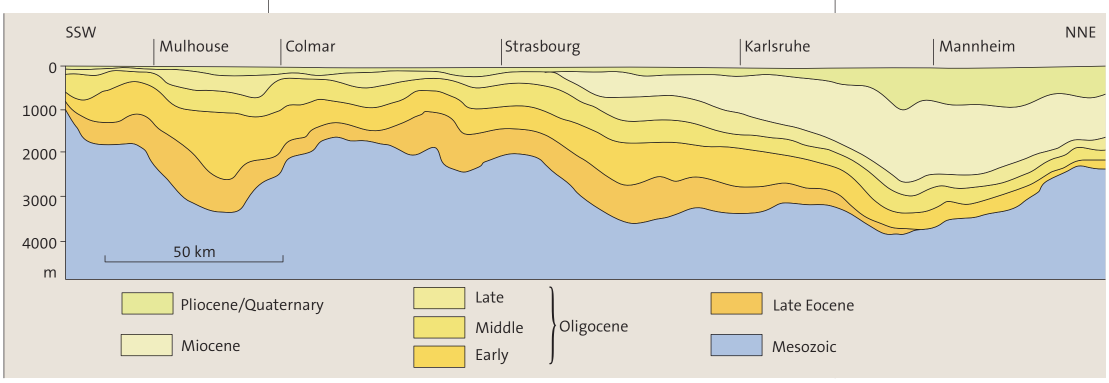
exists. Here, hot and probably partly molten mantle rocks that rise because of the lower density, feed volcanism that is related to the graben formation. This suggests that a mantle bulge is responsible for the bulge of the crust (**Fig. 3.6**).

Formation of the graben, as indicated by initial normal faults, started in the Eocene at ca. 45 Ma. The first sediments were deposited in the downthrown graben block. Extensional forces orthogonal to the graben axis enabled the opening of the graben.

Today, the surface bulge extends more than 200 km orthogonal to the graben axis. Uplift of the graben shoulders varies regionally. More than 2 km of uplift have been documented along the southern end. There, the pre-Tertiary peneplain erosional surface (eroded today) would be more than 2500 m above sea level (**Fig. 3.7**; Illies, 1974b). Total structural displacement across the graben varies from more than 5 km in the south to 4 km in the north. The graben shoulders are not significantly developed in the northern part, although the subsidence of the graben is generally greater and the Tertiary sedimentary fill has a thickness of more than 3 km. The result is a significant topographic gradient parallel to the graben axis. As explained below, subsidence of the northern part occurred distinctly later than that of the southern part.

The history of the Upper Rhine Graben

The Upper Rhine Graben has been filled with nearly 20,000 km^3 of Tertiary sediments (Roll, 1979). Most sedimentary rocks, both pre- and syn-rift, are eroded from the area of the graben shoulders. Along the edges of the graben, coarse-grained clastic sedimentary rocks include conglomerate and immature sandstone. The graben center is dominated by finer-grained clastic sedimentary rocks including siltstone and mudstone; non-clastics include limestone, dolomite, marl, and evaporites (salt rocks). Marine incursions generated saline to brackish conditions. During arid periods, evaporite



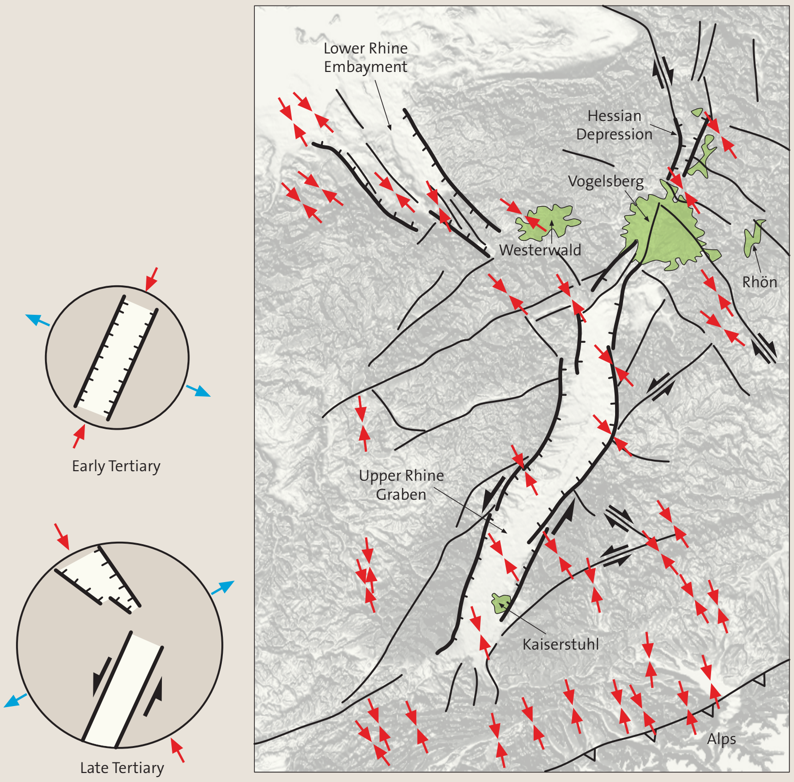
◀ Fig. 3.9 Thickness of sediments in the Upper Rhine Graben in a profile parallel to the graben axis (Pflug, 1982). Differences in the sedimentary thickness indicate that graben formation initiated in the southern area during the Early Tertiary, and that subsidence migrated to the north during the Late Tertiary.

The Upper Rhine Graben in the Middle European stress field

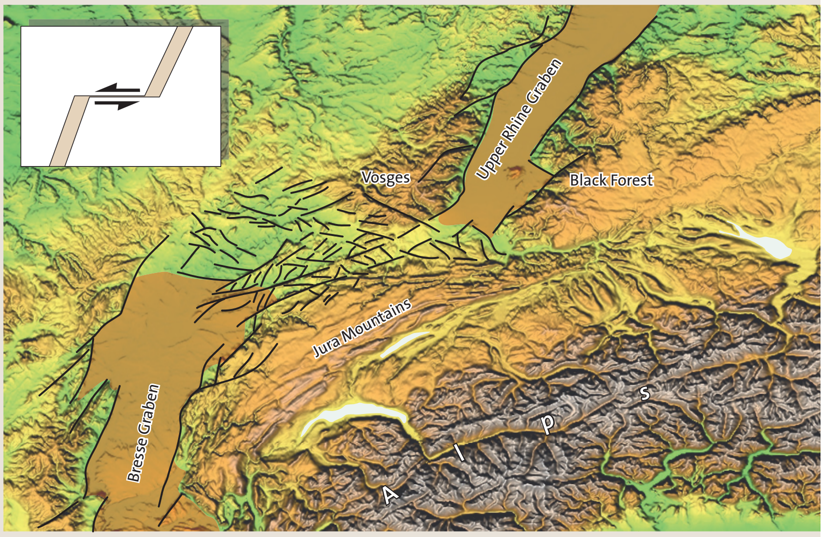
Presently the extensional forces in the Upper Rhine Graben are oblique (SW-NE, with an azimuth of 050–060°) and not orthogonal to the graben axis. Consequently, the principal axis of compression is not parallel but oblique to the graben axis (NW-SE or 140–150°; **Fig. 3.10**). However, when the graben formed, the extensional forces acted orthogonal to the graben. An anticlockwise rotation of the stress field occurred during the evolution of the graben system during the Late Tertiary and caused a left-lateral component to develop that overprinted the normal faults. Therefore, as the Vosges and Black Forest diverged, the Vosges also moved parallel to the graben, southward in relation to the Black Forest (**Fig. 3.10**).

The northern continuation of the Upper Rhine Graben trends along the Lower Rhine Embayment (**Fig. 3.10**). Neither a mantle bulge nor graben shoulders occur in this region. The graben system varies in width and opens towards the NW. Throughout its development, the extensional stress has been orthogonal to the Lower Rhine Embayment since the Early Miocene. The missing mantle bulge indicates that it is a passive rift structure and that extensional forces are the primary cause for the graben development.

The southern continuation of the Upper Rhine Graben, the Bresse Graben occurs approximately 120 km to the west (**Fig. 3.11**). As in the Upper Rhine Graben, subsidence in the Bresse Graben ceased during the Early Miocene. Offset between the Upper Rhine Graben and Bresse Graben is only apparent; in fact, the distance between the two graben segments remained unchanged since their formation. The situation is that of a transform fault, although such a fault is not developed as one distinct fracture. Rather, the rift structure is transformed by a complex system of mostly W-E trending minor faults linking the Upper Rhine Graben and the Bresse Graben (**Fig. 3.11**). The individual faults in the transformation zone carry out left-lateral (sinistral) offset (see box in **Fig. 3.11**).



▲ **Fig. 3.10** Directions of the present maximum horizontal stress (red arrows) in the Upper Rhine Graben and Lower Rhine Embayment (Blundell et al., 1992). The change in the orientation of the stress field is shown in the two schematic diagrams. The older stress field caused the formation of the Upper Rhine Graben, the younger one led to the formation of the Lower Rhine Embayment and to left-lateral movements in the Upper Rhine Graben. Volcanoes related to the graben formation are shown in green.



▲ **Fig.3.11** Map showing the transition zone from the Upper Rhine Graben into the Bresse Graben. The transition is accomplished by a bundle of faults that in total mark the locus of a transform fault. A simplistic representation of the connection between the two graben structures is shown in the insert. The distance between the two graben axes remained unchanged through the course of time.

deposits including halite and potash salts formed in the narrow, restricted seaway. Potash salt is particularly soluble in water and thus it is precipitated only at very high concentrations of evaporate minerals. Because of the low density and high mobility of evaporitic sediments, they rise as diapirs after being overlain by denser rocks. Some salt diapirs nearly reach the surface and are in contact with Quaternary sediments (Pflug, 1982).

The uplift of the shoulders initiated in the southern part of the graben during the Eocene (**Fig. 3.8**). Vertical displacement of more than 1000 m between the graben basin and its shoulders existed at the end of the Eocene as major stream systems were carved deeply into the rift shoulders and spewed coarse conglomerate onto the rift plain. Subsidence of the graben allowed marine ingressions from both the south and north. During parts of the Oligocene, a marine passage existed between the Alpine Molasse zone in the south through the Upper Rhine Graben into the enlarged North Sea Basin. Conglomerate input from graben shoulders decreased, indicating reduced tectonic activity and a lowering of relief. In the Late Oligocene, subsidence concluded in the southern part of the graben while further to the north fresh water lakes expanded as subsidence continued.

The last marine tansgression occurred in the Early Miocene, this time from the Lower Rhine Embayment, which was now established as a new branch of the rift system (**Fig. 3.5**). The southern half of the graben was filled with sediment by this time and completely lacks deposits of younger Tertiary age. Meanwhile, the Miocene Kaiserstuhl volcano was formed (**Figs. 3.6, 3.8**). As the southern rift waned, subsidence continued to the north, a trend clearly documented by a longitudinal section through the graben (**Fig. 3.9**). In the Late Miocene and Early Pliocene, sedimentation was interrupted across the entire graben system signifying general regional uplift. Sediments of Late Pliocene age are dominated by fluviatile deposits, present only in the northern half of the graben. Quaternary sediments locally attain thicknesses of more than 200 m indicating continued tectonic activity coupled with present earthquake activity.

Magmatism and heat flow in the Upper Rhine Graben

Magmas feeding the volcanoes associated with the rift formation of the Upper Rhine Graben are strongly undersaturated in silica and originate from a depth of about 80–100 km, the base of the lithosphere. The magma has been modified at shallower depth by differentiation. Therefore, a broad variety of volcanic rocks evolved within the graben system. Volcanic rocks of the Upper Rhine Graben are of minor volume when compared to the voluminous volcanic rocks of the East African graben system (see below).

The most famous volcanoes of the graben system are the Kaiserstuhl and the Vogelsberg (**Fig. 3.10**). The Kaiserstuhl is constructed of lavas and tuffs that erupted during the Early and Middle Miocene. Its location corresponds to where the crust is thinnest and large faults exist (**Fig. 3.7**). The volcanic rocks are strongly alkaline and undersaturated in silica. This is reinforced by the presence of carbonatites which contain mostly silica-free carbonate minerals. Globally, carbonatites are very rare rocks and are almost exclusively related to graben structures. They are derived from mantle-sourced magmas that contain carbonic acid. The occurrence of carbonatites in the Kaiserstuhl is one of only a very few in Europe.

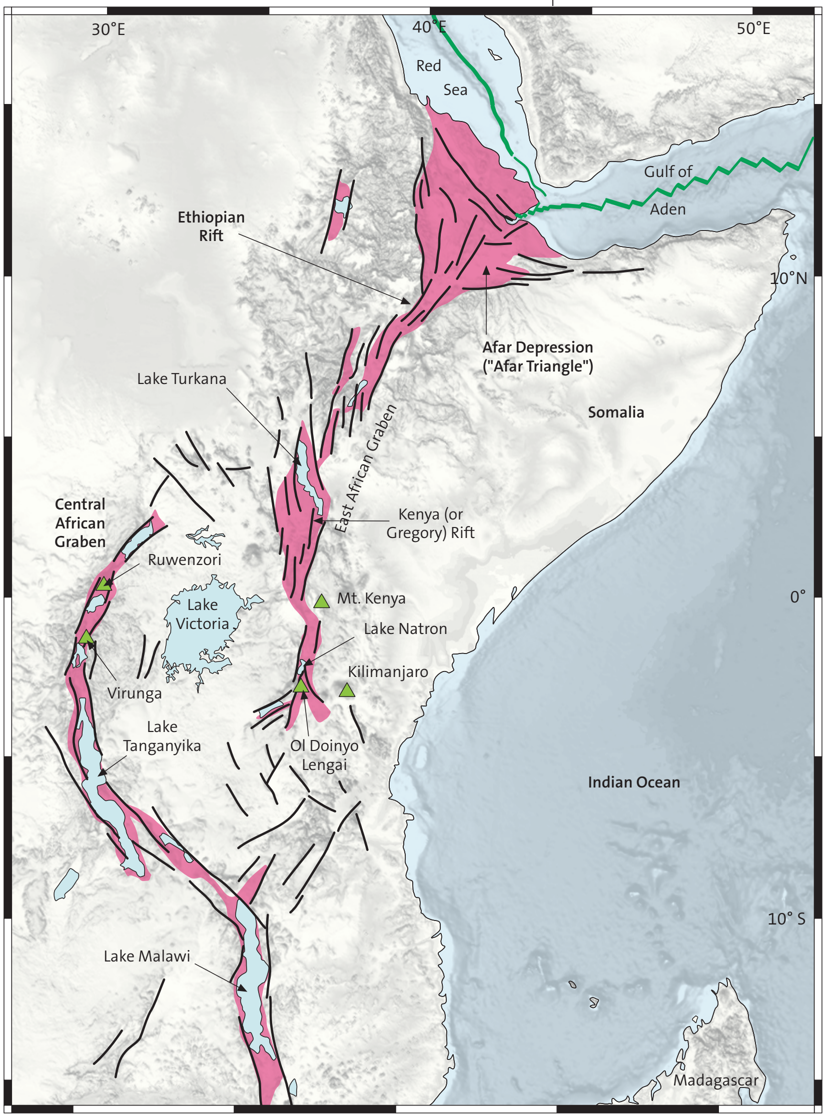
The largest volcano associated with the evolution of the Upper Rhine Graben system is the Vogelsberg. It is located in the northern continuation of the Upper Rhine Graben (**Fig. 3.11**). The broad shield volcano formed in a similar manner to the Kaiserstuhl. It was constructed during the Early and Middle Miocene, 19–10 Ma, and covers an area of approximately 2500 km². Emitted rocks are mostly alkaline basalts.

Because of the relatively low depth to the mantle and associated magmatic acitivity, graben systems are also zones of high heat flow. Below the Upper Rhine Graben, the temperature at the upper boundary of the mantle is at least 200 °C higher than beneath the graben shoulders. Surficial heat flow is about 50–80 milliwatt per square meter (mW/m²) outside the Upper Rhine Graben but increases to values between 80 and 120 mW/m² in the Upper Rhine Graben (Blundell et al., 1992). Locally values greater than 150 mW/m² are possible. Temperature at 1 km depth at these localities is 80 °C (Werner and Doebl, 1974), whereas in areas with a normal geothermal gradient it would be 30 °C.

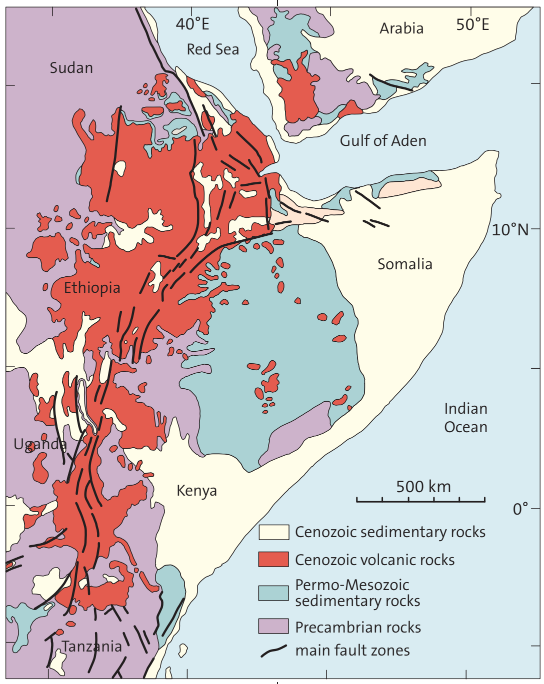
Such high thermal anomalies cannot be explained by heat conduction through the rocks alone because rocks are poor heat conductors. Rather, hot circulating water that migrates rapidly along fault zones is responsible. Cold surface water percolates to the depth, becomes heated, and ascends by convection. Such water cycles are expressed at thermal springs, common along the master faults of the Upper Rhine Graben, such as the health resort of Baden-Baden.

The large East African rift system

Along the greater East African rift system and the three-pointed graben star of the Afar Depression,



◀ **Fig. 3.12** Map of the principle elements of the East African graben system.

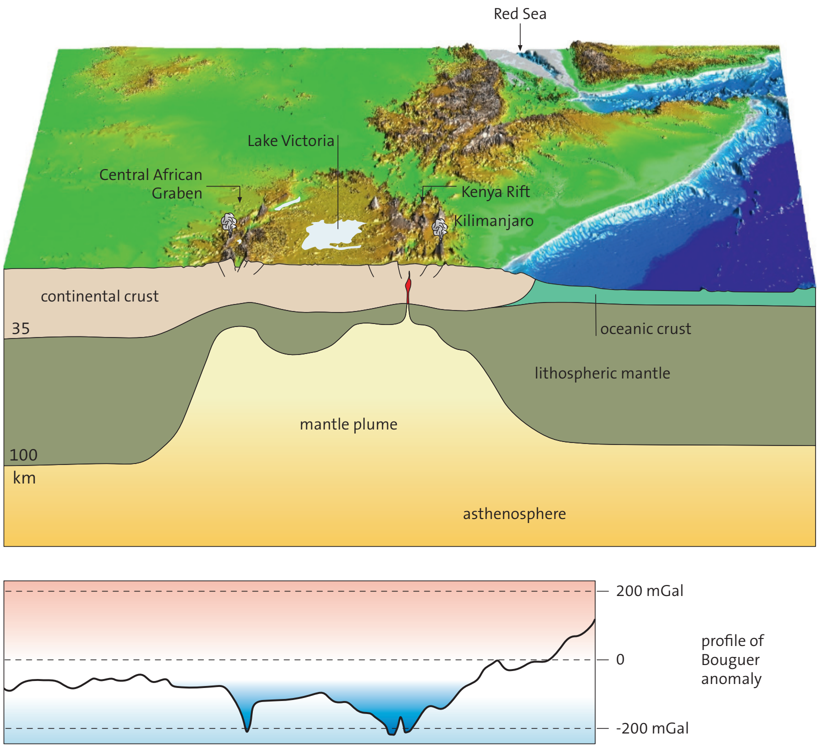


▲ **Fig. 3.13** Generalized geologic map of East Africa.

where the East African rift system, the Red Sea, and the Gulf of Aden meet, different stages of continent break-up are represented (**Fig. 3.12**). The East African Rift, from which the Central African Rift between Lake Malawi and Lake Turkana branches off, is a presently active system with abundant volcanism (**Fig. 3.13**). The East African rift system has not matured enough to have formed a new plate to the east, although the crust is nearly severed at some places and the term "Somalian Plate" is used by some geologists (**Fig. 3.14**). At its northern end, the Afar Depression, which because of its triangular shape is also called the Afar Triangle, has partly generated new oceanic crust.

The region is characterized by two broad topographic uplifts, each more than 1000 km in diameter (**Fig. 3.12**), and each underlain by mantle diapirs with broad mushroom-shaped heads. The northern uplift includes Ethiopia and Yemen and has the three-pointed graben star at its center. The southern uplift area is in Uganda, Kenya and Tanzania and is marked by the intersection of the Kenya and Central African rifts. The faults of the rift systems are generally parallel to structures of the Precambrian basement, an observation that suggests that the graben structures follow old zones of weakness in the crust.

The East African Rift has evolved since the Late Oligocene or Early Miocene. It comprises the Ethiopian Graben and the Gregory or Kenya Rift (**Fig. 3.12**). The Gregory Rift Valley, eponymous for the geological term "rift", has shoulders that rise more than 3000 m above sea level and 1000 m above the inner part of the graben. Collective vertical displacement along the main graben faults much as 4 km. The graben area is cut by densely



▲ **Fig. 3.14** Block diagram of the East African graben system. The lower cross-section through the Central African Rift and the Kenya Rift demonstrates the strong thinning of the lithospheric mantle that causes a negative gravity anomaly (Baker and Wohlenberg, 1971). 1 Gal (galilei) = 1 cm/s^2 (unity of acceleration). 1 mGal = 10^{-3} Gal.

clustered faults that parallel the edges of the graben and define a horst-and-graben structure. A horst is a higher block between two down-dropped graben structures. The total graben has a width of 60 to 70 km but the width of the inner graben is 17 to 35 km. The base of the inner graben is covered by Pliocene and Quaternary volcanic rocks and sediments and the clear dominance of volcanic rocks contrasts sharply with the rarity of volcanics in the Upper Rhine Graben. Mt. Kilimanjaro, an active volcano, towers nearly 6000 m high and along with the older and partly eroded Mount Kenya, is located on the eastern graben shoulder.

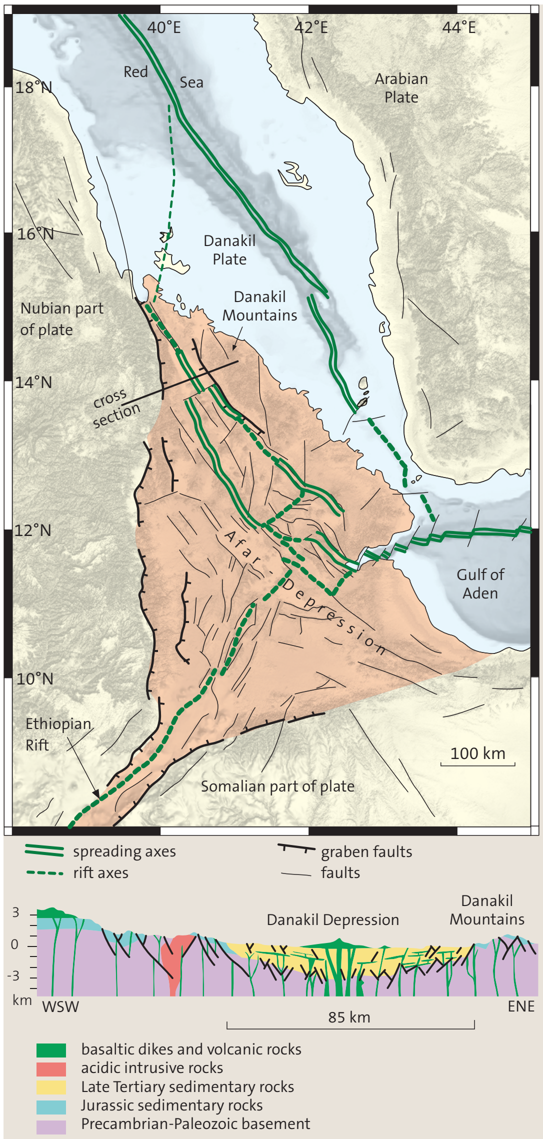
Towards the inner part of the graben, small but steep erosional surfaces limit the transport of sediments, and consequently, deep depressions evolve that are commonly the sites of large lakes. The largest is the 650 km long Lake Tanganyika in the Central African graben system (**Fig. 3.12**); its depth is nearly 1500 m and its bottom lies 700 m below sea level. The lake is situated behind a 2000 m-high mountain range that forms a rain shadow from the trade winds. Therefore, the region is relatively dry with a total annual precipitation less than 1000 mm, and the evaporation rate is high because of its position 3–9° south of the equator. The sediments of the lake are mostly biogenic and chemical in origin with only minor input by rivers. The depth of the water body coupled with the low production of sediment causes the longevity of such lakes. Some of the smaller lakes in the East African graben system, such as Lake Natron in Tanzania, contain high concentrations of sodium hydroxide and with associated high evaporation rates are dominated by evaporite deposits such as sodium carbonate, a by-product of volcanism. Chemical processes are strongly enhanced by bacterial activities (Kraml and Bull, 2001).

Since the Miocene, the Gregory Rift has produced alkaline volcanic rocks, especially strongly alkaline basalts and phonolites. The volcanic rocks have a total volume of about 100,000 km^3, half of which is basaltic. Interestingly the Gregory Rift is the site of the only active carbonatite volcano on Earth, the Ol Doinyo Lengai in northern Tanzania (**Fig. 3.12**). The Central African Rift has produced much less volcanic rock material, although the 4500 m high volcanic chain of the Virunga Mountains in the border region of Congo, Rwanda and Uganda belong to this graben system.

In the Ethiopian Rift, basalts have clearly dominated (ca. 300,000 km^3) since the Eocene; these are dominantly slightly alkaline and range to tholeiitic in composition. The huge volumes of low-viscosity flood basalts have formed 2000 m-thick basaltic plateaus (see Ch. 6). Acidic differentiates in Ethiopia are mostly alkaline rhyolites to trachyandesites. Non-basaltic rocks have a volume of about 50,000 km^3.

The larger volumes of basalts in Ethiopia and their slightly alkaline to tholeiitic chemistry signify a higher percentage of partial melting in the mantle source compared to those in the Gregory Rift. The alkaline magmas of the Gregory Rift are assumed to originate from the lithospheric mantle below the continental crust in a location that is enriched in elements such as alkalis that are incompatible with the mantle rock. In contrast, the tholeiitic magmas of the Ethiopian Rift are mainly derived from rising asthenospheric material that is depleted in these incompatible elements (Wilson, 1989). The differences between the Gregory and Ethiopian rifts are also explained by the different extension rates: total crustal extension in Ethiopia approaches 60 km, but in contrast, much lower rates, 35–40 km in northern Kenya, 5–10 km in southern Kenya, and less than 5 km in northern Tanzania are present in the Gregory Rift. The graben system opened like a pair of scissors, which explains the increasing magma production and decreasing alkalinity (tholeiites originate from higher partial melting in the mantle rocks) from south to north.

The average extension rates in the East African graben system, 0.4–1 mm/yr, are one order of magnitude less than those at a slow-moving constructive plate boundaries such as the nearby Red Sea or Gulf of Aden. Such a graben system is generally



The Afar Depression

The Afar Depression is a lowlying triangular area (**Fig. 3.15**), at the center of a three-point graben star, where the East African graben system (Ethiopian Rift), the Red Sea, and the Gulf of Aden meet. Here, the transition from a continental graben to an initiating ocean basin can be observed. Underlying the depression, a mantle diapir rises and overlying continental crust is strongly thinned and fragmented. In fact, between the separated continental fragments, new oceanic crust has been generated, although it is uncertain whether continuous bands of ocean floor already exist or not. Because the region is above the sea level and thus accessible for direct observation, in the 1970's many aspects concerning the theory of plate tectonics were established here.

Numerous bundles of faults, visible on satellite images of this arid region, pervade the depression. Along these faults, some partly expressed as open-spaced cracks, basaltic lavas with a tholeiitic composition similar that of mid-ocean ridge basalts, periodically discharge. Narrow stripes of quasi-oceanic crust are produced at several spreading axes and have produced a complicated pattern of microplates (**Fig. 3.15**). The tectonic and volcanic activity is concentrated along the inner part of the graben system, an area characterized by both horizontal and vertical movements of blocks. Regional crustal thickness of 30 to 40 km in the upland of Ethiopia is thinned to less than 16 km in the northern Afar Depression (Makris et al., 1975). The principle of isostasy is illustrated in the Danakil Mountains where a block (microplate) greater than 20 km in thickness rises as a highland above the surrounding plain. The adjacent Danakil Depression is an area below sea level and marks a portion of the plate boundary (**Fig. 3.15**). This trend continues northward towards the central part of the southern Red Sea depression where crustal thickness is ca. 6 km, typical of normal oceanic crust.

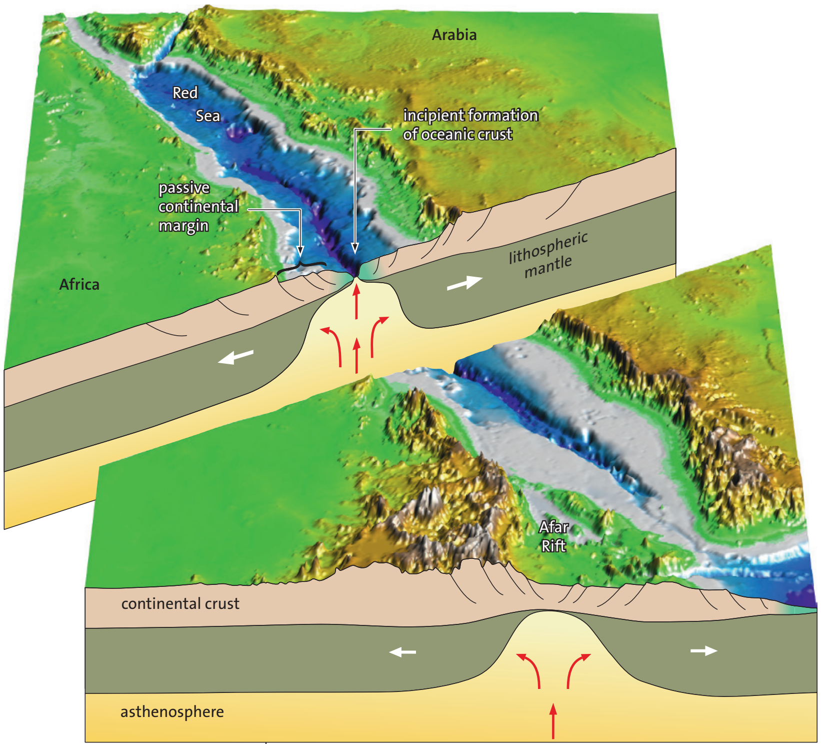
◄ **Fig. 3.15** Map showing structural elements of the Afar Depression or Afar Triangle. The region consists of a mosaic of blocks with thinned continental crust including the small "Danakil Plate". Basaltic rocks lie in the separation of the crust within the narrow spreading axes between the blocks. This relation is shown in the cross section through the Danakil Depression.

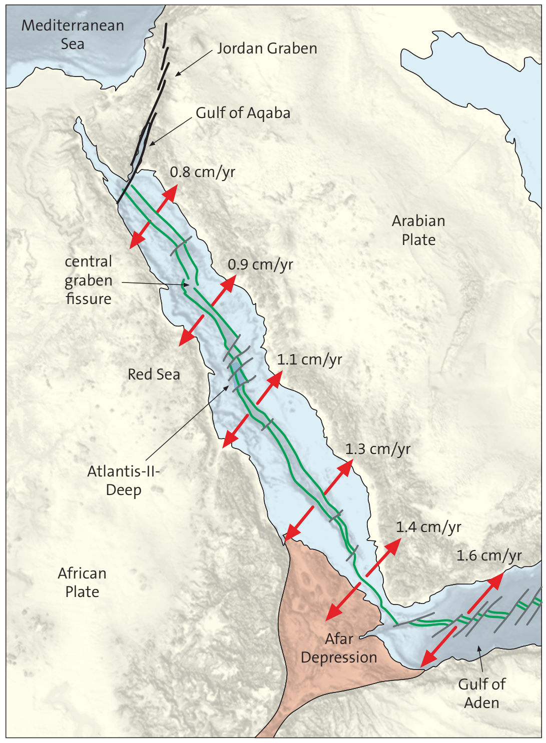
not considered to be a plate boundary but rather the result of intraplate tectonics. Activity in the East African graben system is waning, which is also expressed by its limited seismic activity. This is typical of extensional triple junctions, such as the Afar Triangle, where three arms meet – one arm is shut down after the other two have formed ocean crust between continental blocks.

The entire region is characterized by negative gravity values and local high heat flow. The graben system is underlain by a 1500 km wide bulge of the asthenosphere that nearly cuts through the lithosphere in the Kenya Rift (**Fig. 3.14**); a 20 km-wide intrusion has protruded to a depth of only 3 km below the sole of the graben. The intrusion is detectable in the gravitational profile by a slight positive anomaly within the broad negative anomaly of the graben (**Fig. 3.14**; Baker and Wohlenberg, 1971). The negative gravity anomaly mirrors a widespread mass deficiency caused by the bulge of the asthenosphere, which is less dense than the replaced lithospheric mantle. The mantle lithosphere has been transformed into asthenosphere by the temperature increase.

The Red Sea – from rift to drift

The Red Sea is a relatively recent constructive plate boundary that consists of a band of oceanic crust up to 100 km in width that was formed during the Late Tertiary by the separation of Arabia from Africa (**Fig. 3.16**). Here, the rift stage transformed into the drift stage and the Red Sea forms a nascent





▲ Fig. 3.16 Block diagram of the Red Sea region. Note the graben-in-graben structure, the high elevation of the graben shoulders, and the central graben fissure on oceanic crust in the middle of the Red Sea. The foreground cross-section that passes through the southern Afar Depression indicates that the continental crust is not severed there.

► Fig. 3.17 Spreading rates in the Red Sea and the Gulf of Aden (DeMets et al., 1990). The central graben fissure in the Red Sea is marked by a double green line. The Atlantis-II-Deep is one of several depressions that contain metal-rich oozes.

ocean. The oceanic band contains a central graben that marks the plate boundary and attains a depth of more than 2000 m. Because of the small amount of drift and the closeness of the continental margins, the central graben is not yet developed as a mid-ocean ridge in the true morphological sense.

Plate divergence in the southern Red Sea is 1.4 cm per year and decreases towards the NW (**Fig. 3.17**). In the Gulf of Aqaba it merges into the transform fault of the Jordan Graben (Ch. 8). The Gulf of Aden represents a more advanced stage of ocean-basin formation with a mid-ocean ridge, a feature that continues eastward into the Indian Ocean. Here the spreading rate along the ridge increases to 7 cm per year (**Fig. 1.2**). Spreading direction of ocean floor in both the Red Sea and Gulf of Aden is in the same, SW-NE, although the spreading axes of both oceans have different orientations.

The history of the Red Sea dates back to the middle Cenozoic. The Red Sea rift is cut through the Arabian-Nubian Shield, an area of Precambrian continental crust (Ch. 12) that is mantled by Upper Cretaceous and Lower Tertiary sedimentary rocks. The formation of the graben initiated during the Oligocene and by Late Oligocene; 25 million years ago, violent volcanic activity with basaltic eruptions commenced in the Afar Triangle, a southern extension of the Red Sea. During the Middle to Late Miocene, sea water intruded the graben system from the Mediterranean Sea; the graben-restricted sea had a blind end to the south. Restricted water exchange, high rates of evaporation in the arid region, and low inflow of freshwater generated more than 3 km of salt deposits.

Sea floor spreading with a rate of 1–2 cm/yr started in the southern Red Sea in the Pliocene at approximately 5 Ma. At the same time the Red Sea opened to the Gulf of Aden and the Indian Ocean. Due to the total separation between the continental blocks of Africa and Arabia, the tectonic activity shifted from the edges of the graben to the new spreading zone in its center, the zone where new oceanic crust was generated along a narrow, newly formed central graben; this formed the present graben-in-graben structure (**Fig. 3.16**). The outer graben shoulders are presently tectonically inactive.

The central Red Sea graben contains depressions with water depths greater than 2000 m. The water temperatures at depth increase to more than 60 °C and the salinity to more than 30% (Brewer et al., 1969). Hot brines are trapped in the depressions because of their high density: mud ooze formed on the seafloor is rich in iron, copper, lead, and zinc sulfides, ferric and manganese oxides, and calcium sulfates (anhydrite), gold, and silver. Total concentrations of non-ferrous metal in the oozes amounts to more than 10 % by weight and are potentially of economic interest. The economic potential is presently offset by the very difficult and expensive mining procedures that would be required to extract the ore. However, if the ocean floor became obducted onto the continental margin in the geologic future, like the Oman ophiolite (Ch. 5), mining might be economical.

The formation of these ore deposits is directly related to the opening of the Red Sea. The brines form an ideal environment in which to precipitate the ore content of hot water solutions that ascend from the depth. Salt solutions on the thinned continental crust along the sea floor next to the central graben were trapped in the depressions due to their high density and highly saline brines were formed. The basaltic lavas that extruded along the axis of the Red Sea heated the deep seawater and increased the solubility of the metals. Isotopic composition of the sulfur in the sulfides indicates that the total metal content is only partly derived from the basalts: the ratio of the isotopes sulfur-34 to sulfur-32 suggests that the ore metals are at least partly derived from the Precambrian crust of the adjacent continental blocks. Kuroko-type deposits that are present in local Precambrian rocks are present in the region and are rich in non-ferrous metal sulfides and gold (Ch. 7).

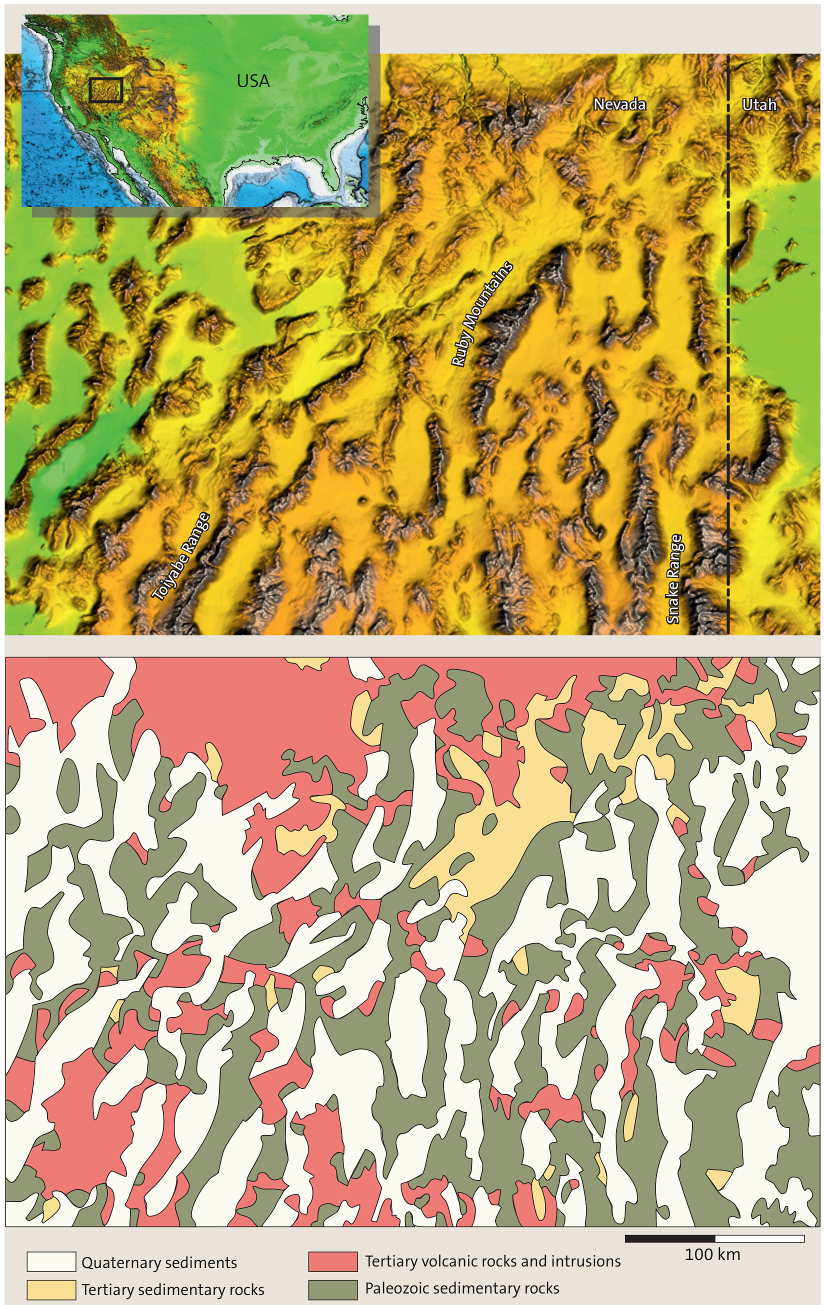
The geologic setting and range of processes that formed the Red Sea were ideal for the generation of metaliferous ore deposits. Submarine basaltic volcanism, faulting, water temperatures, and presence of metal ions in crustal rocks all combined to produce the ores. The ore stock of one of these depressions, the "Atlantis-II-Deep" (**Fig. 3.17**) is estimated to contain 3,200,000 t zinc, 800,000 t copper, 80,000 t lead, 4500 t silver and 45 t gold (Seibold and Berger, 1982).

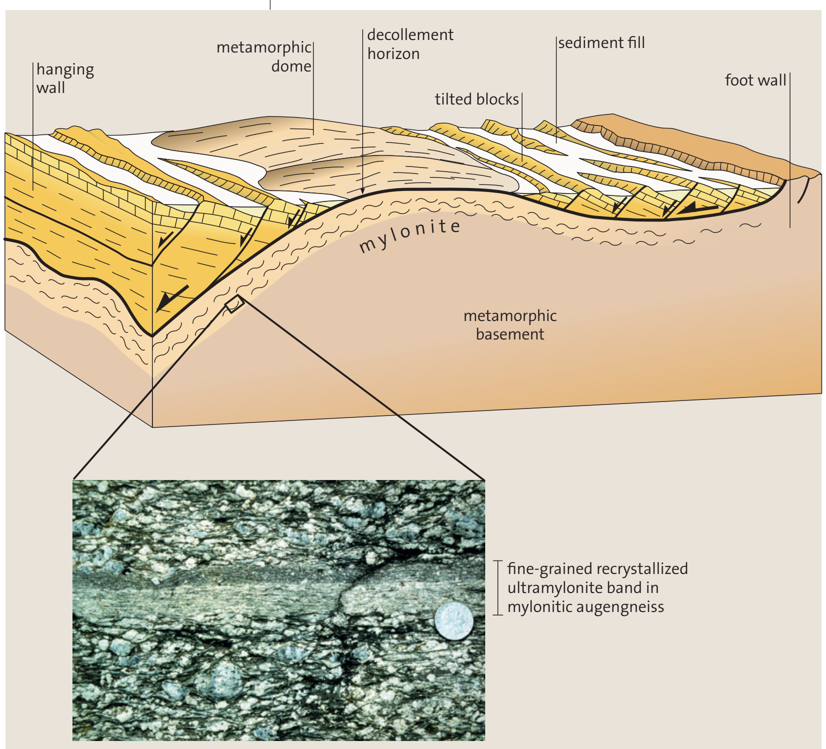
The extensional area of the Basin and Range Province

Continental crustal extension ranges between two styles, (1) areas with narrow zones of tectonic deformation characterized by a single, central graben and adjacent, uplifted graben shoulders, and (2) areas with broad zones of tectonic deformation characterized by numerous, parallel graben and horst blocks or metamorphic domes (see below). Examples of the first style have been described above and include the Upper Rhine Graben system and the greater East African rift system. When rifting continues over long periods of geologic time, such systems can evolve into constructive plate margins and young ocean basins – two plates are generated from a single precursor. Examples of the second style include the Basin and Range Province of western USA and the Pannonian Basin in Central Europe. These graben-horst systems attain great widths and have operated over long periods of geologic time without generating new oceanic crust or new plate boundaries.

The Basin and Range Province comprises two kinds of extensional faults, low-angle normal detachment faults with associated metamorphic core

▼ **Fig. 3.18** Topographic relief map and simplified geologic map of a portion of the Basin and Range Province in North America. The southern Snake Range along the Nevada-Utah border rises to 3982 m at Wheeler Peak, Nevada. Most basins are 1700 m above sea level across the central portion of the relief map. Ranges consist mostly of wide-ranging Paleozoic sedimentary rocks, some of which are metamorphosed across metamorphic domes. Faults, not shown for simplicity, bound the ranges.





▼ **Fig. 3.19** Block diagram of a typical metamorphic dome or metamorphic core complex. The dome is caused by thinning and brittle faulting of the hanging wall (upper plate), under which the foot wall (lower plate) rises in a dome-like fashion. The rocks in the foot wall have risen from the middle crust to the surface. Mylonites are ductilely deformed rocks that formed along the shear zone but are now exposed at the surface; they originally formed at middle-crustal levels. The mylonites (see photo, with band of very fine-grained ultramylonites) are overprinted by brittle deformation during the uplift and subsequent cooling. By comparing this figure with Figure 3.4b, the stages of evolution of a metamorphic dome can be visualized.

complexes (described below) that are responsible for most of the horizontal extension (over 200 % in some places), and high-angle normal faults that create the numerous parallel basins (grabens) and ranges (horsts). The Basin and Range Province is extended in an east-west direction across 600 km and encompasses an area of 550,000 km^3 (similar to that of France). It is oldest to the south in southern Arizona (ca. 30 Ma) and youngest to the NW in Nevada and Oregon where it is still active today. As both a relief map and a geological map clearly show, the characteristics of the ranges contrast sharply with those of the basins (**Fig. 3.18**). Ranges consist of elevated blocks that expose multiple rock types ranging in age from Precambrian to Cenozoic and display wide-ranging internal styles of geologic structure. Basins contain unconsolidated or loosely consolidated erosional debris derived from highlands and deposited in the depressions. Voluminous amounts of Late Cenozoic basalts were deposited within some basins. Elevation of the ranges and basins as well as local relief varies greatly across the province. For example, along the Utah-Nevada border in the center of the province peaks are almost 4000 m high and basin floors fall to 1700 m (relief of 2300 m), in south-central Arizona peaks are 3200 m and basin floors are 850 m (relief of 2350 m), in SW Arizona, the oldest portion of the province, ranges are 700 m and basins are 100 m (600 m relief), and in NW Nevada, the youngest portion of the province, ranges are at 2500 m and basins are at 1200 m (1300 m relief). Thickness of Cenozoic valley fill ranges from a few hundred meters, especially in younger basins, to several thousand meters.

Crustal thickness in the Basin and Range Province averages only 30 km, a low value when compared to the crustal thickness of the adjacent Colorado Plateau, nearly 50 km. Before the Cenozoic collapse of the Basin and Range Province, its crust was even thicker than that of the Colorado Plateau, as during the Mesozoic the Basin and Range Province was the site of the Sevier Mountains, a major thrust belt. The present Basin and Range Province exhibits a high geothermal gradient that is caused by anomalously light, hot, and partly molten asthenosphere.

The development of metamorphic domes

Metamorphic domes, commonly referred to as metamorphic core complexes, develop in regions of asymmetric rifting along a sub-horizontal, large normal fault zone called a detachment fault. The geometry of the overall system is extremely complex and is shown in simplified form in **Figure 3.4b**. Note that the detachment fault buckles in the middle continental crust where it merges into a ductile shear zone (Lister and Davis, 1989). At depth the shear zone breaks through the lower crust and breaches the lithospheric mantle (**Fig. 3.4b**). In the hanging wall or upper plate above the shear zone, steeper inclined and curved normal faults attend the shear zone and separate the upper plate into domino-like gliding blocks (**Fig. 3.19**). Such curved faults and shear zones are called listric faults (*Greek* shovel-shaped). Because of the concave upwards curving, the blocks above the faults are tilted.

Figure 3.4b displays an early stage of extension and **Figure 3.19** represents late-stage development. Comparing the two figures helps explain the origin of the metamorphic core. As the extension shown in **Figure 3.4b** expands, the upper plate thins; at the same time the heat underlying the extension zone bows the lower plate upward. The movement along the decollement horizon causes rocks with considerably higher grade of metamorphism from the middle and lower crust to abut rocks in the upper plate that consist of non-metamorphic or lower-grade metamorphic rocks. The shear zone bulges upwards at the place where the hanging wall (upper plate) is thinned the most; eventually the shear zone tears through the upper plate and the foot wall is exposed at the surface (**Fig. 3.19**). This type of exposure is called tectonic exhumation (as opposed to erosional exhumation). The thermal rise of the footwall (lower plate) coupled with the thinning and eventual rupturing of the brittle upper plate exposes the dome-like high-grade metamorphic rocks – the metamorphic core or dome. The metamorphic core has extended by ductile thinning while the upper brittle plate fractured by faulting. No wonder such complicated and unexpected juxtaposition of contrasting rock types was not adequately explained until 30 years ago!

Good examples of metamorphic domes are found in the Basin and Range Province, e. g., the Ruby Mountains (**Fig. 3.18**). In map view, the metamorphic core or dome forms an irregular elongate- to kidney bean-shaped body; dimensions can range to tens of kilometers. Depending on geometry and topographic relief, the core zone may only expose the mylonite zone (explained below), or it may also expose the underlying non-mylonized metamorphic rocks of the middle and lower crust. Rocks that form the core can have diverse geologic histories. They may be Precambrian basement rocks or Phanerozoic rocks formed at or near the surface, buried during mountain building to lower-crustal levels, and exhumed by the extensional processes discussed above. The metamorphic core is overlain by brittle rocks of the upper plate. Upper plate rocks can consist of sedimentary rocks and volcanic rocks, the latter formed when material from the elevated asthenosphere melts and rises to the surface along fault conduits. Across much of west-central Arizona, these faulted and tilted dark volcanic rocks form a distinctive moon-like landscape.

The large shear zone between the hanging wall and foot wall is ductilely deformed below ca. 10 km depth causing the formation of mylonites. Mylonites are formed by intense ductile, fractureless deformation and consist of fine-grained, recrystallized metamorphic rocks. They overprint the middle and lower crustal rocks of the core complex. Mylonites thus represent the surficial expression of the decollement shear zone that originally formed 10–20 km below the surface. Rocks originally formed at dramatically different crustal levels adjoin along the shear zone. Another characteristic of metamorphic domes is the rapid uplift required for their formation, typically rates of several mm/yr or kilometers per million years. However, the incredible uplift rates are not necessarily the sites of towering mountains. This is because the rapid uplift is accompanied by rapid horizontal extension – as much material is transported horizontally as is pushed up vertically. Therefore, rapid rates of erosion (impossible at this magnitude) are not required to expose the deep crustal rocks of the metamorphic core.

The crust of the Basin and Range Province has been extended up to two times its original width by the mechanism described above. However, the topography of the region is controlled by the later brittle faulting, the high-angle normal faults that flank the numerous ranges, features responsible for a much smaller fraction of crustal extension than the metamorphic domes.

A brief history of the Basin and Range Province

The geologic history and geologic controls that acted upon the Basin and Range region are complex and beyond the scope of this book. However, a brief account is presented here. Much of the Basin and Range Province originally comprised the Early and Middle Paleozoic passive margin of western North America and was the site of unusually thick sedimentary rocks. In the Late Paleozoic and Mesozoic, the region was an active margin characterized by subduction, terrane accretion, and orogenesis. The culminating Cretaceous Sevier orogeny generated thick continental crust, a large thrust belt, and towering mountains. The hinterland bisected the present Basin and Range area, and the future Colorado Plateau and Rocky Mountains provinces lay near sea level to the east.

The initial tectonic collapse of the hinterland initiated in the Paleocene and Eocene as flat-slab subduction formed under the American Southwest. Subduction ended as the East Pacific Rise (the Pacific mid-ocean ridge) intercepted western North America in the Oligocene. As the eastward-plunging Farallon Plate was consumed along portions of the North American Plate, the westward moving Pacific Plate contacted the continent. Although North America has a westward component of motion, the Pacific Plate moved at a much greater rate to the NW. A "space problem" developed along the margin of North America as the Pacific Plate attempted to pull it westward. Consequently, the thick crust below the old Sevier hinterland was extended to fill the gap and the great crustal extension commenced.

The interception of the Pacific and North American plates can be tracked along the northward-moving Mendocino triple junction. In fact, the current location at Cape Mendocino, California, marks the approximate northward extension of the Basin and Range Province; its inception near the latitude of San Diego marks the inboard location of the oldest Basin and Range extension in southern Arizona. Since the Middle Miocene, ca. 15 Ma, much of the motion between the two plates has been taken up by transform faulting, especially along the famous San Andreas Fault, the current plate boundary. The San Andreas Fault and its precursors have continuously carved eastward into North America resulting in several hundred kilometers of North America being "captured" by the Pacific Plate.