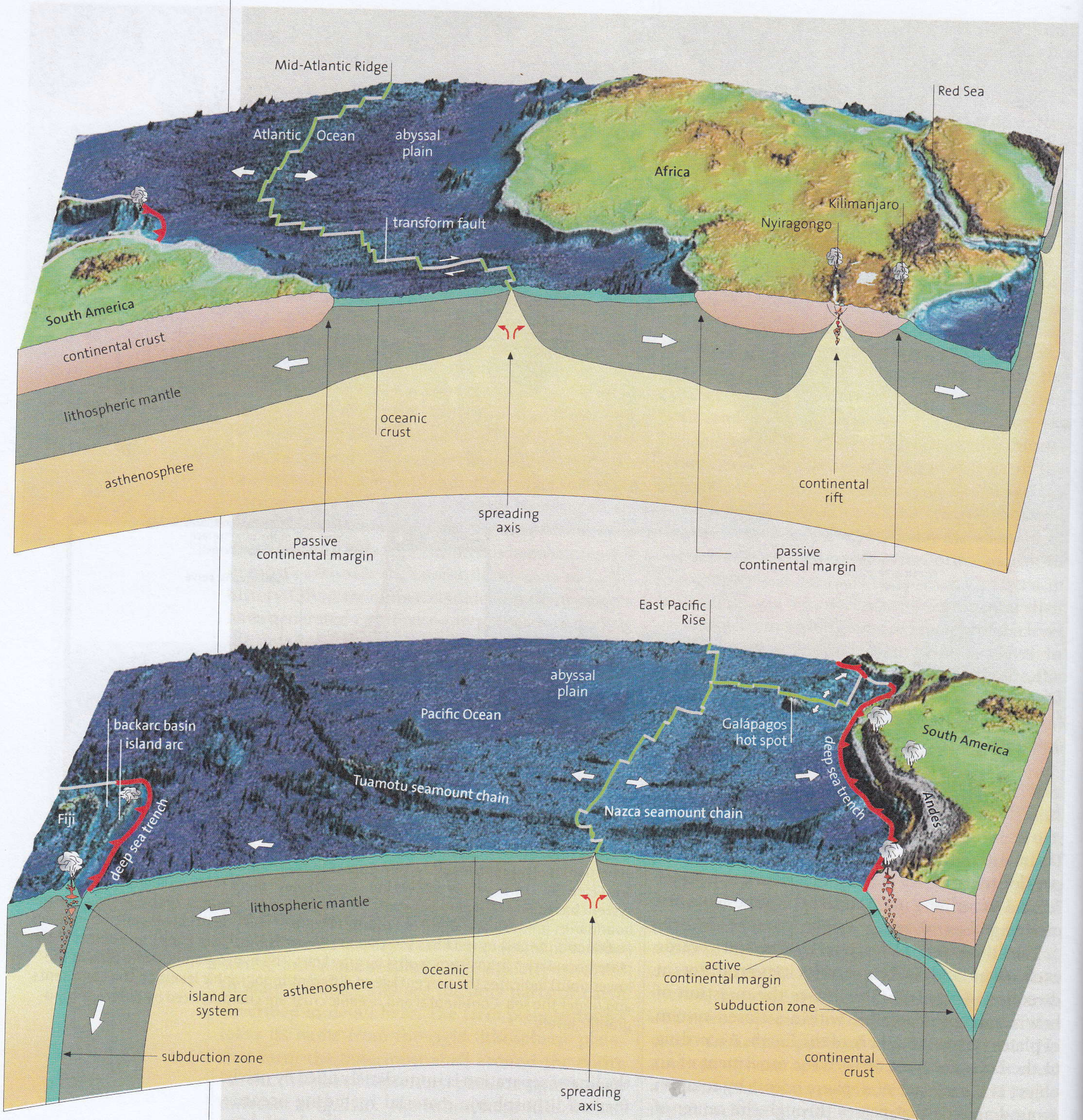
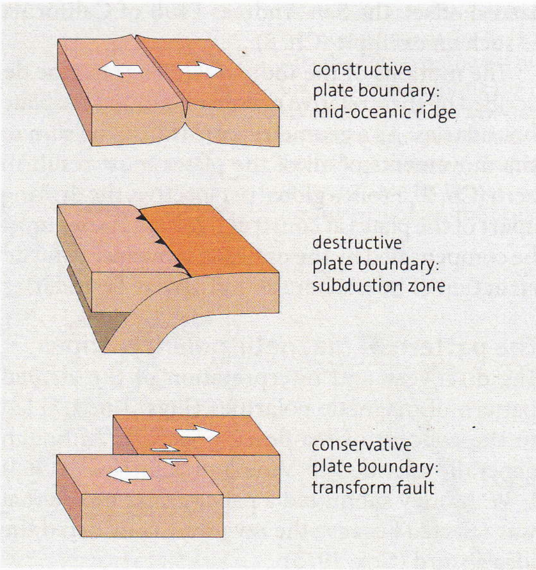


▲ Fig. 1.3 Block diagrams of the outer shells of the Earth in the Atlantic and the Pacific region. Shown are the three types of plate boundaries, passive and active continental margins, island arcs, volcanic chains fed by hot-spot volcanism, and a graben system (strong vertical exaggeration). The plates consist of crust and lithospheric mantle. Relief data are from etopo30 (land surface) and gtopo2 data by Smith and Sandwell (1997), and etopo1 data by Amante and Eakins (2009).

warm, less dense lithosphere that underlies it (older oceanic lithosphere is more dense and expresses the low topography of the abyssal ocean plains). The ocean floor spreads from constructive plate boundaries, hence "sea-floor spreading".

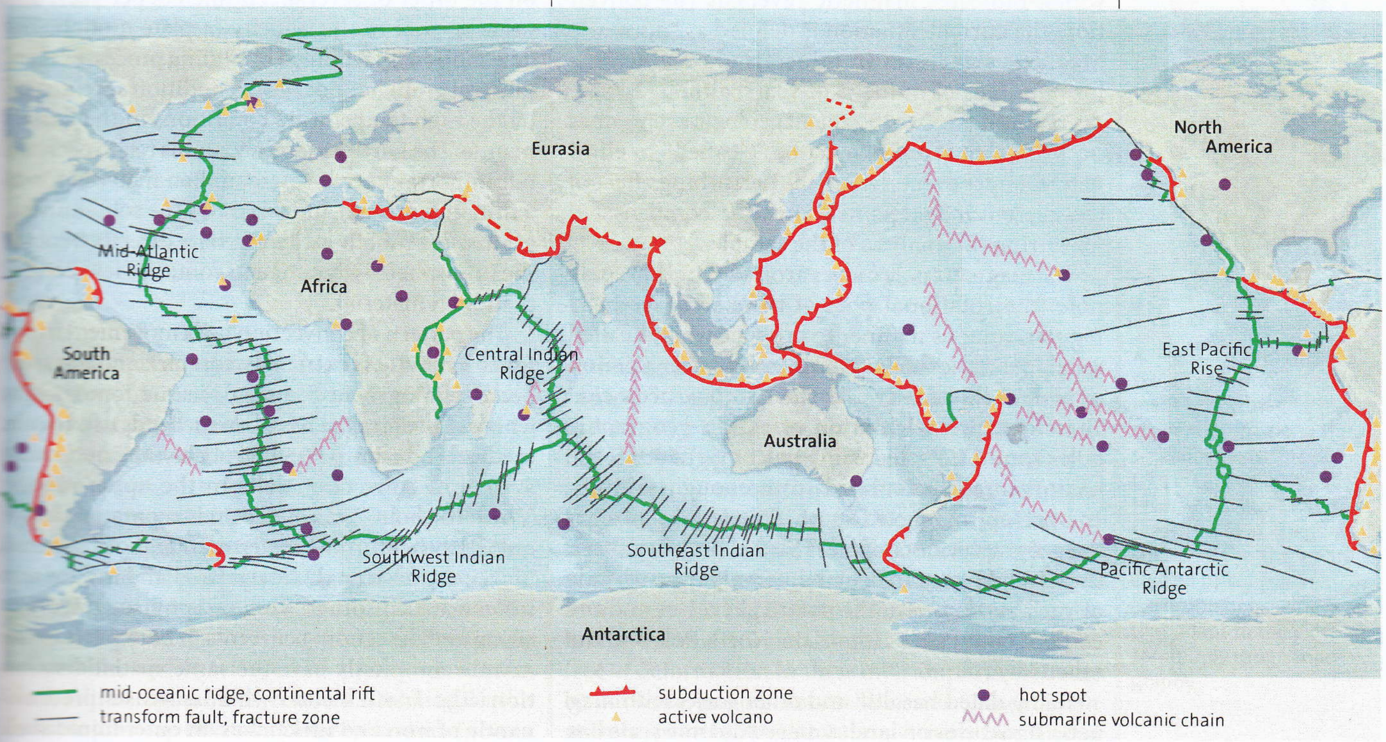
Destructive plate boundaries are characterized by converging plates. Where two plates move towards each other, the denser plate is bent and pulled beneath the less dense plate, eventually plunging downward at an angle into the depths of the sub-lithospheric mantle. Such areas are called *subduction zones*. Eventually the subducted plates become recylced into the mantle and thus destroyed. At convergent boundaries as they are commonly termed, only dense, oceanic lithosphere can be diverted into the sub-lithospheric mantle in large quantities; thicker, less dense continental lithosphere can not subduct very deeply – this explains why old continental crust, billions of years old exists today while no ocean crust greater than ca. 180 Ma (*Mega anni* – millions of years ago) is present – older ocean crust has all been recycled. Surface expression of subduction zones is manifested in the deep-sea trenches, common features around the Pacific Ocean (**Figs. 1.3, 1.5**).

Conservative plate boundaries occur where two plates slip past each and little or no crust is created or destroyed in the process. These boundaries are also characterized by strike-slip or *transform faults* and are commonly called transform margins. Transform faults are rare in the purely continental realm. Mid-ocean ridges on the other hand are cut by numerous, mostly relatively short transform faults (**Figs. 1.3, 1.5**). Such faults connect two segments of the ridge that are apparently shifted relative to each other. The prolongation of oceanic transform faults contain *fracture zones* with little tectonic activity that in many cases can be traced for long distance on the ocean floor. Where oceanic transform faults intercept continental crust at oblique angles, the faults can penetrate deeply into the adjacent plate and create large distances of



◄ **Fig. 1.4** Block diagrams showing the three types of plate boundaries.

▼ **Fig. 1.5** First-order tectonic elements of Earth. Each of the present plates is readily discernable.



lateral offset; the San Andreas Fault of California is such an example (Ch. 8).

The motions of the individual plates can be described by their relative movements along the plate boundaries. As a geometric constraint, the sum of the movements of all of the plates must result in zero (Ch. 2). From a global perspective, the drifting apart of the plates at constructive boundaries must be compensated by the opposite movement and destruction of lithosphere at destructive boundaries.

The pattern of magnetic polarity stripes

The discovery and interpretation of the striped pattern of magnetic polarities (**Figs. 1.6, 1.7**) led to the concept of sea-floor spreading. Although generally ascribed to Vine and Matthews (1963), L. W. Morley submitted a paper a year before that was rejected because the reviewers considered the idea absurd (Cox, 1973).

Minerals and the rocks in which they are contained acquire a magnetic signature as a given mineral cools below a certain temperature, its Curie temperature. Below the Curie temperature, named after the physicist Pierre Curie, a given mineral acquires the magnetic signature of the Earth's magnetic field that was present at that time. As an example, magnetite has a Curie temperature of 580 °C. Three signatures of magnetism are generally infused into magnetic minerals: inclination, which reflects latitude; declination, which reflects direction to the poles; and normal or reversed polarity, which indicates magnetic reversals (by convention, the current situation is defined as "normal"). Magnetic signatures in minerals are maintained for hundreds of millions of years, although some overprinting from subsequent geologic events does occur so that samples must be "cleaned" to eliminate younger events. Also, the perturbing effect of the current magnetic field must be compensated for during the analysis of the sample.

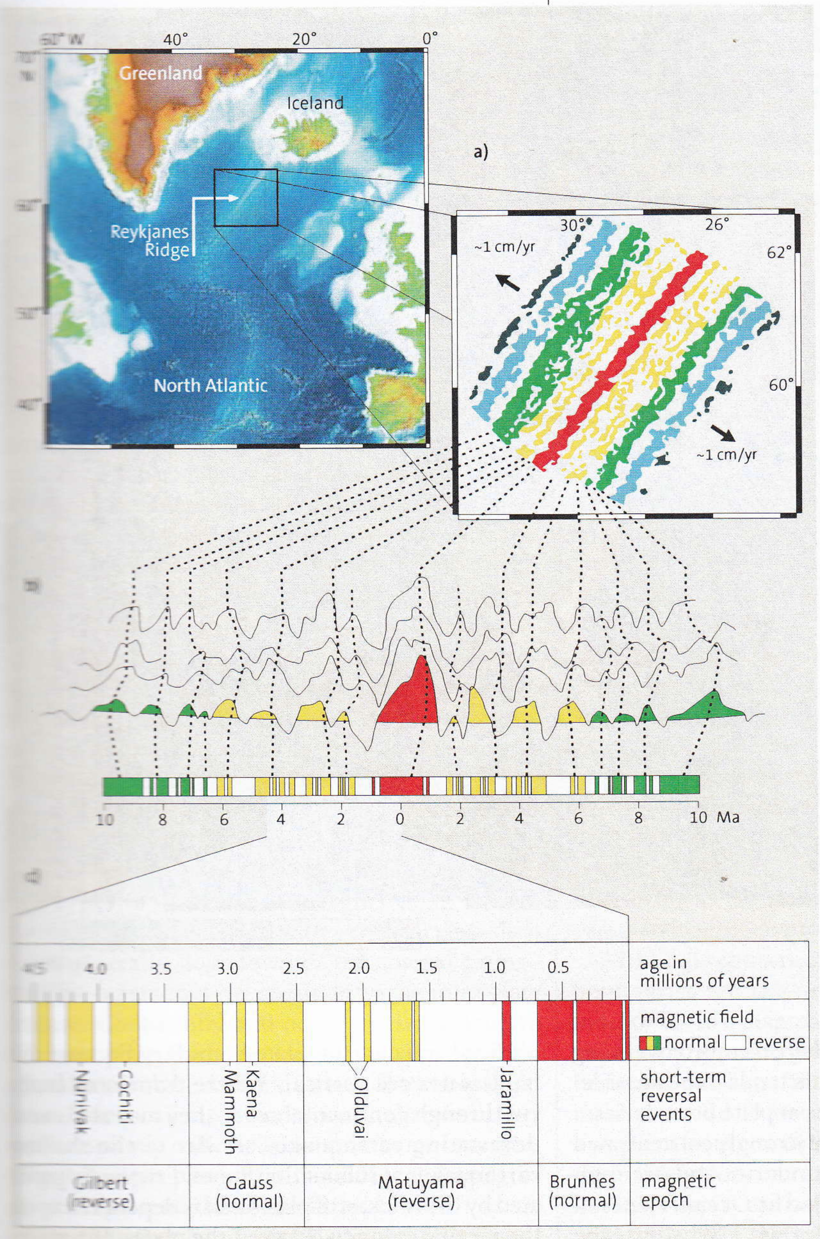
The magnetic pole moves around the geographic pole (the rotational pole of the Earth) in an irregular, sinuous manner to produce what is called secular variation. However, averaged over a period of several thousand years the two poles coincide. Therefore, the orientation of earlier geographic poles can be detected using paleomagnetics if the mean value is calculated from enough samples. The present magnetic South Pole is located near the geographic north pole. This has not alwas been the case. At very irregular intervals over periods of variable duration, the polarity reverses and the earlier South Pole becomes the North Pole and the other way round.

Using dated basaltic and other rocks with magnetic signatures on land, a magnetic time scale has been defined that displays the periods and epochs with normal (like today) and reversed magnetization. These patterns of magnetization are found parallel and symmetrically aligned to the oceanic ridges (**Fig. 1.6**). Based on the characteristic patterns of normal and reversed magnetization, the stripes can be dated by comparing them with known sequences. This is very strong proof for sea-floor spreading because the method shows that variable magnetic stripes of oceanic crust are formed parallel to the ridges and that they become older with increasing distance to the ridge (**Fig. 1.7**). It was the discovery of this symmetric pattern parallel to the ridges that proved in the early 1960's the concept of sea-floor spreading and associated drifting of continents, two of the most basic tenants of plate tectonics. Magnetic reversals in oceanic rocks only yield data back to approximately 180 Ma, the Early Jurassic (see **Fig. 2.12**) – all older oceanic crust has been subducted. A paramount reason for this fact is that older ocean crust is colder and more dense, and therefore subducts more readily; for example, if 20 Ma ocean crust and 150 Ma ocean crust collide, the older will be subducted (Ch. 4).

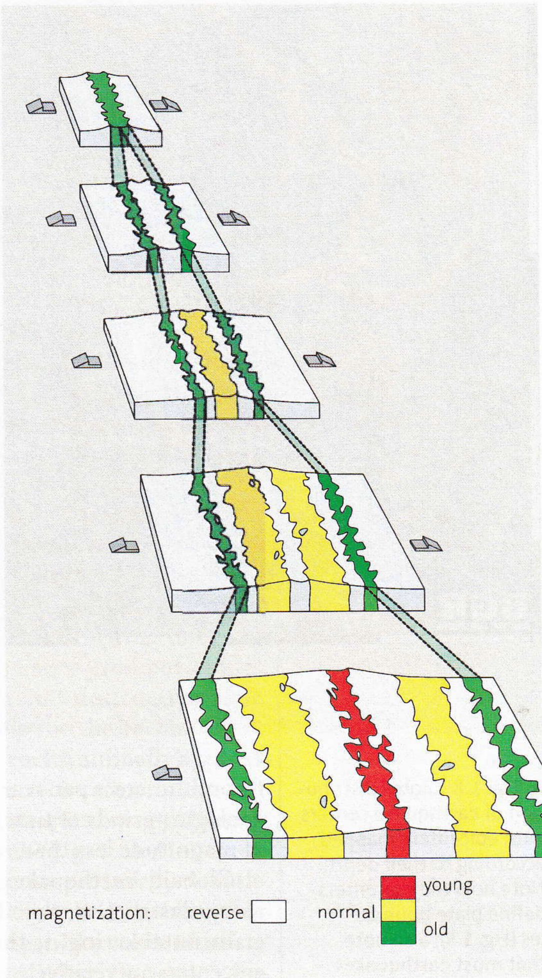
Plate motions and earthquake zones

Convectional currents in the sub-lithospheric mantle are interrelated to overlying plate motions. Based on the propagation behavior of earthquake waves, it is known that the Earth's mantle is primarily in a solid state. Nevertheless, it is able to flow on the order of several centimeters per year; this value is close to the velocity of plate motion. The flow motion is facilitated by gliding processes along mineral grain boundaries, a condition accentuated by the high temperature conditions in the Earth's mantle. The Earth's mantle contains relatively small but important areas where molten material forms a thin film around and separates the solid mineral grains. The mobile asthenosphere directly beneath the lithosphere is assumed to contain a few percent of molten material.

The pattern of convection cells movement in the Earth's mantle is extremely complex and has even been "photographed" using seismic tomography, which is a technique based on methods used in the medical industry (Ch. 2). Probably, the outermost system of convection cells in the upper mantle (down to about 700 km depth) is separated from a second system in the lower mantle; both systems, however, are strongly interrelated and induce and influence each other. Tomography suggests that rising and descending currents in both parts of the mantle commonly have the same spatial distribution. The Earth's core, which consists predominantly of iron and nickel, has an outer liquid shell



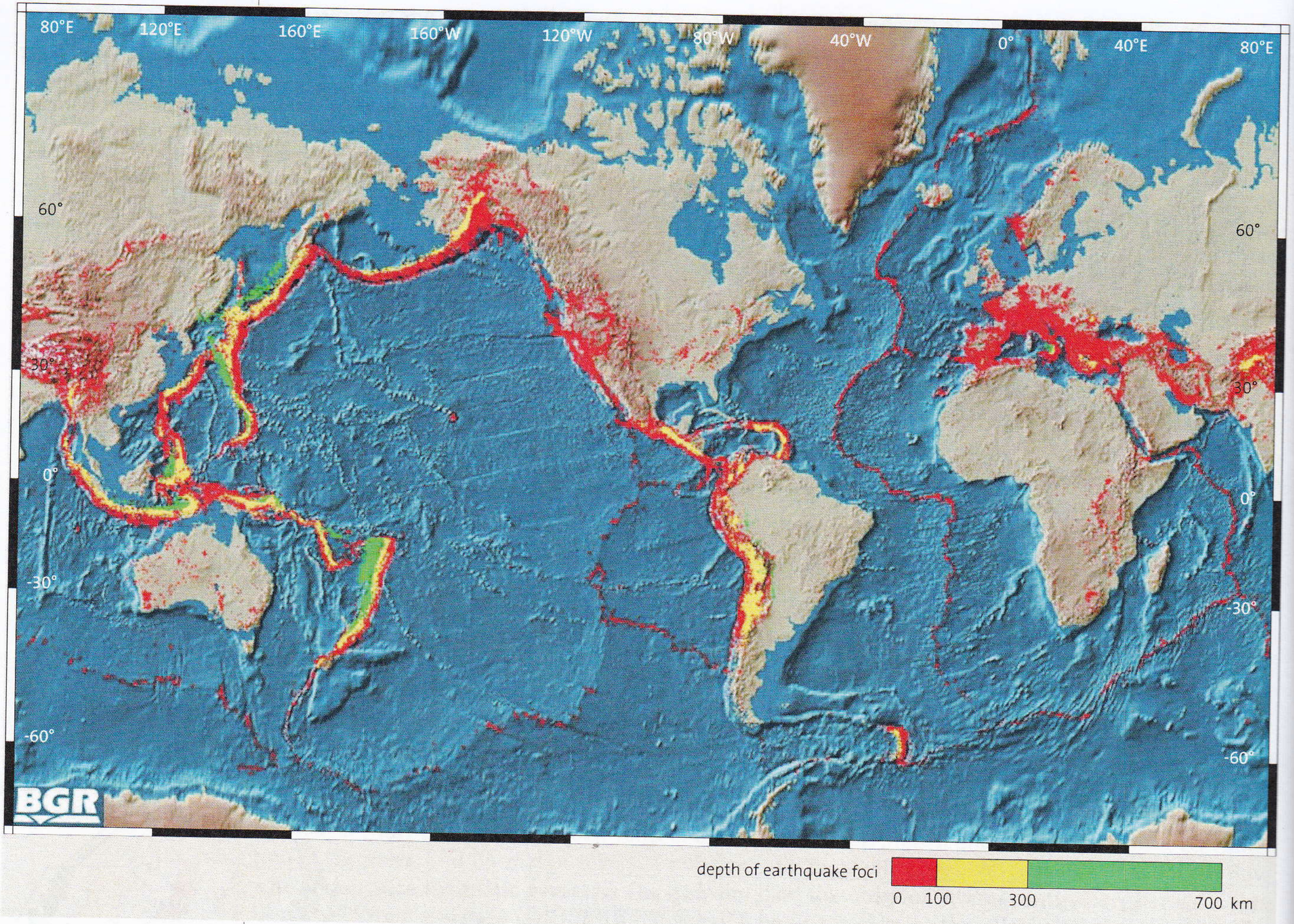
◀ Fig. 1.6 a) Stripe pattern of magnetic polarities on the ocean floor at the Reykjanes Ridge, part of the Mid-Atlantic Ridge southwest of Iceland (Heirtzler et al., 1966). b) Curves representing the magnetic field strength measured along the track of ships crossing the ridge. Normal (in colors) and reverse magnetization can be obtained from these curves. c) Graph showing detailed magnetic stripe pattern for the last 4.5 Ma. By comparison with measured profiles, the ocean floor can be dated.



▲ Fig. 1.7 Simplified sketch showing development of the magnetic stripe pattern along the spreading axis. The pattern is caused by repeated reversals of the Earth's magnetic field. Irregularities of the stripes are caused by submarine extrusion of basaltic lavas that adapt to the existing, commonly rough topography.

that surrounds an inner solid sphere. The relations of the core to convection cell currents and plate motions are still under investigation. However, it can be assumed that interactions and material transfer occur.

Relative movements of the plates along the plate boundaries induce earthquakes. The gliding process between the plates produces great but variable stress. Stress is induced along slip planes within rocks, which to a certain degree are elastically deformable, and is released in a jerky movement when a limiting value is reached. Looking at the distribution map of earthquake epicenters (points at the Earth's surface directly above the earthquake center) it impressively shows that earthquakes are mainly restricted to narrow zones around the globe (**Fig. 1.8**), the present plate boundaries. Distribution of earthquakes varies at different types of plate boundaries. Deep earthquake centers occur only along subduction zones, whereas shallow earthquakes occur at all plate boundaries. Moreover, distributed earthquake centers can be found elsewhere indicating that the plates are not free of deformation in their interior parts that may be cut by large fault zones. The rates of movement at intra-plate fault zones are generally less than a



▲ Fig. 1.8 Global distribution of earthquake centers with epicenters mapped according to their depth. Note how the epicenters define plate boundaries (Fig. 1.5); also note that most earthquakes occur within 100 km of the surface except along subduction zones where they deepen under the upper plate (produced with the kind support of Mrs. Agneta Schick, Federal Institute of Geosciences and Resources, BGR, Hannover, Germany).

few millimeters per year when averaged over long geologic periods of time; this tends to be an order of magnitude less than rates at plate boundaries.

Globally, earthquakes are strongly concentrated along destructive plate boundaries and are especially notable ringing the Pacific Ocean. Zones of epicenters are relatively wide (Fig. 1.8) because the subducting plates that produce the earthquakes plunge obliquely into the mantle. Subduction zones can be traced downward, using earthquake foci (the location and depth of the earthquake) to depths of approximately 700 km. The map view of the plate boundary is located at the trench side of the earthquake belt where the centers are at a shallow depth; the boundary plunges downward at different angles. Earthquake foci at shallow depths, with epicenters mostly near the surface of the plate boundary, may have devastating consequences – these are the locations of the Earth's most destructive earthquakes.

Along transform faults the epicenters of earthquakes are much more concentrated near the surficial trace of the plate boundary because the fault zones are vertical. Where transform faults cut through continental crust, they may also cause devastating earthquakes similar to the shallow earthquakes at subduction zones. Friction is generated by the thick, stiff plates and is dependent upon the motion velocity between the plates. Examples of continental transform faults include the San Andreas Fault in California and the North Anatolian Fault in Asia Minor.

Earthquake activity is much less at mid-ocean ridges. Uprising currents transport molten rock material to the Earth's surface and the stiff shell that accumulates and releases stresses is quite thin. Hot and recently solidified rock material is more likely to deform plastically. Therefore, only small, shallow earthquakes occur. Nevertheless, the constructive plate boundaries are also clearly visible on the earthquake map (Fig. 1.8).

Young mountain ranges like the Alps-Himalaya belt or the Andes-Cordillera belt, which are still tectonically active, are also characterized by frequent earthquakes. Because of the diffuse collision of large continental masses, wide zones of deformation with numerous slip planes develop. Therefore, exceptionally wide belts of shallow earthquakes occur in these zones (**Fig. 1.8**). Occasional deeper earthquakes testify to the preceding subduction activity.

Two kinds of continental margins

Nearly all of the presently existing plates contain areas with both continental and oceanic crust. A good example includes the large plates on either side of the Mid-Atlantic Ridge, one of Earth's most prominent plate boundaries. This plate boundary separates the two American plates from the Eurasian and African plates, all of which have large amounts of both continental and oceanic crust (**Figs. 1.2, 1.5**). The Indo-Australian, the Antarctic and many smaller plates also contain both crustal types. In contrast, the huge Pacific Plate which extends westward from the East-Pacific Rise to the eastern Asian island arc systems, contains only very small amounts of continental crust, mostly in California and New Zealand. The Philippines, Cocos, and Nacza plates – smaller plates that surround the Pacific Plate – only contain oceanic crust.

The fact that most plates contain both crustal types means that some boundaries between oceans and continents occur within a given plate; hence, two types of continental margins exist. Where continental crust merges with oceanic crust, shelf areas generally slope towards the abyssal plains – thus the ocean-continent boundary is an intra-plate feature. Continental and oceanic crust belong to the same plate. Such continental margins are widespread around the Atlantic Ocean. Here only slight (mostly vertical) movements occur; therefore, they are commonly called *passive continental margins* (**Fig. 1.3** upper part). Passive continental margins do not represent plate boundaries.

On the other hand, *active continental margins* are those margins where a plate boundary exists between continent and ocean. Two types occur – subduction margins and transform margins. At subduction margins, a part of a plate with oceanic crust is being subducted beneath the continental crust. At transform margins, the oceanic plate slides laterally along the continental margin. A deep sea trench forms along subduction zone plate boundaries. This type of continental margin is today prominent along the Andes (**Fig. 1.3** lower part) and along numerous subduction zones around the Pacific Ocean that are characterized by island arc systems. The margin of the upper plate in these cases is characterized by chains of volcanic arcs, built either on continental crust or on continental pieces that were separated from the neighboring continent.

Magmatism and plate tectonics

Magmatic belts as well as earthquake activity are closely related to plate boundaries. The average yearly production of magmatic (volcanic and plutonic) rocks formed at destructive plate margins is slightly less than 10 km^3 (Schmincke, 2004). The melting that produces magmatism is caused by complex interrelations between the asthenosphere and the subducting plates plunging into it. These melts, which are marked by specific chemical characteristics, intrude into the upper plate and feed volcanic chains above subduction zones (**Figs. 1.3, 1.5**) to produce subduction related magmatism. Modern examples include the eastern Asian island arcs (island arc magmatism) and the Andes (magmatism at an active continental margin).

Mid-ocean ridges are the location of major production of basic magmatites, namely basalts and gabbros. High temperature and pressure release beneath the ridges combine to generate partial melting of up to ~20 % the rocks of the mantle (peridotite). Oceanic crust develops from these melts and annually more than 20 km^3 of new crust is formed (Schmincke, 2004). Therefore, mid-ocean ridges generate more than twice the amount of melts than are generated above subduction zones. At transform faults significant melting does not occur so magmatic processes are unimportant.

Although constructive and destructive plate boundaries are responsible for the formation of most of Earth's magmatic rocks, annually approximately 4 km^3 of magmatic rocks are produced in intraplate settings. This intraplate magmatism is mostly related to hot spots (**Fig. 1.5**). Hot spots are point-sources of magma caused by mantle diapirs and occur on either the continents or oceans. Diapirs are hot, finger-like zones of rising material within the mantle. When they reach the upper asthenosphere beneath the plates, melting is induced that creates volcanic eruptions and doming of the surface over long time periods. Hot spots are less commonly superimposed on constructive plate boundaries.

Modern continental hot spots include the Yellowstone volcanic field in North America, the French Central Massif and the volcanic Eifel Mountains in Europe, and the Tibesti Mountains and the Ahaggar (Hoggar) in North Africa (**Fig. 1.5**). Modern oceanic hotspots include the active part of the Hawaiian Archipelago and the Canary Islands; Iceland is an example of a hotspot superimposed on a mid-ocean ridge. As plates drift over hot spots, long volcanic chains develop with the hot spot located at the active end; Hawaii is a good example of this (Fig. 1.5). On continents, hot spots are commonly related to graben structures characterizd by extensive, deep fault systems that cut through the entire thickness of continents; the best known example are the volcanoes of the East African graben system. Graben structures are characterized by crustal extension and bordered by faults; such areas cause thinning of the lithosphere and provide the opportunity for magma to rise along fault zones. If extension continues, new ocean can be formed at these structures. An example for such a newly developing ocean is the northern part of the East African graben system (Afar) and the Red Sea (Ch. 3). The so-called Afar triangle is also characterized by a hot spot. Graben structures may be transferred into constructive plate boundaries where the hot spot commonly plays an important role.

What drives the plates and what slows them down?

The growing plate boundaries at mid-ocean ridges always forms new oceanic lithosphere because the basaltic/gabbroic oceanic crust is a product of partial melting directly from the mantle; on the other hand, continental crust forms by much more complicated melting and recycling processes above subduction zones (Ch. 7). Only oceanic lithosphere can be completely reintegrated into the mantle at subduction zones whereas subducted continental crust experiences strong buoyancy and accretion to the overlying plate. Plate movements are thus mainly controlled by the formation of oceanic lithosphere at the oceanic ridges, and its subduction and reintegration into the Earth's mantle at destructive margins. Oceanic lithosphere, therefore, forms the conveyor belt of plate tectonics whereas the continental blocks go along for the ride.

In fact, the driving forces for the plate movement are to be found under the mid-ocean ridges and in the subduction zones – at the plate boundaries. Plate motion is thus orchestrated by rising magma at the mid-ocean ridges and sinking dense lithosphere at the subduction zones (Bott, 1982). These processes are called "ridge push" and "slab pull". Ridge push is caused by the upward movement of hot and relatively light rock melts at the mid-ocean ridges where, in the area of newly forming lithosphere, the vertical movement is transferred into a horizontal vector that pushes the plates apart. Slab pull arises because of the higher density of cooled lithosphere with respect to the mantle underneath. Of the two driving forces this is the more important one. Mineral changes to denser species that, because of lower temperatures in the descending plate, occur at shallower depth as compared to that of the surrounding mantle, intensify this process. An earlier idea that suggested that the carrying of middle parts of plates on top of the horizontal currents of the asthenosphere may, however, not be important in plate motion and, in contrast, actually hinder the process in certain regions.

Ridge push and slab pull act in accord with the state of stress in the plate interiors. They produce compression near the mid-ocean ridge and extension near the deep sea trenches. If plates were actually carried by currents, the state of stress would be the other way round. Also, from a thermodynamic view, it is logical that rising hot and descending cold material supply the driving forces. These considerations are supported by the following observations. The velocities of plate movements are independent of the size of the plates, and plates with subduction borders move faster than those without subduction borders; this emphasizes the importance of the slab pull as the driving force. Plates with a large percentage of thick continental crust move more slowly, an observation that suggests that dragging at the bottom of the plates (like the keel of a boat on sand) negatively influences the movement (Kearey and Vine, 1990).

Collision and mountain building

Subduction zones tend to form in locations with mature (older), cool, and thus denser lithosphere. These conditions exist at the edges of large oceanic basins like the present Pacific Ocean. If an oceanic basin is not bounded by subduction zones, it will widen as is the case with the present Atlantic Ocean. Spreading rates at middle oceanic ridges range from a low of 1 cm/yr to a high of 15 cm/yr (Fig. 1.2). The rate of subduction (at present up to 9 cm/yr – in the past probably faster) in an ocean basin can exceed the rate of new crust formation at a ridge, especially if opposite sides of the basin both contain subduction zones. In such a case, the oceanic basin shrinks and adjacent continental blocks move closer together. Continuing convergence finally leads to the collision of the continental blocks and the passive continental margin of the subducting (lower) plate is dragged beneath the active boundary of the upper plate. The low density of the subducted part of the continent prohibits extensive subduction and it cannot be dragged down to great depths. Rather, it buoys up on the surrounding denser mantle material and rises, following the principle of isostasy.

Buoyancy and strong frictional forces following collision of two continental blocks eventually brings the subduction of continental crust to a standstill. During this process, complex tectonic structures such as folds and nappes develop and