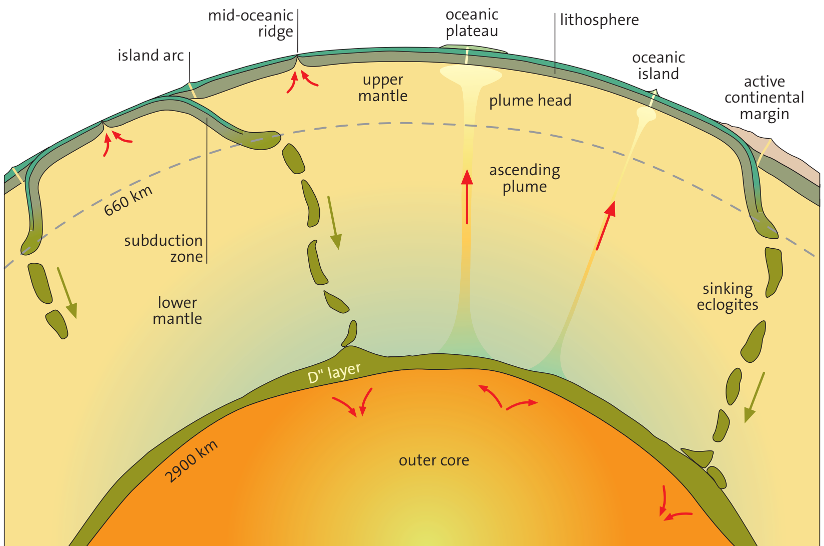
Hot spots

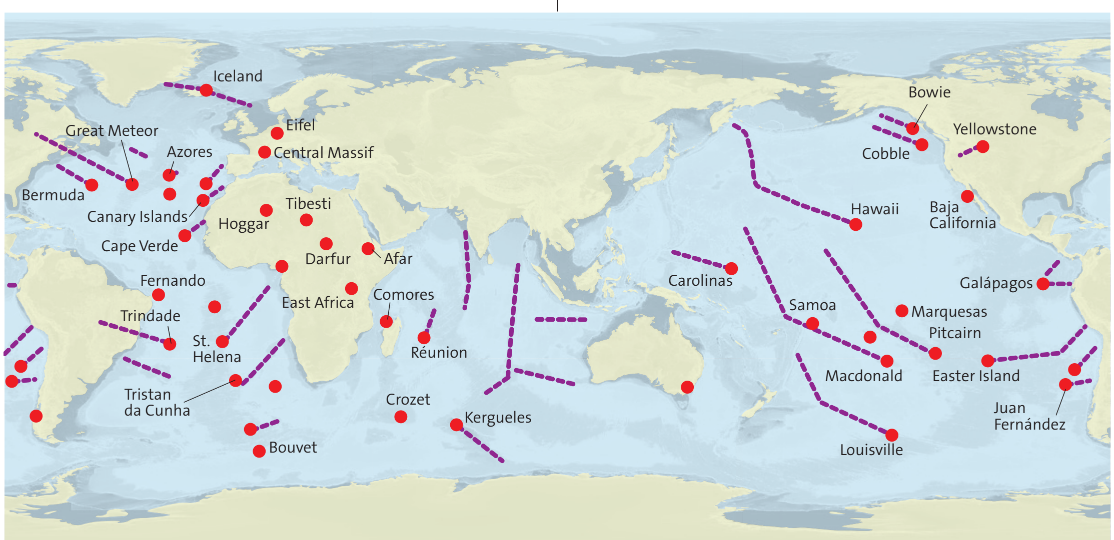
Although the vast majority of volcanoes on Earth are related to plate boundaries – mid-ocean ridges and subduction zones –, approximately 5 % are classified as “hot spots”. Hot spots are the product of mantle diapirs (*diapirein*, *Greek* to perforate) or plumes (because of their shape) that rise through the mantle as finger-shaped hot currents and penetrate the crust. In spite of their relatively low numbers, they play an important role in the convection system of the Earth’s mantle and are responsible for about 5–10 % of the melts and energy emitted by the Earth. At present, approximately fifty hot spots have been identified in both continents and oceans (**Fig. 6.1**). Although most are located in the interior of plates (“intraplate volcanism”, e. g., Hawaii), some hot spots are either coincident with mid-ocean ridges (e. g., Iceland) or in close proximity (e. g., Azores, Tristan da Cunha). Mantle plumes and the resulting hot spots are responsible for the formation of large volcanic complexes; these include volcanic chains up to several thousand kilometers long and huge flood extrusions, commonly called large igneous provinces or LIP, that consist of basaltic lavas. Plumes are produced in the lowermost mantle adjacent to the core within the so-called D'' layer (see below; **Fig. 6.2**), at a depth of about 2900 km. The hot source in the mantle is generally considered to be fixed in its position over long periods of time. The existence of hot spots and their significance as fixed points was first established by J. Tuzo Wilson, one of the doyens of plate tectonics (Wilson, 1963).

The boundary between the metallic core and the siliceous mantle is not only a material boundary



▲ **Fig. 6.2** Schematic cross section through the mantle. Mantle plumes generally initiate at the base of the mantle within the D” layer. Heavy parts of subducted plates sink down to the base of the mantle to feed the D” layer. The liquid outer core generates convection currents that are in concert with the mantle. Mid-ocean ridges have shallow mantle sources.

▼ **Fig. 6.1** Map showing global distribution of hot spots. Volcano chains formed by the hot spots (dashed lines) indicate the plate movement over the hot spots.



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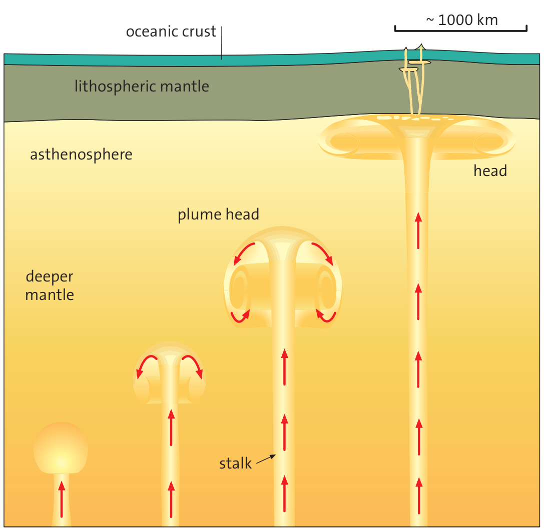
but also a thermal one where temperatures change drastically across a short distance. Temperature decreases from about 3000 °C by several hundred degrees within the lowermost 200 km of the mantle. Plumes are apparently triggered by hotter areas at the base of the mantle by heat transfer from hot spots in the liquid outer core. However, a plume is only generated if the temperature anomaly is large enough and exceeds a limiting value. In spite of its high temperature, the rising mantle rock is not molten because of the high pressure.

The plume ascends because its material is hotter than the surrounding mantle and thus has slightly lower density and viscosity; the mantle generally is able to flow ductilely in spite of its solid state. The plume has a cylindric conduit approximately 150 km diameter. Experiments indicate that it develops a bulbous head that moves upward through the mantle (**Fig. 6.3**). Processes associated with the rising head of the plume are complex; the head is slowed down by the viscosity of the surrounding mantle (in particular the lower mantle exerts a strong resistance); hot but less viscous material below the head causes expansion of the head so that its slower upward motion is compensated by the faster but smaller stalk below. As the head expands and rises, surrounding mantle material is also incorporated into it.

Melting and magma generation occur as the plume rises near the base of the lithosphere at a depth of 100–150 km. Here pressure is reduced and the temperature of the plume is 250 °C higher than that of the surrounding asthenosphere; this leads to the formation of significant amounts of magma by partial melting of the peridotite in the plume (White and McKenzie, 1995). The content of water in the mantle rock is also of importance as water reduces the melting point. Portions of the upper mantle are "wet" or "damp" and accordingly larger amounts of melt are produced. The wet mantle plays a role in Hawaii as well as in the Azores (see below). Water can be detected because it is bound as hydroxyl ions in basaltic minerals that evolve from the melts.



▶ **Fig. 6.3** Model for the evolution of a mantle plume based on experiments with glucose syrup (Griffiths and Campbell, 1990).

When the plume reaches the bottom of the rigid lithosphere, a broad mushroom-like head is formed that may achieve a diameter of 1000 km (**Fig. 6.3**). This head generates surface uplift because the push of the plume and the thermal expansion of the heated area causes bulging of the lithosphere; elevations greater than 1000 m can be attained. The magma generated in the head penetrates through the lithospheric mantle and crust. The final result is the extrusion of enormous volumes of basalt. Smaller plumes such as one under the French Central Massif do not have a distinctive head.

The consequences of hot spot magmatism can be extreme. The large volcanic edifices at Hawaii form the highest single mountain on Earth. They can also form huge flood basalt fields where magmatic extrusions occur within amazingly short periods of time, commonly within one million years. Such an extrusion occurred in the Deccan Trap basalt in India. Here the generation of huge volumes of basalt was related to the break-up of a portion of the large continent Gondwana. Huge, extensive intrusions and basalt flows can lead to thickening of ocean crust over wide areas such as the Ontong-Java Plateau in the western Pacific and other similar oceanic plateaus (**Fig. 6.14**).

Because of the enormous loss of magma material, more than 2 million cubic kilometers in some cases, the head of the plume cools rapidly and the magmatic activity subsides. However, the active life of some hot spots may persist more than 100 million years, commonly through pulsating events of volcanic activity.

Where hot spots underlie continents, the magma penetrates and is influenced and contaminated by continental crust. The basaltic melts differentiate and their heat may cause large volumes of crustal rock to melt. This process generates acidic magmatic rocks. Therefore, most continental hot spots coevally produce basaltic mantle melts and rhyolitic crustal melts. These two groups of rocks differ considerably in their composition. Basalts have SiO_2 contents of approximately 50 % and rhyolites approximately 70 %. Close associations of clearly distinctive magmas are called bimodal. Continental bimodal magmatic provinces are characteristic of rift and hot spot settings.

The fixed position of mantle plumes correlates with their origin near the core boundary but would not harmonize with an origin in the upper mantle that is dominated by strong convection. The "superplume" theory that explains the dominance of mantle plumes in the Cretaceous also supports the formation of plumes at the base of the mantle (see below). However, for some hot spots such as Iceland or Yellowstone, a shallow source in the upper mantle has been postulated.

Hot spots and mid-ocean ridges

Magmatic rocks associated with hot spots have a different mechanism of formation than those of the mid-ocean ridges even though both basaltic melts evolve in the uppermost mantle. The former, which originate within plates and are called intraplate magmatites, have characteristic chemical signatures that distinguish them from basalts of mid-ocean ridges (**Fig. 5.8**).

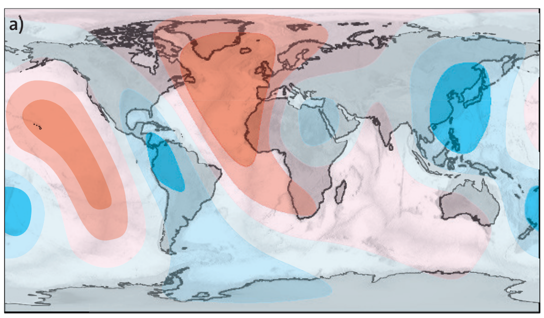
Mid-ocean ridge basalts (MORB) are tholeiites. Tholeiites evolve through relatively high percentages of partial melting of peridotite (mostly 15–25 %). Hot spots with high magma production generate flood basalts and island chains like Hawaii and also develop tholeiites, albeit of slightly different composition from that of MORB. Hot spots associated with low volumes of magmatic activity produce alkaline basalts. Compared to MORB, hot spot basalts contain distinctly higher contents of elements that are incompatible with mantle rocks including potassium, rubidium, phosphorus, titanium and the light rare earth elements. This mirrors the low degree of partial melting of alkaline basalts in the mantle (mostly below 10 %) as well as a mantle source that is different from the MORB source. The MORB source is solely the asthenosphere. In contrast, the intraplate basalts of the plumes contain components of the lower mantle, the asthenosphere, and the lithospheric mantle. In intraplate tholeiites, the asthenospheric portion dominates, in alkaline basalts the lithospheric mantle dominates (Wilson, 1989). If hot spots are located on a mid-ocean ridge, their chemical composition indicates a particularly high asthenospheric portion. Alkaline basalts generated from a fading hot spot reflect the waning melt formation and are rather coupled to a lithospheric source.

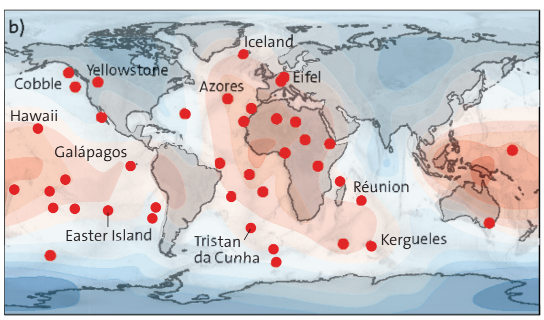
Isotope chemistry is a powerful tool for differentiating the source of magmas. All isotopes of a given element have the same number of protons but vary because of their different numbers of neutrons and therefore, have different atomic masses. The mantle reservoirs of hot spots are richer in incompatible elements. As confirmed by strontium isotopes, these different reservoirs have not been substantially mixed during the last 2 billion years with the exception of the mixing processes that occur in areas where hot spots are coincident with mid-ocean ridges. Presently forming MORB have $^{87}Sr/^{86}Sr$ isotope ratios of 0.7027 whereas basalts from hot spots have ratios of about 0.7040 (Nicolas, 1995). Because ^{87}Sr evolves from the decay of ^{87}Rb, there must be a deep mantle source for basalts derived from hot spots. The deep mantle has Rb/Sr ratios that are higher than those generated in the Rb-depleted asthenosphere, the source of MORB.

Helium isotopes also have a characteristic signature in basalts generated from hot spots; the high ratio of ^{3}He to ^{4}He is considered to be a signature of melts derived from the lower mantle. The primordial mantle contains ^{3}He whereas ^{4}He evolved during the Earth's history by the radioactive decay of uranium and thorium (alpha decay). Because the upper mantle degasses easier, it contains less ^{3}He and thus lower $^{3}He/^{4}He$ values than the lower mantle.

The mysterious D'' layer and the dented Earth

The D'' layer forms a zone at the base of the mantle. It is suggested that it is critical in the generation of mantle plumes. The D'' layer is generally 200 to 250 km thick but has extremes that range from 100 to 500 km. The origin of the term comes from an obsolete system of classifying the layers of the Earth devised by New Zealand geophysicist Keith Bullen (1942). The D shell in his classification stood for the lower mantle and was later subdivided in the main part of the lower mantle (D') and a seismically newly defined bottom layer (D''); this term





◄ **Fig. 6.4** Maps showing a) bulges on the core and b) geoidal bulges. Bulges (red) and indentions (blue) of the core have ca. 10 km of total relief (Morelli and Dziewonski, 1987; Vogel, 1994). Bulges and indentions of the geoid surface have ca. 200 m of total relief. Hot spots (red dots) are concentrated in the bulges (Chase, 1979).

alone has survived from the the Bullen system. An internet site by A. Alden sarcastically suggests we call it “hell”.

Current geophysical thought suggests that the D'' layer, because of its irregular and heterogeneous characteristics, represents the ultimate fate of subducted plates (Vogel, 1994). Below the long lasting, active subduction zones around the Pacific Ocean, seismic waves travel relatively rapidly in the D'' layer, indicating the presence of slightly cooler and denser material. This characteristic suggests to geophysicists that the D'' layer consists of remnants of oceanic crust that was subducted into the lower mantle where it sank to the bottom because of its high density (**Fig. 6.2**). It is assumed that these rocks contain stishovite, a high-pressure form of quartz that has a density of 4.34 g/cm^3, nearly twice that of normal quartz at 2.65 g/cm^3. The weight of this dense material from subduction zones causes the core to be indented several kilometers (**Fig. 6.4**). The presence of cooler subducted material at the core boundary likely induces downward-directed convection currents in the liquid outer core.

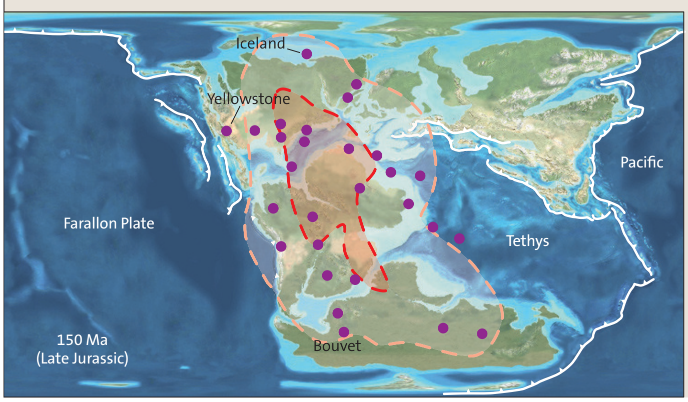
Beneath the Central Pacific, lower velocities of seismic waves indicate that the D'' layer is slightly hotter, less viscous, and therefore, easier to move. In this region, both the core and the surface bulge (**Fig. 6.4**) indicate that relatively hot and light material is ascending in the outer core and mantle. In the Pacific realm, the concentration of hot spots is especially high; their origin is assumed to be above the bulge of the core. The resulting bulge on the surface is represented by the geoid surface. After subtracting the influence of subduction zones, a broad equatorial bulge of the geoid surface 80 m high remains and is related to the concentration

Hot spots of Pangaea

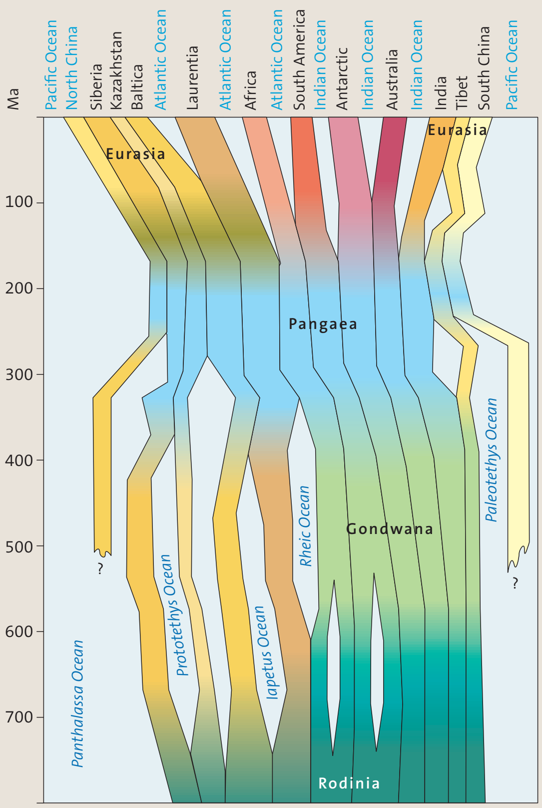
In theory, there should be a connection between the bulges of the geoid and the concentration of hot spots. Geoid bulges form because large areas of thick continental lithosphere have poor heat conductivity and have an insolating effect on rising heat from the sub-lithospheric mantle (Anderson, 1982). Mantle temperatures rise and the region forms a thermally uplifted bulge. Resulting bulges with 50 m of relief can form within 100 m.y. They tend to be centered near the equator where their configuration is most stable within the rotating Earth. Therefore, if large continental masses are concentrated near the equator, hot spots and resulting bulges will also concentrate near the equator. This causes the axis of Earth to adjust by true polar wandering.

An example of concentrated hot spots and large continental masses near the equator occurred during the break-up of Pangaea. Many hot spots are located within the African-Atlantic bulge at present (**Fig. 6.4**, lower) and as reconstructed for the continental masses of Pangaea during the Jurassic (**Fig. 6.5**). The concentration of so many hot spots in this configuration suggests that the present, currently active hot spots are old and initially formed during the time of Pangaea 300 to 175 Ma. Because mantle diapirs require approximately 100 m.y. to develop and ascend, most of the hot spots would have formed towards the end of Pangaea's existence. The hot spot bulges and the subduction zones at the margins of Pangaea (**Fig. 4.8**) produced extensional stress in the supercontinent. The plume locations formed corners for subsequent rifting – the sites of three-pointed graben stars (Ch. 3). Rift zones propagated by extensional forces, followed weak zones in the crust, and eventually connected the hot spot corners – the fate of Pangaea was sealed (**Fig. 4.7**). The crust of the supercontinent was fatally cracked and post-Pangaea plate patterns would rift it apart. The history of the Atlantic is closely connected to the activity of hot spots and corroborates the above picture. Individual parts of rifted continents drifted away from the zone of ascending mantle currents towards zones of descending mantle currents. New subduction zones formed at old passive margins and continents eventually rejoined, sometimes in approximately the same positions that they were torn apart previously. In the past 60 Ma, new collisions have occurred to form the Alpine-Mediterranean mountain range at the southern margin of Europe and Asia – a starting signal for the formation of a new Pangaea?

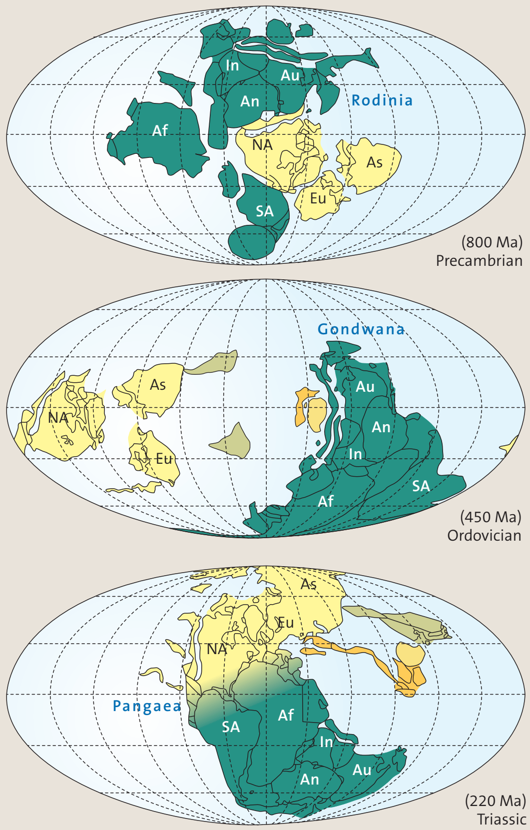
A second geoid bulge occurs in the Pacific, directly opposite to the African-Atlantic bulge, where a conspicuous concentration of hot spots is located. The present Pacific bulge is the broader and higher one of the two. The length of the Pacific bulge corresponds to half of the circumference of the Earth



◀ **Fig. 6.5** Bulge of the present geoid in the area of Africa–North Atlantic with related hot spots, superimposed onto the continental pattern at the end of the Jurassic (Anderson, 1982; map after Ron Blakey, http://jan.ucc.nau.edu/~rcb7/150marect.jpg) .



▲ **Fig. 6.6** Graphic illustration of major landmasses before, during, and after Pangaea. Pangaea lasted from ca. 300 to 175 Ma. The supercontinent Rodinia preceded it in the Late Precambrian. Colors used to show general affinities of continents following major rifting events.

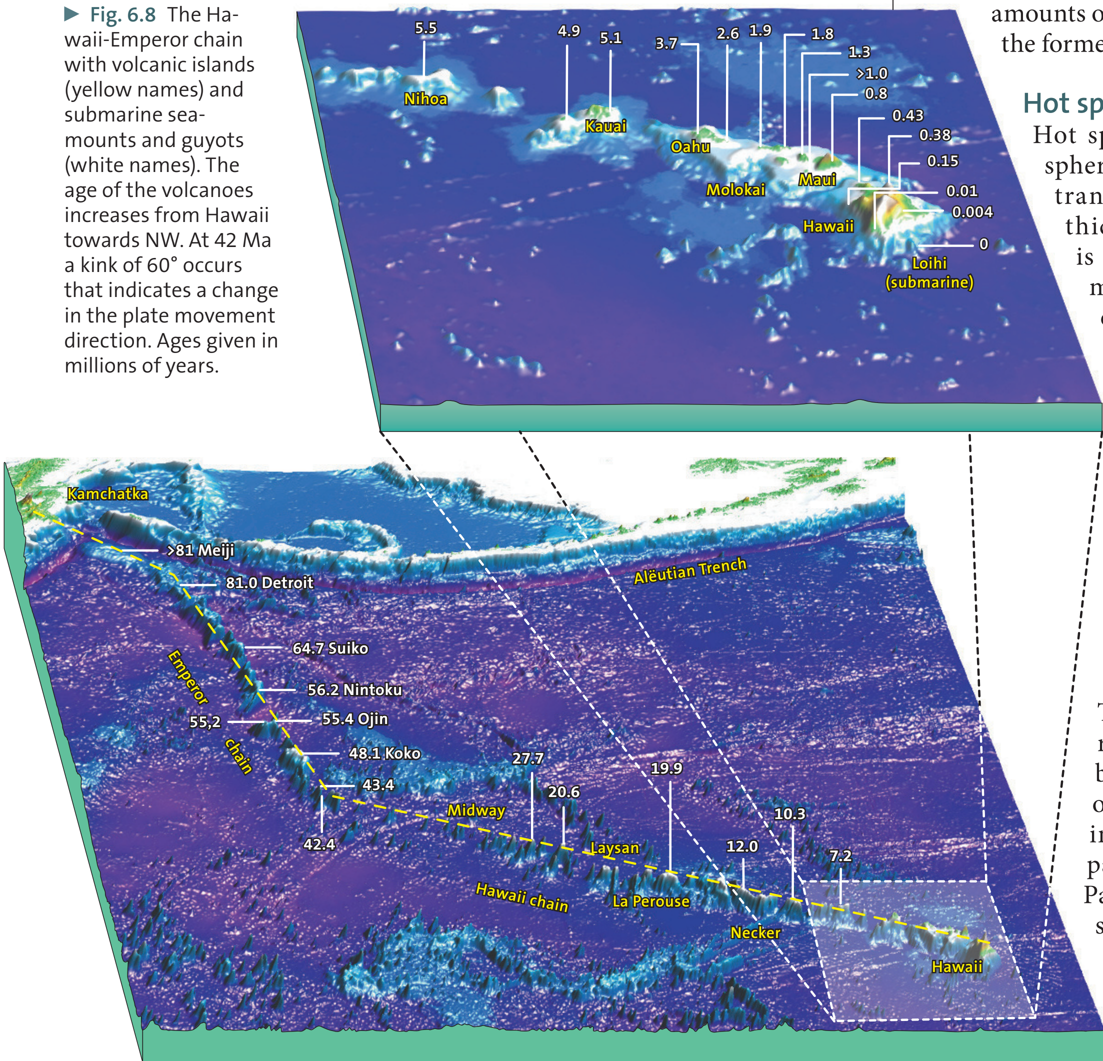


▲ **Fig. 6.7** Reconstruction of the supercontinents Rodinia in the Late Precambrian, Gondwana in the Early Paleozoic, and Pangaea in the Early Mesozoic (Hoffman, 1991). Af = Africa, An = Antarctica, As = Asia, Au = Australia, Eu = Europe (corresponds to Baltica), In = India, NA = North America, SA = South America.

along the equator (**Fig. 6.4**) and thus defines the present location of the pole of the Earth because the axis of rotation is vertically oriented to it. *If* the African–Atlantic geoid bulge, which is stretched in NW-SE direction and only slightly smaller, were the stronger bulge, the axis of rotation of the Earth would be oriented vertically to it through polar wandering – the equator would align with Greenland and Western Europe according to the opinion of some scientists (Chase, 1979).

Supercontinents probably form at intervals of several hundreds of millions of years and then break apart. Pangaea was formed by several continent-continent collisions ca. 300 Ma and gradually broke up 130 m.y. later (**Fig. 6.6**). Pangaea formed when Gondwana (named after the kingdom of the Gond in Central India, i.e., the present Madhya Pradesh), an accumulation of the present southern continents as well as Arabia, India, and Southwest Europe, was welded together at the end of the Precambrian 550 Ma and later combined with North America, Baltica, and Asia (**Fig. 6.7**). Rodinia (after ***rodina***, ***Russian*** fatherland), a supercontinent similar to Pangaea that comprised all large land masses, amalgamated ca. 1000 Ma and disintegrated ca. 750 Ma. The spatial distribution of the land masses in Rodinia was fundamentally different from that of Pangaea (**Fig. 6.7**).

An even older supercontinent, Panotia, may have existed ca. 1400 Ma. The history of supercontinents suggests that they form and break-up in cycles of several hundreds of millions of years duration. As suggested above, the processes of supercontinent formation and destruction are orchestrated by events within the mantle including the activity of hot spots.



▶ **Fig. 6.8** The Hawaii-Emperor chain with volcanic islands (yellow names) and submarine seamounts and guyots (white names). The age of the volcanoes increases from Hawaii towards NW. At 42 Ma a kink of 60° occurs that indicates a change in the plate movement direction. Ages given in millions of years.

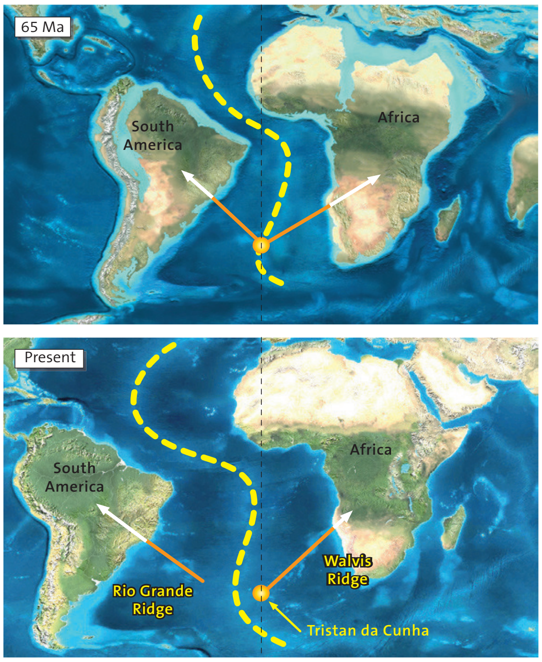
of hot spots (Chase, 1979). The second bulge of the geoid is located in the West Africa–Atlantic region where the concentration of hot spot is also high. Here, the corresponding bulge of the core appears to lie slightly off the geoid bulge (**Fig. 6.4**).

Summarizing these complex relations, it appears that hot zones of the D'' layer correspond with hot zones of the core and that these coincidences produce mantle plumes. Consequently, more heat is transferred into the D'' layer above the hot bulges of the core than is transferred by the core outside the bulges. Moreover, there is an indication that a some amounts of iron from the core are transferred to the D'' layer in such places. This transfer is actually suggested by relatively high osmium contents and a high $^{187}Os/^{188}Os$ ratio seen in some basalts that are derived from hot spots (Walker et al., 1995). The complex processes and features of the core and the adjacent D'' layer suggest that both heat and limited amounts of material are transmitted from the former to the latter.

Hot spot tracks in the ocean

Hot spots penetrate oceanic lithosphere relatively easy whereas the transfer of magmas through the thicker continental lithosphere is retarded. Therefore, hot spot magmas form tracks on the ocean floor in response to the motion of the ocean plate relative to the fixed hot spot. The tracking the hot spots and dating the volcanoes along the track allow the velocity and direction of plate motion to be accurately calculated. The most famous example for such a track is the Hawaii-Emperor chain in the Pacific (**Fig. 6.8**). The mantle plume is presently located below Hawaii. The 6000 km-long track, which runs towards WNW and then bends abruptly NNW, consists of extinct volcanoes of increasing age to the northwest. The pattern demonstrates how the Pacific Plate drifted over the hot spot and during the Eocene at 42 Ma, changed its direction of movement by 60°. The oldest remaining volcanoes in this chain are older than 80 Ma. However, the hot spot is somewhat older because the northernmost volcanoes are already subducted along with the northwestern part of the Pacific Plate beneath Kamchatka.

Similar tracks with a kink are also found in other regions of the Pacific (**Fig. 6.1**) and reveal a consistent picture of the drift of the Pacific Plate since the Late Cretaceous with a change of the movement direction in the Eocene. Other tracks in the Pacific do not show a kink and are thus younger than 42 Ma. The kink is most probably caused by global plate tectonic patterns, specifically a plate reorganization that took place during the Alpine-Himalayan orogeny in the Eocene. The resulting mountain belt extends through Southern Europe and South Asia across more than 10,000 km and was initiated by the collision of Africa and India with Eurasia. Such a major plate tectonic event caused a re-orientation of the plate drift pattern and is reflected on the other side of the world by



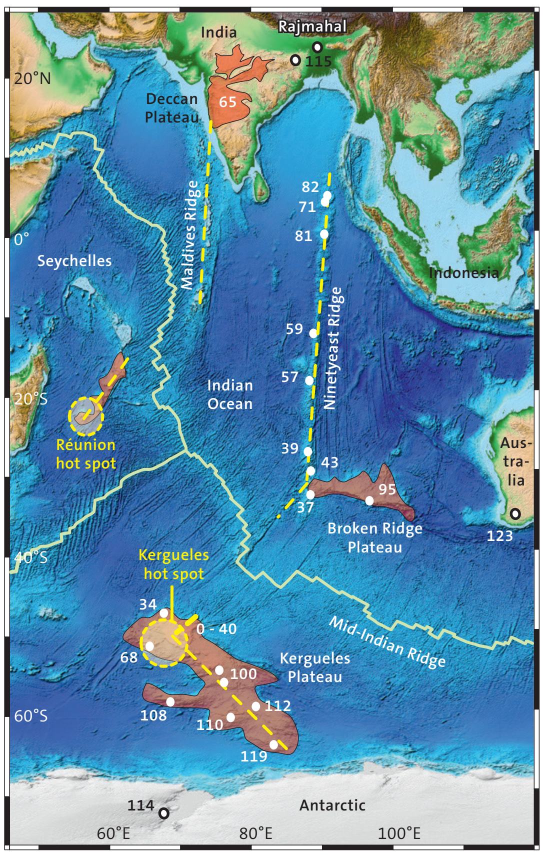
◀ **Fig. 6.9** Early Tertiary and present maps that demonstrate the drifting of Africa and South America and location of the hot spot of Tristan da Cunha. Following a period of symmetric drift when the hot spot lay on the mid-ocean ridge, the plate boundary shifted 30 Ma, and the hot spot tracked across only the African Plate; therefore, the most recent track is restricted to this plate (compare **Fig. 2.6**; Burke and Wilson, 1976).

the abrupt movement change of the Pacific Plate.

Linear volcanic chains that represent tracks of mantle plumes are also found in the Atlantic and the Indian oceans. In the Atlantic, the hot spot of Tristan da Cunha produced two volcanic chains, the Rio Grande Ridge on the South American Plate and Walvis Ridge on the African Plate. Both ridges have northerly components of movement in addition to their west and east motions, respectively. This pattern evolved because the hot spot spent most of its life on the Mid-Atlantic Ridge, ca. 125–30 Ma, and the volcanoes were thus partly built on each plate (**Fig. 2.6**). Approximately 30 Ma in the Oligocene, the drift of Africa slowed significantly although the spreading velocity remained unchanged. Therefore, the Mid-Atlantic Ridge was pushed westward resulting in a stronger westward drift of South America. The hot spot of Tristan da Cunha remained stable and was thus completely enclosed by the westward extending African Plate (**Fig. 6.9**). Presently, Tristan da Cunha is located about 400 km east of the Mid-Atlantic Ridge and produces its track only on the African Plate.

In the Indian Ocean the Maldives Ridge and the Ninetyeast Ridge, which got its name from its position on the 90th east line of longitude, were generated by the hot spots of Réunion and the Kergueles, respectively. Each displays kinks in plate direction, also ca. 40 Ma (**Fig. 6.10**), and both produced voluminous basaltic rocks, especially early in their respective histories. After significant early basaltic production, each continued with sufficient magmatism to leave prominent and rather continuous tracks of plate drift in the ocean.

The hot spot of the Kergueles, the older one of the two, built a plateau with a volume of 20 million km^3 of rocks across an area of 2 million km^2. Over its geologic history that began 120 Ma, the hot spot has had a great influence on the greater Indian Ocean and adjacent continents. Volcanic rocks were especially abundant during its first 30 million years of history and are presently located in the Kergueles Plateau, the Antarctic, the Broken Ridge Plateau, Western Australia, and Rajmahal, India. The basalts of the Rajmahal are not widely distributed and were separated from the occurrences further south by the formation of the early Indian Ocean. The track of the hot spot is found in the Ninetyeast Ridge and records when India and the Antarctic drifted apart. At 40 Ma, the present spreading axis between the Indian Ocean was formed and separated the Broken Ridge Plateau from the Kergueles Plateau. This reorientation of the plate drift pattern in the Eocene caused the kink in the track. The recorded hot spot history of Réunion began with a huge bang as it generated the Deccan Traps of the Deccan Plateau at the Cretaceous-Tertiary boundary 65 Ma. Again, the hot spot track was generated in two episodes,



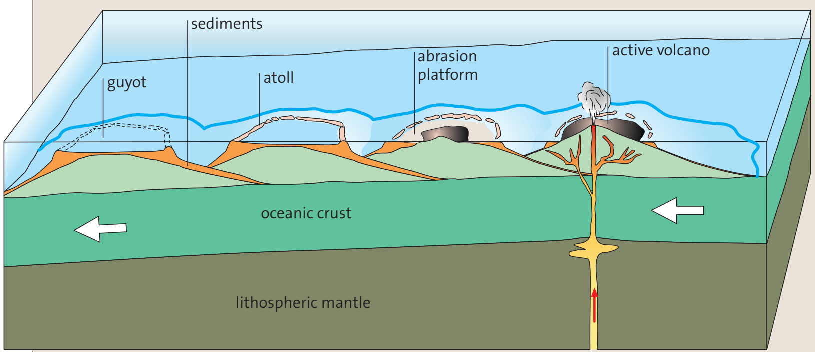
◀ **Fig. 6.10** Tracks of the hot spots of Réunion and the Kergueles. Ages of dated basalts (in Ma) show the series of events that separated India, Australia, and Antarctica (Zhao et al., 2001). At ca. 120 Ma, the three continents were adjacent to the Kergueles–Broken Ridge hot spot. From 115 to 40 Ma, India moved rapidly northward and the Ninetyeast Ridge marks the track of the hot spot on the Indian Plate. At ca. 40 Ma, Australia and Antarctica separated and the hot-spot plateau was severed into two major pieces. The Maldives Ridge records a similar though younger history of movement of the Indian Plate. Note that both hot spots were separated from their tracks on the Indian Plate by the establishment of the present Mid-Indian Ridge around 40 Ma. Their most recent tracks are only on the Antarctic and African plates.

A guyot evolves

One of the most important American geologists in the middle of the 20th century was Harry H. Hess, one of the principle formulaters of the theory of plate tectonics. Hess was a naval officer in the 2nd World War where he was assigned to submarines. Through voyages in the Pacific, he discovered volcanic seamounts in the Pacific. Some seamount summits are 2000 m below sea level, are remarkably flat, and have fringe reefs on their tops (Hess, 1946). He called these peculiar seamounts guyots, after Arnold Guyot, a Swiss geographer who emigrated to America and who, like Hess, taught at Princeton University but a century earlier. Hess concluded from his observations that the summits of these seamounts must have been at sea level earlier in their history. Marine erosion beveled the peak of the volcano and created a flat abrasion platform. During the gradual subsidence of the seamounts, shallow marine sediments and coral reefs were formed on the planed summits. Eventually, the summits subsided too rapidly for coral growth to keep up and the reefs drowned and sank to the depths.

Plate tectonic processes readily explain the evolution of flat-topped seamounts (**Fig. 6.11**). A volcano that develops above a hot spot on ocean crust may grow, as in the case of Hawaii, above the sea level and form an island. As the volcano drifts away from the hot spot area, volcanic activity becomes extinct and erosion exceeds volcanic production. Continued drifting causes the island to slowly submerge because its plate basement also subsides with increasing age. Marine processes bevel the volcanic island to near sea level and create an abrasion platform. In tropical regions, the flat top may be rimmed with coral reefs, usually in a circular pattern. Because the vertical growth of coral can keep up with all but the most rapid rates of subsidence, the fringing coral atoll continues its upwards growth. However, eventually the subsidence, perhaps coupled with rapidly rising sea level, wins out and the reef drowns and subsides passively below the sea. With continuously increasing age of the crust, the eroded, flat-topped volcano sinks deeper into the deep sea.

Seamounts that formed above hot spots in the deep sea but never rose above the sea level maintain their summit as a peak . Volcano chains formed above hot spots may consist of various volcanic islands, guyots, and seamounts – each records the balance between volcanic production, subsidence, and sea level.



◄ **Fig. 6.11** Evolution of a guyot. Stage 1 – the hot spot builds a volcano above sea level that in tropical regions may be fringed by reefs. Stage 2 – the ocean plate moves off of the hot spot and the volcano erodes by marine processes to a flat-topped extinct volcano. Stage 3 – the fringing reef builds around the margin of the truncated volcano. Stage 4 – the volcano subsides and the reef builds upward; eventually subsidence lowers the volcano to depths at which reefs can no longer grow. The now flattened volcano becomes a guyot whose summit can be in water depths as great as 2000 m.

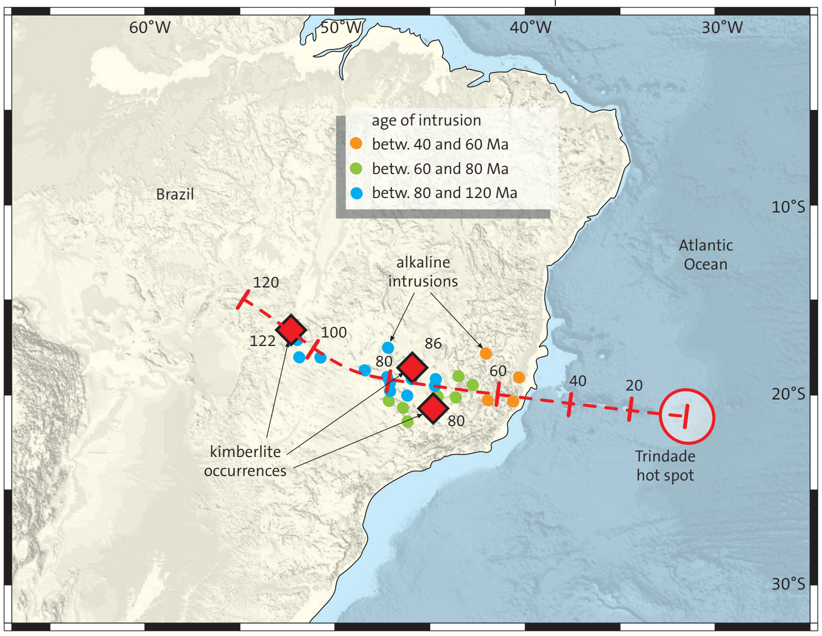
before and after 40 Ma. The shorter tracks of the two hot spots on the African and Antarctic Plate, respectively, resulted from slow motion of these plates relative to the hot spots since the establishment of the actual Mid-Indian Ridge in the Eocene. In contrast, the long hot spot tracks north of the ridge reflect the rapid drift of India northward during the Late Cretaceous and Early Tertiary.

Deviations from the regularity of a plume track are caused by various reasons. A large plume can produce magma at different locations. If a fault zone drifts over the hot spot, the magma might be displaced along this fault zone and diverted across the track. It is also conceivable that the magma might find its way back along the plume track that has been weakened by fractures and older supply channels. Magma chambers can form and drift within the lithosphere and extrude at a later time. Such behavior could disrupt the linear sequence of volcanoes. The three long and kinked hot spot tracks in the Pacific (**Fig. 6.1**) do not coincide exactly. The two southern tracks display irregularities in their volcanic sequence that may be related to one or the other of the mechanisms described above.

Hot spot tracks on the continent

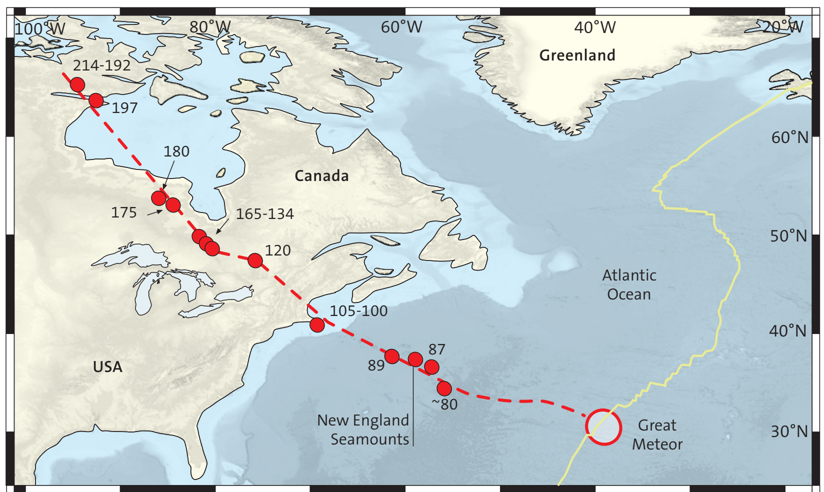
The tracks of hot spots are more difficult to follow across continental crust where they are less distinctive and commonly not marked by volcanic chains. The thick continental lithosphere is difficult for the magmas to penetrate and disruptions of magma paths and changes of magma composition make detection more difficult. However, examples of hot spot tracks across continental terrain do exist. The hot spot of Trindade in the South Atlantic left a track during the Cretaceous when the Brazilian Shield

▶ **Fig. 6.12** Track of the hot spot of Trindade that today is in the South Atlantic adjacent to the Brazilian coast (Crough et al., 1980). The track is marked in Brazil by occurrences of kimberlite and alkaline intrusions that decrease in age from west to east. The dashed red line tracks the position of the hot spot as reconstructed from the global plate drift pattern. Ages are given in million years.



drifted over the hot spot (**Fig. 6.12**). Subsequent erosion has removed the volcanoes but has left a track of originally shallow alkaline intrusions, some diamond-bearing kimberlite occurrences amongst them. Kimberlites are formed in diatremes (kimberlite pipes) that originate in the sub-lithospheric mantle; many are thought to be related to hot spots. The magmatites of the Trindade track decrease in age towards the present position of the hot spot. The positions of the hot spot below the South American continent as reconstructed from the plate drift history and the age of the intrusions coincide well even though the magmatism above hot spots in continental areas scatters across a larger radius.

The track of the Mesozoic Great Meteor Plume was imprinted across the North American continent where it is marked by kimberlites and other alkaline magmatic complexes. Only magmatic rocks that formed deep in the crust are preserved. As North America moved across the hot spot for 100 m.y. during the Jurassic and Early Cretaceous, a 4000 km-long track was formed that extends from Hudson Bay to New England (**Fig. 6.13**). As it migrated off shore, it produced the New England Seamount Chain in the NW Atlantic. Some of these volcanoes were islands earlier in their history but today their summits are in water depths of more than 1500 m. The Great Meteor Plume is presently located at the Mid-Atlantic Ridge.



▲ **Fig. 6.13** Track of the hot spot of the Great Meteor Plume on the North American Plate as reconstructed from dated magmatic rocks (Van Fossen and Kent, 1992; Heaman and Kjarsgaard, 2000). Ages in million years.

Two mantle plumes located beneath the European continent, the hot spots of the Central Massif in France and the Eifel in Germany (**Fig. 6.1**), did not form a track. Both are considered to be the product of a shallow source in the upper mantle and not of deeper mantle origin (Granet et al., 1995); each was relatively short-lived. The activity in the Central Massif culminated in the formation of two volcanoes, the Cantal (11 to 2.5 Ma) and the Mont Dore (4 to 0.3 Ma). The last eruption occurred 4000 years ago. Although the volcanism is considered to be extinct, this may turn out to be a misjudgement in the long term.

The volcanoes of the Eifel were active in the Early Tertiary, especially between 40 and 25 Ma. Activity recommenced in the Quaternary ca. 0.6 Ma and the youngest eruptions are dated ca. 10,000 years old. The formation of NW-SE striking rows of volcanoes corresponds to the alignment of fissure systems (Schmincke, 2004) that are related to the NE-SW extension of the Lower Rhine Embayment (Ch. 3). The alkaline melts are dominantly of basaltic composition and only slightly differentiated. Contacts of the ascending magmas or the already solidified hot magmatites with ground water led to explosive events (phreatomagmatic resp. phreatic eruptions; *phreas*, *Greek* fountain, water container) accompanied by thick deposits of tuff. These volcanic deposits comprise mostly cinder cones and highly explosive maars and are well exposed in the numerous quarries of the Eifel region. These are phreatic pipes (diatremes) with ring walls of pulverized rock material that was ejected around the explosive crater without the eruption of magma. The craters are the sites of scenically attractive lakes. Presently the volcanism is dormant but it is not considered extinct as testified by frequent escape of magmatic carbon dioxide.

Flood and trap basalts

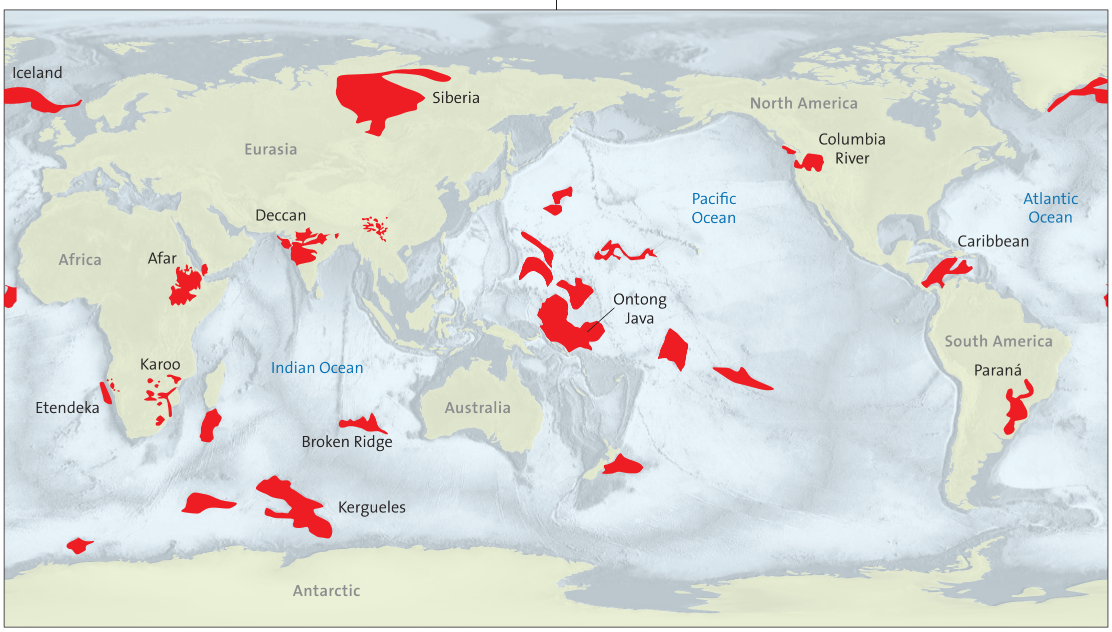
Flood basalts, also called large igneous provinces (LIP), are characterized by layered basaltic deposits several kilometers thick that form huge plateaus on land as well as below sea level. On land the basaltic plateaus typically erode at their edges into distinctive stair steps. The steps are caused by erosion of the intersecting horizontal bedding planes between lava flows and the vertical contraction joints that develop on cooled lava sheets. Erosion takes advantage of the horizontal and vertical planes to form the distinctive landscape. The term "traps" or "trap basalts" was introduced by Carl von Linné, Axel Fredrik Cronsted, and Johan Gottschalk Wallerius for the Deccan basalts, India, in the 18th century – *trappa* is *Swedish* for stair (Faujas de Saint-Fond, 1788).

Typically huge eruptions involve millions of cubic kilometers of basaltic rock that is extruded during an interval of only one or several million years. The hot basalts that emanate from fissures or erupt as lava curtains have a low viscosity that enables successvie flows to cover areas as large as several hundred-thousand square kilometers. Multiple layers of basalts several meters thick stack up to form widespread plateaus. During brief periods of magmatic quiescence, soils can develop on top of the basalts to form prominent stratigraphic horizons within the sequence.

Large eruptions of flood basalts have occurred throughout Earth history (**Fig. 6.14**). Notable examples include:

- The Siberian trap basalts formed 250 million years ago at the Permian/Triassic boundary with a volume of 2.5 million cubic kilometers.
- The hot spot of Tristan da Cunha (see above) that contributed to the opening of the South Atlantic, created the Paraná flood basalts in South America and the Etendeka basalts along the Southwest African coast. Each is associated with alkaline magmatic complexes that formed before they were separated by continental break-up in the Early Cretaceous.
- The Deccan trap basalts in India are related to the break-up of part of Pangaea 65 Ma where the hot spot of Réunion produced the enormous eruptions that now form the Deccan Plateau (**Fig. 6.10**). These eruptions splintered India from a continental remnant that today is represented by the Seychelles Islands in the Indian Ocean. Paleomagnetic investigations indicate that the Deccan traps were formed more than 30 degrees farther south than their present position which confirms the location of origin as close to the hot spot of Réunion. The Deccan traps erupted in less than one-half million years with a volume greater than 2 million cubic kilometers.

When continental crust is strongly thinned by the rifting process, hot asthenosphere rapidly ascends. If magma was already enriched in the head of a mantle plume, additional melting of the mantle occurs because of pressure release. Huge quantities of basalt pour out across the continental margin and the adjacent oceanic depression in a geologically



▶ **Fig. 6.14** Global distribution of large igneous provinces in the oceans (submarine plateaus) and on land (flood basalts).

very short time interval. The eruptions during this process occur in phases. Calculations indicate that many cubic kilometers of basalt may be produced per year. Therefore, the magma production of one hot spot may exceed the total production along the global mid-ocean ridge system that is presently ca. 20 km^3 per year.

Large submarine basaltic plateaus are formed within intraplate oceanic settings. Basaltic extrusions at intra-ocean hot spots can thicken the basalt/gabbro oceanic crust to thicknesses up to 30 km. The resulting large plateaus rise by up to 3000 m above the surrounding deep sea floor. The largest extant oceanic plateau is the Ontong-Java Plateau east of the Bismarck Archipelago in the Western Pacific (**Fig. 6.14**). It covers over two million square kilometers and includes up to 30 million cubic kilometers of gabbro and basalt that ascended from a plume (Coffin and Eldholm, 1993). Additional examples of large submarine basaltic plateaus include the Kergueles Plateau in the Indian Ocean (see above) and the Caribbean Large Igneous Province. All three of these plateaus were formed during the superplume event in the Cretaceous (see below) during several episodes that individually lasted for only a few million years.

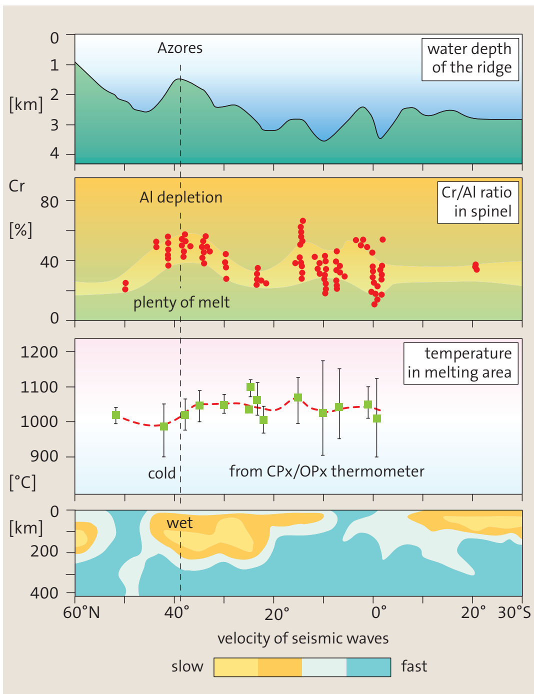
The Azores – hot, cold or wet spot?

The Azores are a group of volcanic islands near the Azores Triple Junction where the transform fault representing the plate boundary between Eurasia and Africa branches off from the Mid-Atlantic Ridge towards east (**Fig. 1.5, 6.1**). For years the Azores were considered to be a hot spot. Detailed investigations indicate that the mantle immediately below the Azores is not hotter but rather cooler than other segments of the Mid-Atlantic Ridge (Bonatti, 1994).

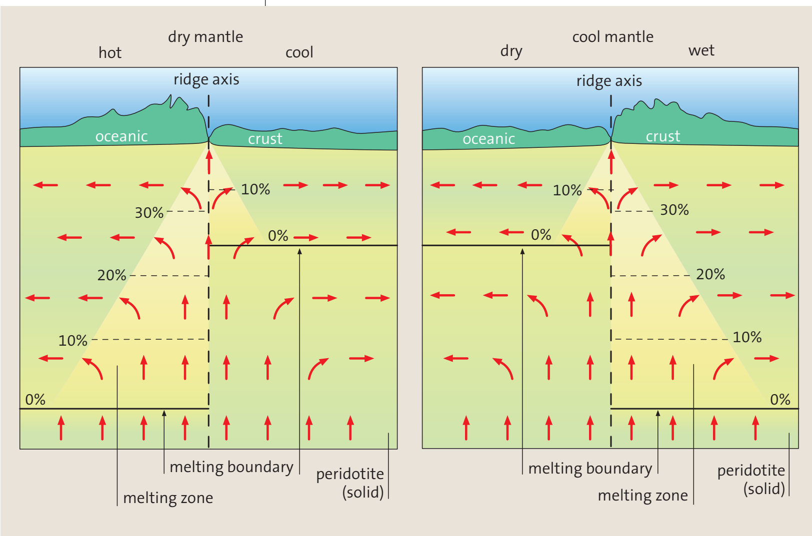
Peridotites were examined from the lithospheric mantle directly below the crust along the Mid-Atlantic Ridge from the equator to northern latitudes (**Fig. 6.15**). The basalt formed from upper mantle peridotite by partial melting (Ch. 5). Clinopyroxene (diopside) is easier to melt than the other minerals of the peridotite and thus preferentially goes into the melt. Also, the elements iron and aluminum preferentially go into the melt whereas magnesium and chrome mostly remain in the peridotite. The peridotite is thus depleted by partial melting in Fe and Al and enriched in Mg and Cr. This process is recorded in the mineral spinel that occurs in low quantities in peridotite. Generally, mixed crystals between chrome and aluminum spinel are formed. From the Cr/Al ratio of the spinel, it can be estimated how much basaltic melt was extracted from the peridotite. The electron microprobe was used to measure the exact chemical composition of the minerals.

The profile along the Mid-Atlantic Ridge indicates that the Cr/Al ratio of spinel is low in the equatorial area but high in the area of the Azores (about 39° N; **Fig. 6.15**). This confirms that low degrees of partial melting prevailed in the equatorial Atlantic (therefore, the oceanic crust at the Vema Transform Fault is of low thickness; **Fig. 5.12**) whereas around the Azores, a high degree of partial melting is assumed. In the area of the Azores, a 25 % partial melting of the original rock is to be expected; other segments of the profile show values between 10 and 20 %. These results support the theory of a hot spot below the Azores.

Meanwhile, temperature of the peridotites during the processes of partial melting and subsequent cooling can be determined with the aid of geothermometers. Within mineral pairs in rocks, an exchange of elements occurs and a thermodynamic equilibrium according to the current temperature develops. For example, the distribution of Mg and Fe in orthopyroxene and clinopyroxene can be used as geothermometer. Corresponding measurements in peridotites along the Mid-Atlantic Ridge indicate that the area of the Azores is not hotter than the other ridge segments but has slightly lower temperatures (**Fig. 6.15**). Is this a cold spot? But how does one explain the high percentage of partial melting that is responsible for the high elevation of the ridge in the area of the Azores?



◀ **Fig. 6.15** Topographic profile along the Mid-Atlantic Ridge between 60° N and 30° S latitude and conditions in the mantle below (Bonatti, 1994). Beneath the Azores, the formation of basaltic melts is high; the temperatures, however, are low in the area of melting. Seismic investigations show that the mantle below the Azores is rich in watery fluids. Cpx = clinopyroxene, Opx = orthopyroxene.



▲ **Fig. 6.16** Amounts of partial melting of the ascending mantle peridotite if it is dry and hot, dry and cool, cool and "wet" (Bonatti, 1994). The percentages indicate the portion of basaltic melt extracted from the peridotite; melt proportion increases during the ascent of the peridotite because of pressure release and related lowering of the melting point. Watery fluids ("wet" mantle) reduce the melting point of peridotite and thus increase the degree of partial melting.

The explanation is that the melting point of rocks is not only dependent on the rock composition and temperature but also on the pressure and the content of fluids. As already discussed, the degree of partial melting increases when the mantle current ascends because the pressure release reduces the melting point (**Fig. 6.16**). Fluid phases included in the rock act in the same direction. Fluid phases are liquid or gaseous components that occur in practically all rocks; however, they occur in different amounts, and are dissolved in magmas. Magmatic rocks containing very little fluid phases are described as "dry", and those with a lot of fluids (in particular water) are classified as "wet". According to chemical analyses, basalts of the Azores contain three to four times the amount of water as normal basalts of mid-ocean ridges indicating a "wet" melting region. This is corroborated by seismic investigations (**Fig. 6.15**). The seismic waves are relatively slow in the uppermost mantle in the area of the Azores, a characteristic attributed to high contents of fluids.

Therefore, the substantial production of basaltic melts in the area of the Azores Triple Junction is not related to a hot spot. But this area is not a cold spot either! Low temperatures can not be responsible for the high magma production. The critical feature is the high content on fluid phase in the mantle rocks that effectively reduces the melting point of the peridotite. At given temperatures between 1000 and 1100 °C, large portions of basaltic melt develop that are otherwise only possible with higher temperatures (**Fig. 6.16**). Additionally, with melting, a loss of heat develops in the remaining peridotite and may explain the somewhat lower temperatures according to the geothermometer. Wet areas are not rare in the mantle. Water is permanently transported into the mantle in subduction zones as water-containing sediments or within basalts whose original dry minerals are converted into water-containing minerals. Water may even come from the mantle itself by methane degassing from the deeper mantle that is oxidized to water and carbon.

Hawaii – a typical oceanic hot spot

The main island of Hawaii, the locus of most recent volcanism in the Hawaiian chain, has been connected to the hot spot plume for less than one million years (**Fig. 6.8**). The resulting volcanic complex rises at the peak of Mauna Loa to more than 4000 m above the sea level and has its base at a water depth of more than 5000 m below sea level and thus, with nearly 10,000 m of relief forms the tallest mountain on Earth measured from its base. Its diameter at the ocean floor is ca. 1000 km, roughly corresponding the diameter of the plume head below. In spite of the fast drift of the Pacific Plate, the magma production was rapid enough to produce the present mountain with a volume of 40,000 km^3 (Schmincke, 1981). This volume would cover Switzerland with a 1 km thick layer of basalt; however, such an event probably would be prevented by a referendum! Geophysical investigations indicated that a pear-shaped magma chamber is located between 3 and 10 km below the summit of the Kilauea. This magma chamber is fed by a 30 km-long vertical, tube-like feeder channel that terminates in the mantle.

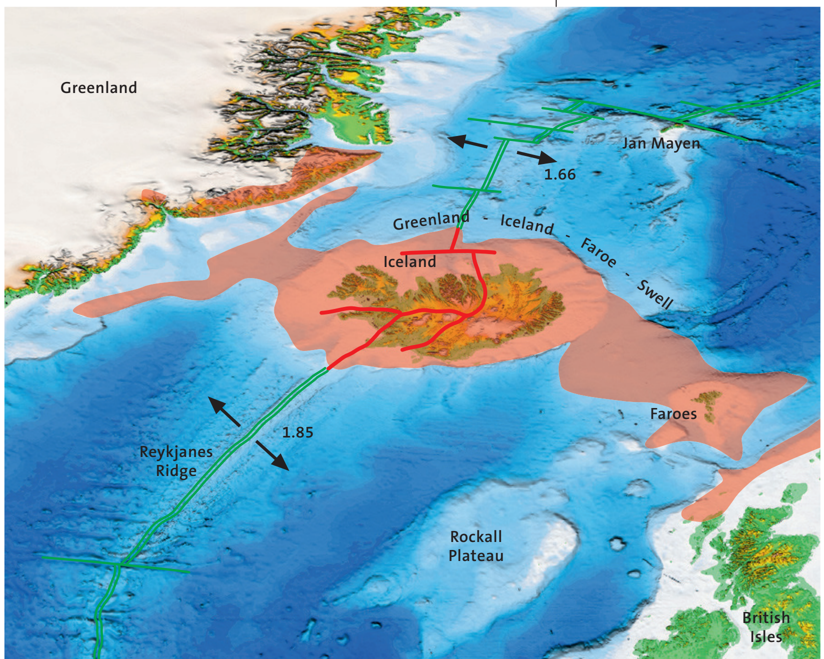
The Island Hawaii will drift out of the area of influence of the mantle plume within the near geological future because the Pacific Plate drifts over the plume at a rate of ca. 100 km in one million years. Approximately 30 km southeast of Hawaii is a new volcano, Loihi, currently under construction (**Fig. 6.8**). Therefore, the plume is not currently located directly beneath Hawaii. The summit of Loihi is presently at ca. 1000 m below sea level.

Islands such as Hawaii typically evolve in a predictable fashion. Incipient submarine volcanoes are built on the sea floor above hot spots and comprise pillow lavas and massive dikes and sills. As the lavas build vertically and gradually approach the surface of the water, gas in the magma can blast through water to the surface and result in volcanic explosions that intensify as seawater penetrates into fissures and joints and is mixed with magma in shallow magma chambers. The explosions generate volcanic ashes that solidify to tuff or are transported into deeper areas where they form aprons of volcanic fragments. If the production of lava continues, the volcano builds a large oceanic island that is composed of a sequence of tuffs and lavas with diagonally cutting dikes that represent the solidified feeder channels. The island of Hawaii is currently in this mature stage of development; however, as the island moves off the locus of the hot spot, erosion processes will take over and eventually reduce the island to a low-lying pile of volcanic rocks and debris.

Iceland

Slow plate motions over active hot spot generate large amounts of basalt. This is the case in Iceland where the hot spot is directly at the Mid-Atlantic Ridge and the spreading rate is low (ca. 1 cm/year in each direction; **Fig. 6.17**). The coincidence of a large production of basalt from the hot spot and a slow plate drift results in an elevated mid-ocean ridge that in Iceland is exposed on land for greater than 300 km. Iceland is currently one of the most active hot spots on Earth with a plume head of probably 1000 km diameter. The crust was thickened to more than 25 km by the intense magmatism and thus raises the question about the possibility of a continental splinter below Iceland. A negative gravity anomaly appeared to support this assumption. Seismic investigations demonstrate that the Icelandic crust is basaltic although acidic volcanic rocks (rhyolites) also occur. The rhyolites are differentiates of the basaltic magmas and not developed from the melting of continental crust as it is frequently observed at continental hot spots. Approximately 8 % of rhyolitic melts may form from a basaltic magma by differentiation during cooling (Schmincke, 2004). The negative gravity anomaly of Iceland is thus not related to lighter continental crust but to the anomalously hot and molten, and thus relatively light, uppermost mantle, the head of the plume and the MORB source. The thick basaltic crust and the large plume head explain the high elevation of the 100,000 km^2 island.

Iceland is a product of the activity of a mantle plume, which ca. 70 Ma (Late Cretaceous) was located below Western Greenland and the Canadian Arctic and probably contributed to the aborted separation of North America and Greenland (**Fig. 4.9**). Greenland then drifted over the hot spot, which 55 Ma was located below Eastern Greenland. At that point, the opening of the Atlantic between Greenland and Scandinavia initiated. Eruption of basalts occurred in pulses that lasted one to three million years and recurred at intervals of 5 to 10 m. y. (O'Connor et al., 2000). Basalts derived from the hot-spot acitivity are found in Eastern Greenland, along the Greenland-Iceland-Faroe-Swell in the North Atlantic, at the northwestern edge of the British Isles, and at the continental margin of Norway (**Fig. 6.17**). These basaltic centers make up the North Atlantic Volcanic Province. The young mid-ocean ridge came under the influence of the mantle plume in the Eocene ca. 40 Ma. Since that time, the Greenland-Iceland-Faroe-Swell was formed as a diagonal ridge by the hot spot on both sides of the spreading axis.

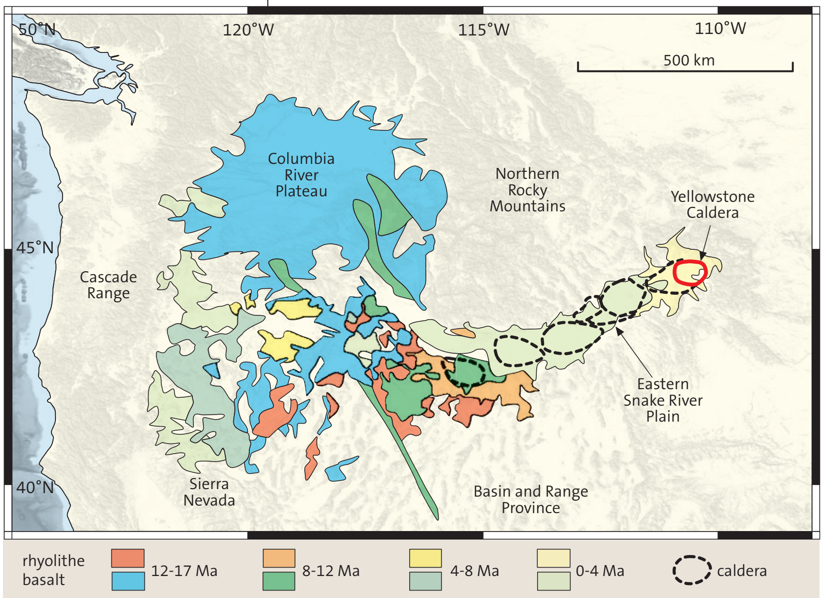


▲ **Fig. 6.17** Map of the Greenland-Iceland-Faroe-Swell. This topography evolved because the Iceland hot spot was located below the young mid-ocean ridge of the North Atlantic 40 Ma. Red shading marks basaltic rocks produced by the hot spot. Numbers indicate the spreading rates in cm/year.

Yellowstone

The Yellowstone hot spot appeared ca. 17 Ma in the Miocene and has left a track as the relatively slow-moving, WSW-moving North American Plate drifted over it at 4 cm/yr. The track has left major volcanic centers that shift to the ENE. The earliest record of volcanism occurs in the Columbia River Plateau and comprises chiefly basalts but important rhyolites as well (**Figs. 6.18, 6.19**). This acidic volcanism arises because the upwelling magmas pass through thick continental crust.

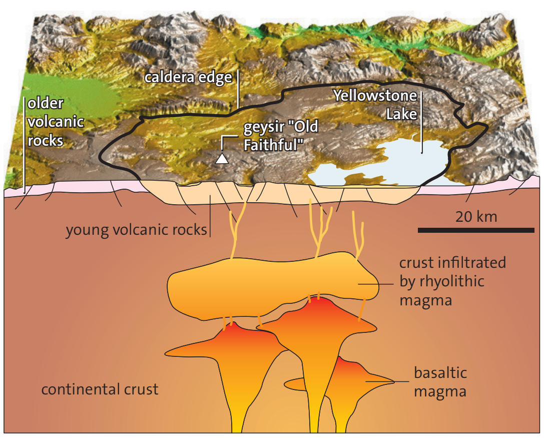
In fact, the Yellowstone hot spot is currently the most prolific producer of acidic magmatic rocks on Earth. Compared to basalts, rhyolitic magmas are highly explosive and produce violent eruptions. The explosive nature is a result of the high viscosity of the magma and water dissolved in the magma – during the explosion, water is abruptly released in a gaseous state. These types of volcanic eruptions drain very large magma chambers in a very short time span; the emptied chamber immediately collapses to produce *calderas* (*Spanish* cauldron), bowl-shaped craters that commonly fill with water to form a lake. Because of the explosive nature of the volcanism, the vast majority of the rhyolites are erupted as tuffs.



▲ **Fig. 6.18** The bimodal (basaltic and rhyolitic) volcanic rocks along the track of the Yellowstone hot spot (Christiansen et al., 2002). Because of the plate drift of North America towards WSW, the volcanic rocks decrease in age towards ENE. Collapse craters (calderas) with diameters of several tens of kilometers developed in areas of large rhyolite eruptions.

The heat supply from the Yellowstone mantle plume has repeatedly generated large volumes of rhyolite melts that have resulted in violent eruptions and the formation of huge calderas. During the history of the hot spot, nearly 150 highly explosive eruptions have been recorded. The youngest, Yellowstone Caldera, is oval and has dimensions of 75 × 45 km. The last three large eruptions occurred in regular intervals at ca. 2.1 Ma, 1.3 Ma, and 0.64 Ma with a production of up to 2500 km^3 of tuff during one event. It is easy to calculate that another such catastrophic eruption is to be expected in the geologically near future and the devastating consequences will not be restricted to the direct surroundings of the center of eruption.

During the eruption at 0.64 Ma, the volcano edifice collapsed several hundreds of meters. The caldera was filled with volcanic lavas and lake deposits.



▶ **Fig. 6.19** Schematic block diagram of the Yellowstone Caldera and the magmas below.

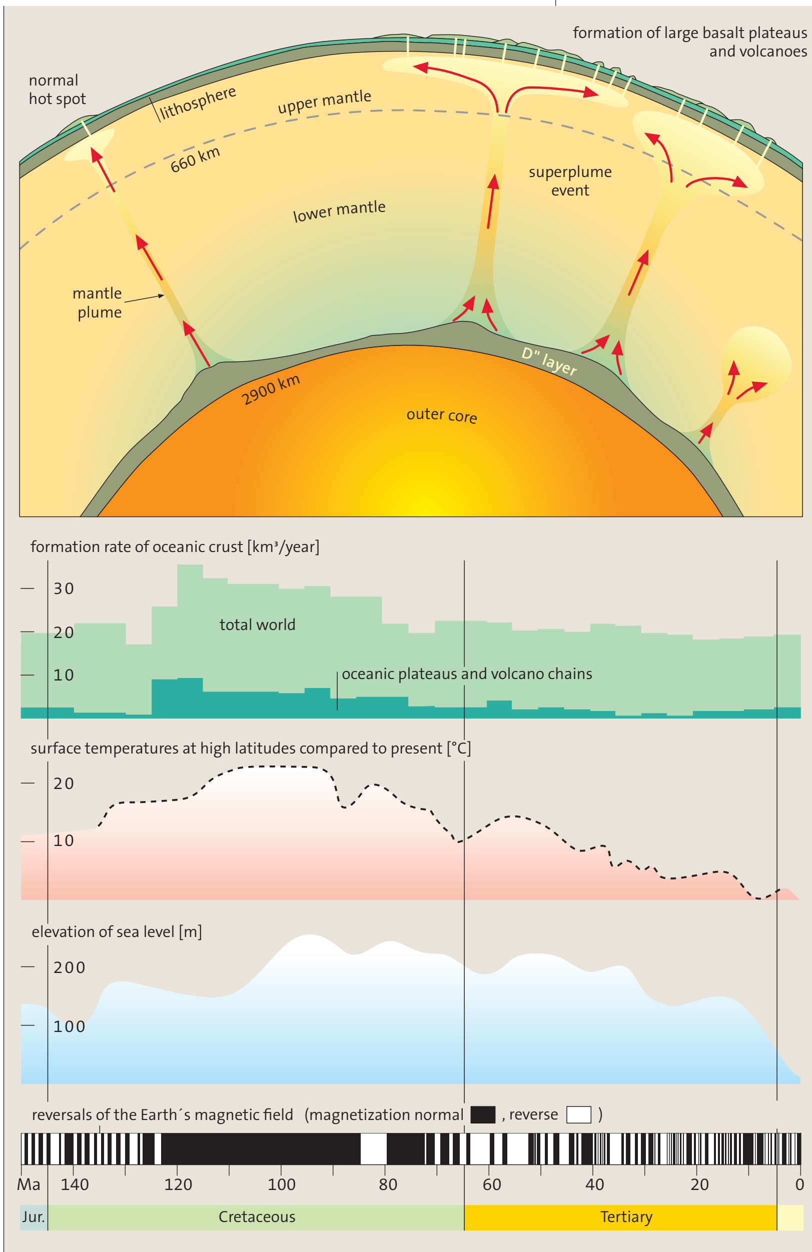
Collapse was followed by the formation of dome-like bulges, an indicatation that magma chambers at depth were again filled. At present, the surface continues to pulsate suggesting magmatic agitation in the depth. Because of the permanent heat transfer from the hot basaltic magma chambers to the surrounding crustal rocks, increasing amounts of explosive acidic magmas develop. If the internal pressure becomes excessive, a devastating eruption will occur because the pressure release during the explosion abruptly releases all gases within the melt. According to seismic investigations, the actual magma chamber is swollen over a distance of 40–50 km and a thickness of up to 10 km (**Fig. 6.19**). In addition to volcanic destruction across a wide area, a new eruption would have global climatic consequences. A similar huge eruption at Toba, Sumatra 0.74 Ma formed a caldera 100 × 60 km. A global drop in temperature of ca. 5 °C drastically reduced the population of mankind at that time. However, Toba is not a volcano located above a hot spot but rather above a subduction zone.

The superplume event in the Cretaceous

The Cretaceous from ca. 125 Ma to 85 Ma was a time of extremes. According to a theory of Roger L. Larson, extreme conditions were caused by the exceptionally high activity of mantle plumes that were expressed in many large and unusually productive hot spots on the surface. This event is called the "Cretaceous superplume event" (Larson, 1995). Large igneous provinces (LIP) generated by superplume events apparently occur at irregular intervals throughout geologic time, although two large LIP, the Siberian Traps at the Permian/Triassic boundary and the Deccan Traps at the Cretaceous/Tertiary boundary, are not associated with superplume events.

The high activity of mantle plumes in the Cretaceous is the result of a heat accumulation in the boundary zone between the core and mantle, the place of origin of the large mantle plumes. Overheating was caused by convection currents in the outer core. A persistent and stable convection current pattern could be the reason that the boundary zone to the mantle (the D'' layer) was extensively heated at different locations and finally led to heat accumulation at various global locations to subsequently ascend from hot mantle material in the form of plumes (**Fig. 6.20**). As the plume heads reached the base of the lithosphere, they were converted into particularly large mushroom-like heads that provided the intense, abrupt magmatic activity.

The superplume activity not only increased the activity of hot spots within the plates, but also those at mid-ocean ridges where the magma production



▶ **Fig. 6.20** The superplume event in the Cretaceous and some of its global effects (Larson, 1995). Shown are rates of formation of ocean crust (at hot spots and mid-ocean ridges), high-latitude temperatures, and sea level elevation, each compared to the present. Also shown is the long normal magnetic polarity period that lasted 40 m.y. This period may have been caused by a stable convection current pattern in the liquid outer core that was related to the superplume event.

increased significantly. As a consequence of the strong magmatic activity within the plates, as well as at the mid-ocean ridges, new oceanic crust, highly elevated oceanic plateaus, oceanic islands, and seamounts were rapidly formed. In the mid-Cretaceous, the Pacific Ocean was a place of enormous expansion of basaltic crust. Both these trends continue to the present. Globally, almost all submarine high plateaus and a large number of seamounts were formed in the Cretaceous superplume period. Approximately 125 Ma, the total production of oceanic crust increased, within a time interval of only a few million years, from around 20 km^3 per year (this corresponds more or less to the present value) to about 35 km^3 per year (**Fig. 6.20**).

The intense magmatic activity had global effects on other geologic and climatic processes. Large amounts of gases that were dissolved in the magma were released to the atmosphere, especially the well-known greenhouse gas carbon dioxide (CO_2). This caused an additional increase of the global mean annual temperature that already was at a higher level than today. The polar regions were particularly affected by the warming (**Fig. 6.20**). The time interval between 115 and 90 Ma was presumably the warmest period since the Precambrian.

Rapid sea-floor spreading generated broad, large-volume mid-ocean ridges. The ridges, the oceanic plateaus and the hot-spot volcanoes caused a decrease in volume of the ocean basins thereby forcing sea-water onto the continents. Also, few or no glaciers were present so little water was tied up in ice (today's sea level is depressed some 80 m by extensive Antarctic and Greenland glaciation). The result was the highest sea level since the Ordovician, as much as 250 m above present sea level (**Fig. 6.20**). Shelf areas expanded as coastlines retreated towards the land. Approximately 93 Ma during the Early Turonian division of the Cretaceous, the zenith was reached.

The broad shelf areas and the warm, equitable climate led to an enormous increase in the biomass of organisms that produce calcareous shells. The calcareous production is particularly high in shallow warm water. Therefore, shallow water limestones from the Cretaceous period are especially widespread. The increased amount of carbon to build calcite ($CaCO_3$) in the form of limestone from the organic carbon in living things was provided by the carbon dioxide that was released to the atmosphere and the oceans by the basaltic volcanism. Among the most impressive and beautiful landscapes built from this sedimentary formation are the Cretaceous chalk rocks of Dover (Great Britain) and Rügen (Germany), and the chalk bluffs from the Dakotas to Texas (USA); each is composed of the remnants of calcareous microfossils. But the rich habitat also produced organic matter that could not be decomposed completely because of the lack of oxygen in the sediments; this led to the formation of bitumen-rich clays that matured to black shales rich in petroleum. Mid-Cretaceous black shales were formed repeatedly and represent important petroleum host rocks. Oil deposits thus derived from Cretaceous black shales comprise nearly half of the known global petroleum reserves. Another giant storehouse of Cretaceous carbon occurs in the huge coalfields of the Western Interior of North America. Coal formed in coastal plain settings when the rapid accumulation of plant material greatly exceeded the rate at which it could be oxidized; the coals now comprise one of the greatest energy reserves in the world.

The increased formation of oceanic crust at mid-ocean ridges must have been compensated by a faster removal of ocean floor in the subduction zones. Subduction produces a characteristic magmatism above the subduction zone. The faster the subduction rate, the greater are the amounts of subduction-related magmatites. Therefore, thick bodies of Cretaceous plutonic and effusive rocks are found in the core of the Andes, the North American Cordillera and other mountains around the Pacific. The 110–80 Ma magmatic event in the Sierra Nevada and adjacent batholiths represents one of the most concentrated plutonic events in the geologic record.

Curiously, coincidentally with these other extreme events, most of the diamond deposits of the world evolved. Diamonds are, except from the pressure developing at meteorite impacts, only stable under the high pressure of the mantle and occur at depths greater than 100 km. Diamonds were formed in the mantle over eons of time and were transported to the surface by the Cretaceous plume magmatism. They reached the surface when magma from the upper mantle rapidly ascended through pipes to form kimberlites, the host rock of diamonds. Diamonds are commonly found in river placer deposits because they can withstand long river transportation due to their hardness.

Yet another extreme phenomenon that seems to have a causal relation to the superplume event is the mid-Cretaceous magnetic normal period. The reversals of the Earth's magnetic field occur successively in very irregular time intervals from some thousand to several tens of millions of years. The magnetic field of the Earth is generated within the outer core that is mostly composed of iron. Convection currents in this liquid layer and electric currents caused by the convection play the decisive role. The reversals of the magnetic field are not understood completely, but turbulent currents from the outer core may cause more frequent reversals than convection currents that are stable over long time periods. The intense and fast convection currents demanded for superplume periods also stabilize the magnetic field over long time periods and reversals do not occur (Larson, 1995). The removal of hot mantle material by the plumes maintains the steep temperature gradient across the core/mantle boundary which in turn stabilizes the convection in the outer core.

The mid-Cretaceous is the time interval with the longest span without a reversal in the Phanerozoic (**Fig. 6.20**). Over some 40 million years, the magnetic field maintained a polarity that was oriented in the same sense as the current normal magnetization. During the Late Cretaceous and Early Tertiary, the frequency of reversals increased gradually until such events became very frequent in the last 30 million years, typically occurring several times within one million years. The frequent reversals indicate slow and turbulent currents in the outer core. This increase in reversal frequency is coincident with the onset of icehouse conditions in the Early Oligocene. This is in sharp contrast to the greenhouse conditions of the Cretaceous.