2. Plate Tectonics

In this chapter we repeat basic aspects of the theory of plate tectonics. In the first part of the chapter a summary of the history of the plate tectonic model is presented and some basic principles how to describe it are discussed. Some arguments that suggest why plate tectonics may *not* work are also presented. In the second part of this chapter we discuss the layered structure of the earth and the geographic distribution of lithospheric plates. We will also use this chapter to introduce the terminology that is used in the remainder of the book. As such, the chapter is meant as a basis for all following chapters.

2.1 Historical Development

Observations pertaining to the theory of plate tectonics are at least 500 years old. In the late 16th century, Sir Francis Bacon observed that the coast lines of the American and African continents have a matching shape. At Darwin's times the connections between the two continents were already well-established. However, it was only Alfred Wegener who presented the first synthesis explaining these similarities with a theory of plate motion. In part, his synthesis was based on his own observations on the climatological connections between the two continents. While Wegener's synthesis was ultimately proven wrong by the first detailed bathymetric surveys of the oceans in the middle of the past century, his publications (and those of others around his time, e. g. Taylor 1910) are still viewed as the basic foundation of plate tectonic theory (Wegener 1912a,b; 1915). At the time of Wegener, the significance of mid-oceanic ridges and subduction zones were still unknown. However, the deepest point in the world (the Mariana Trench with 11.5 km below sea level) and the eastern Pacific rise had already been discovered by the research vessel H.M.S. Challenger around 1875. Also, mantle convection was already established as the driving force for plate motion (Holmes 1929; Griggs 1939). Nevertheless, Wegener had no model for the processes on the ocean floors and thought of plate motion as some "plough-like" motion of the continents through the oceans. From the time of this theory, only the names of the ancient super continents *Gondwana* and *Laurasia* (du Toit 1937) are still being used.

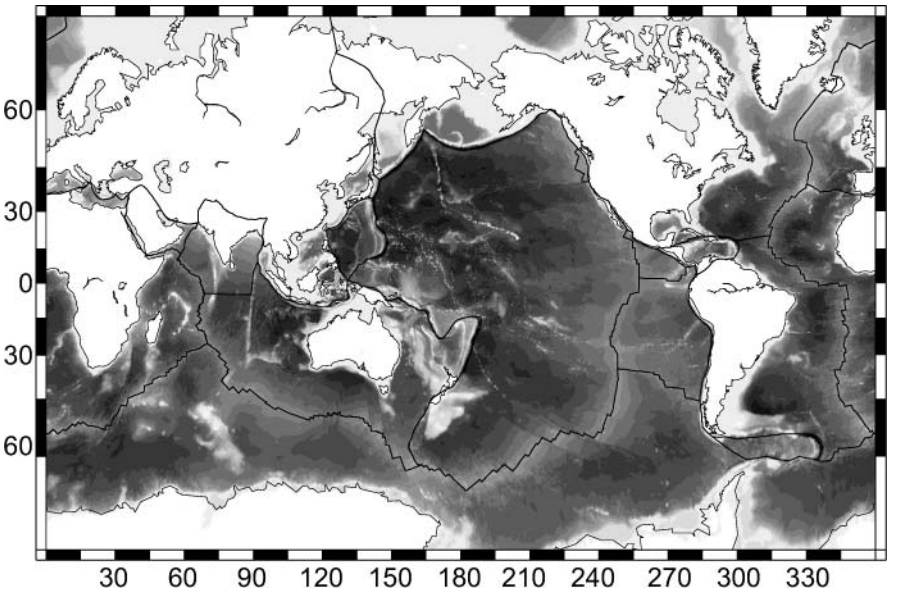


Figure 2.1. Topographic map of the ocean floor. Most regions where the water depth is less than 200 m (nearly white regions) are made of continental lithosphere. Note that in some regions substantial portions of continental lithosphere are actually below sea level, for example around New Zealand, between North-America and Siberia, in the Mediterranean or east of southern South America

The real break through of plate tectonics came not until the mid-twentieth century when the first detailed bathymetric surveys of the Atlantic were performed around 1950 to 1965 (Heezen 1962; Menard 1964). During these surveys, the gigantic mountain ranges and valleys of the ocean floors were discovered that are now known as mid-oceanic ridges and subduction zones. Earlier on, however, these valleys and ridges were interpreted by all kinds of theories, for example the *expanding earth theory* (Carey 1976; King 1983). However, since the mid-sixties it is well-established that mid-oceanic ridges are areas of lithosphere production, while subduction zones are areas where lithosphere is being consumed (*destructive plate margins*). The chains of volcanoes that produce new oceanic lithosphere along the mid-oceanic ridges had already been predicted by some authors, but they were only discovered by the submarine research vessel Alvin (s. Edmond and Damm 1983). Subduction zones and mid oceanic ridges are in volumetric balance so that there is no need for an expanding earth theory (Hess 1961; Vine and Mathews 1963). Today our knowledge of mid-oceanic ridges and subduction zones is well-established as the basic foundation of plate tectonic theory (sect. 2.4.3) (e. g. Morgan 1968).

Modern Understanding of Plate Tectonics. Our modern understanding of plate tectonic theory is largely based on observations from the ocean floors (Le Pichon et al. 1976). The total length of mid-oceanic ridges is of the order

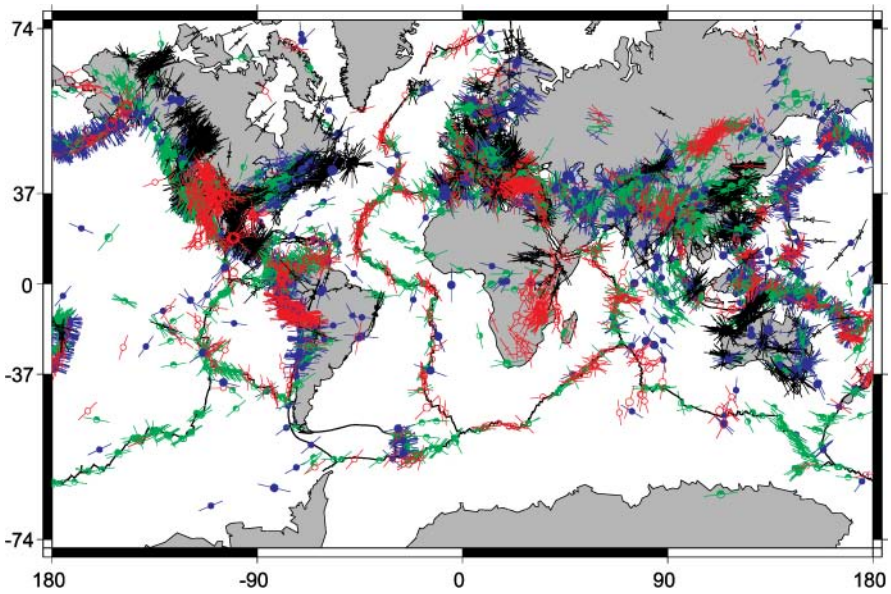


Figure 2.2. The intraplate stress field of the world. The different symbols indicate different methods of determination including earthquake focal mechanisms, borehole breakouts and geological indicators. The different shadings indicate different deformation regimes: darkest are thrust faults, medium gray are normal faults and light shading are strike slip faults. For detailed resolution of this figure see the original CASMO facility on the world stress map home page, which was used to create this map (Müller et al. 2000)

of 60 000 km. Most of them are located on the ocean floors (s. sect. 2.4.3). The average rifting rate at these ridges is about 4 cm per year, (Table 2.3) which implies that the total production rate of new surface on earth is about is about 2 $km^2\ y^{-1}$. This surface production is balanced by surface wasting at the subduction zones. The total surface of the oceans is about $A_0 = 3.1 \cdot 10^8\ km^2$ which means that all oceanic lithosphere is being renewed about every 155 my. This is a geologically short time span and means that oceanic lithosphere is one of the younger features of this globe. It is therefore ironic that plate tectonic theory – which is now mostly being applied to our very detailed observations in the continents – has its origin *not* on the continents, but at the deepest points of the ocean floors. Cox (1972) has summarized the revolution of plate tectonic theory. He considers the plate tectonic theory to be based on four independent data sets:

- The topographic maps of the ocean floors (Fig. 2.1).
- The magnetic maps of the ocean floors.
- The age dating of the magnetic maps (Fig. 2.4).
- Detailed maps of the epicenters of global earthquakes (Fig. 2.5).

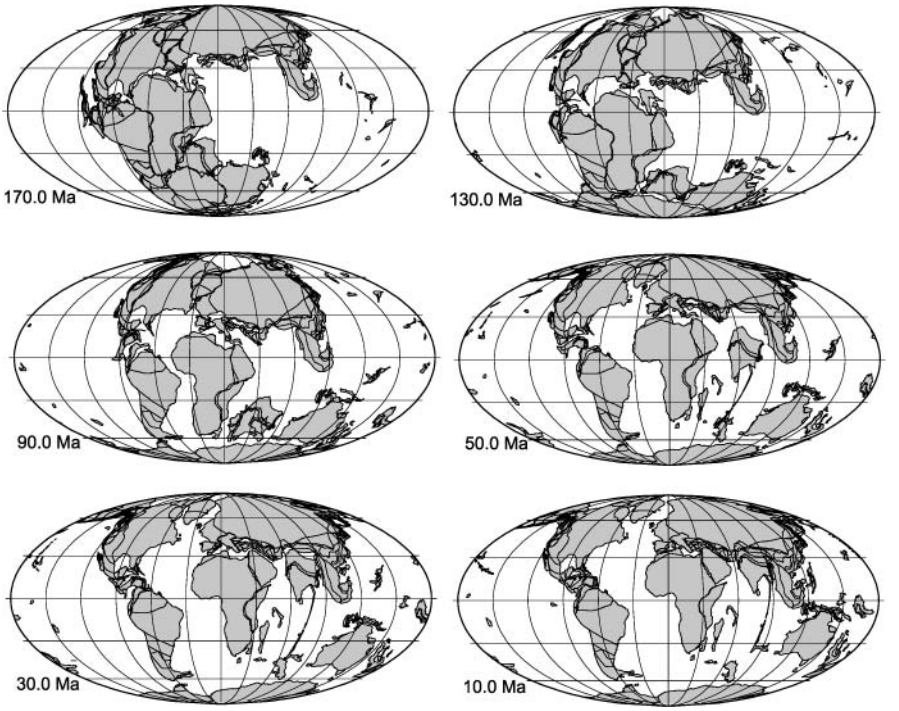


Figure 2.3. Plate reconstructions for the last 170 my. Produced with the ODSN home page (s. p. 455). Maps are in Mollweide projection reconstructed relative to the magnetic reference frame

With the advance of GPS measurements of active plate motions over the last 10 years (e.g. Kreemer et al. 2003), it may be said that a 5th data set must be added to the list:

– Direct measurement of plate motions.

Since the fundamental break through of the sixties, plate tectonic theory has made dramatic and rapid advances. It was soon discovered that many observations can be explained by astoundingly simple physical models, all within the plate tectonic concept. For example, the startlingly simple quadratic relationship between space- and time scales of diffusion processes (sect. 3.1.4) has now been used to explain an enormous variety of processes including the water depth of the oceans (sect. 4.4.1), the duration of metamorphic events, the magnitude of contact metamorphic aureoles (sect. 3.6.2) or the shape of chemical zoning profiles in minerals (sect. 7.2.2).

Such amazing success of simple physical models has lead in the past 30 years to an unparalleled development of plate tectonic theory. Much of this development has been characterized by the application of simple analytical

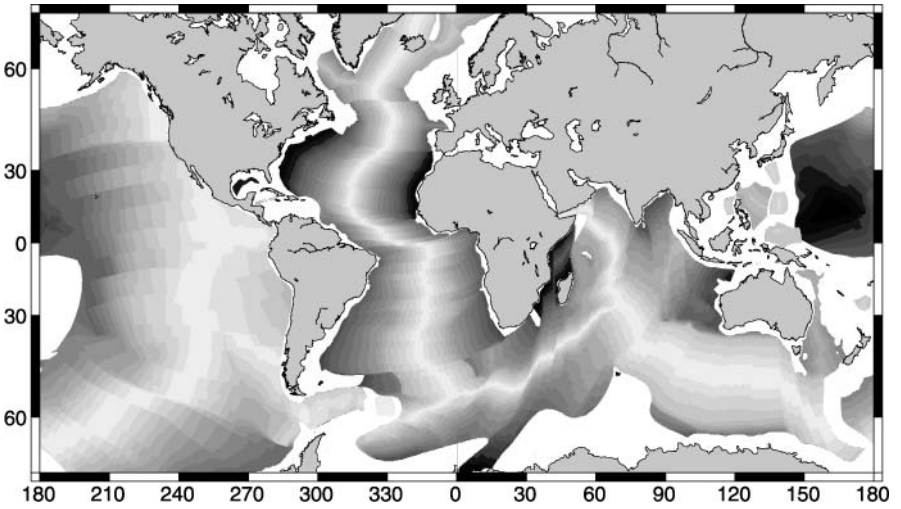


Figure 2.4. The age of the ocean floor (Müller et al. 1997). Shading intervals are every 10 my from 0 (white) to 160 my (black). Ocean floor older than 160 my is black. Areas with no data ar white. These regions are both on continental and on oceanic lithosphere. Landmasses are grey. Oldest known parts of oceanic lithosphere are around 180 my in the western Pacific just east of the Mariana Trench, between Madagascar and Africa and in the westernmost Atlantic just east of the US. Compare this map also with the topography of the ocean floor (Fig. 2.1) and note the similarities.

models to geological problems. Many of these models are introduced in this book.

Are We Sure that it Works? Despite the wide spread acceptance of the plate tectonic model, there are observations and ideas that indicate that it may not work after all. Most important among these observations are the occurrence of zircons with continental signatures found in oceanic islands, for example in Mauritius or Iceland (e.g. Schaltegger et al. 2002). It also has been argued that there is a discrepancy between the observed long narrow zones of axial volcanism at the mid ocean ridges and the geometry of upwelling mantle plumes (McKenzie 2001). As an alternative, it is being discussed if mid ocean ridges may reflect zones of asthenospheric downwelling and that the magnetic striping of the oceanic plates is the consequence of narrowing of formerly much more expanded zones. Within this view substantial parts of the oceanic plates may represent the ultimate end member of an extremely thinned continental margin. Even if arguments can be found that refute such models, some of these hypothesis remind us not to accept the plate tectonic model uncritically.

Nevertheless, there is no debate about the fact that plates move and that the surface of earth includes zones of divergence and convergence that separate zones of less internal strain which we often refer to as cratonic interiors

or plates (Fig. 4.2). More than calling the plate tectonic concept "wrong" it could be said that it becomes slowly outdated in view of our much better understanding of the processes that govern plate motion and deformation.

Recent Developments and Future. Although the global distribution of earthquakes shows that *oceanic* plates act according to the principles of plate tectonic theory, continental plates do not: Oceanic plates are large and flat and their margins are narrow compared to their size and seismicity is confined to the rims of these plates. *Continents* behave differently. They are characterized by deformation and seismicity which reaches wide into the continents themselves, their thickness varies dramatically and some seismicity even occurs in their centers. Their plate boundaries are diffuse. It has therefore been suggested to use the term "cheese tectonics" as a superceding term for "plate tectonics", because continents seem to deform rather like soft camembert than like rigid plates. Like many good jokes this one has a true core. It has been known since the turn of the century that rocks behave like fluids on geological time scales (s. summary by Gordon 1965 or England 1996). However, only in the past 30 years geologists have begun to actually use the theory of deformation of viscoelastic materials to describe the dynamic evolution of continents (England and McKenzie 1982; England and Jackson 1989). In part these new descriptions were triggered by studies like those of Goetze (1978) or Brace and Kohlstedt (1980) who provided us with the first simple models of lithospheric rheologies.

Because of such models it has now been possible to describe largest scale tectonic processes with simple dynamic models. Even coupled thermomechanical approaches can now be performed with startling simplicity (e.g. Sonder and England 1986) and have found their use even by non-geophysically oriented earth scientists.

The current development of plate tectonics is going more and more towards the use of numerical models and further away from simple analytical models. Digital data sets of global observations – for example the global digital elevation model SRTM3 or the global sea floor topography of Smith and Sandwell (1997) – make it now possible to tackle problems that can only be solved using large numbers of data. Such models can now be used to explain problems that go beyond single observations and pertain to the whole globe. The future will surely be characterized by an increasing use of large data sets and numerical models.

2.2 Working on a Spherical Surface

The earth is nearly a sphere (s. sect. 4.2) and many aspects of the geometry and the mechanics on a sphere are different from its Cartesian equivalent. In this section we discuss some aspects of spherical coordinates that may need to be considered when solving geodynamic problems on very large scales. We

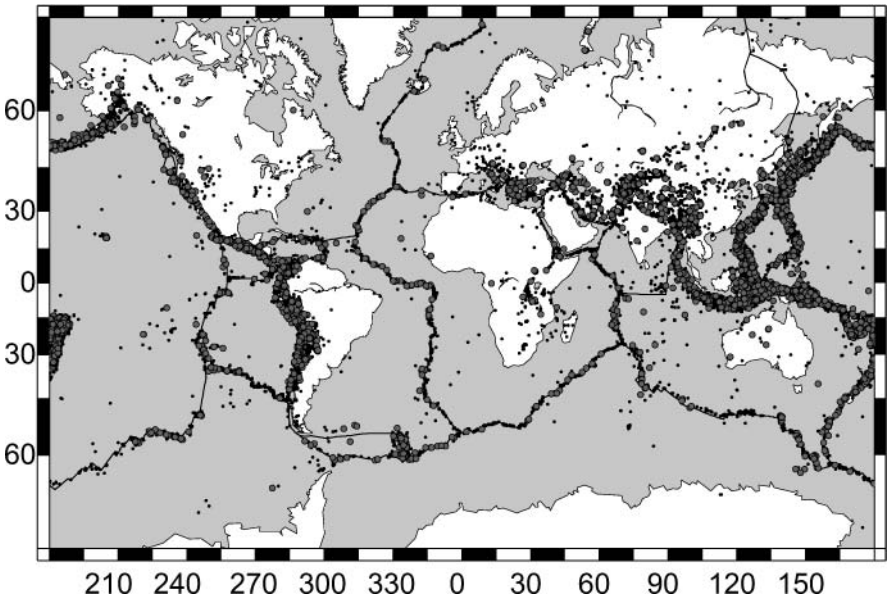


Figure 2.5. Global seismicity. All recorded earthquakes above magnitude 6 since 1973 are shown. Note that most earthquakes along oceanic plate boundaries (e.g. the Mid-Atlantic Ridge) occur very near the plate boundary, while earthquakes associated with plate boundaries in continental lithosphere cover much more diffuse areas (e.g. the India-Asia collision zone). See Fig. 2.20 and Fig.4.2 for more details about the plates and strain distribution

also discuss some ball park estimates that can be used to judge whether it is sufficient to neglect the spherical geometry.

2.2.1 ... or is the Earth Flat After All?

By far the largest number of geodynamic problems can be described assuming a flat earth. Fig. 2.6 shows that over a line two thousand kilometers in length, the curvature of the earth causes about 80 km deviation from a straight line. This is only about 5% of the extent of the feature. For geological features of some tens of kilometers extent, the deviation is only some tens of meters, which is about 0.1% of the extent of the feature. We can conclude that only for geodynamic problems that have a length scale of substantially more than thousand kilometers the curvature of earth must be considered.

A famous example for a problems that can not be described on a flat earth is the shape of long subduction zones. On a flat earth, the trace of a subduction zone should be linear, just like the linear trace of the curvature of a sheet of paper hanging off the edge of a table. In contrast, the trace of many deep ocean trenches is curved along the surface of earth. For example, the Aleutian, Kurile, Japan, Izu Bonin and Mariana trenches from northeast to southwest) along the Pacific ring of fire, the South Sandwich Trench east

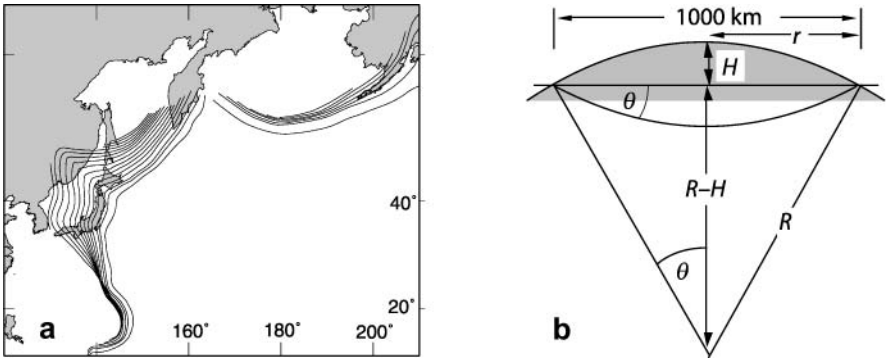


Figure 2.6. Illustration of the ping pong model. **a** Slab contours for the Aleutian slab and the Kuril-, Japan- and Mariana slabs. Depth contours are shown every 50 km from 0 km to 700 km (in the Japan Trench). Data from Gudmundsson and Sambridge (1998), (http://rses.anu.edu.au/seismology/projects/RUM). **b** The difference between a flat and a curved surface of the earth. The maximum deviation of a curved surface from a flat surface, H, is given from geometric relationships by $H = R - \sqrt{R^2 - r^2}$. For $r = 1\,000$ km as shown here and the value for R from Table D.3 the deviation is $H \approx 80$ km

of South Georgia or the Java (Sunda) Trench south of Sumatra and Java are all curved concave towards the subduction direction in plan view. This is often explained with the ping pong ball model. This model compares the curvature of trenches on the surface of the earth with the curvature of the indentation edge on a dented ping pong ball. If the indented part of a ping pong ball is not deformed in itself, then this edge forms a small-circle on the surface of the ball. Exactly this is observed in subduction zones. In fact, the model can be used to predict the subduction angle (2θ), which should be given (according to Fig. 2.6) by:

$$\sin(\theta) = \frac{r}{R} \tag{2.1}$$

where r is the small circle radius and R is the radius of earth. Most of the small circle radii of subduction zones on earth correspond well with the subduction angle predicted by eq. 2.1 (s. Isacks and Barazangi 1977). However, the subduction angles may also depend on a large number of other parameters, for example whether subduction occurs *in* or *against* the direction of convection in the asthenosphere (Doglioni 1993). In fact, it has been suggested that the earths rotation causes a westward drag between lithosphere and asthenosphere which also influences the direction and steepness of subduction zones (Doglioni et al. 1999). Other plate scale examples for the influence of the curvature on plate motions are the transform faults in the oceanic lithosphere (Fig. 2.9).

Geometrical problems can be described on a flat earth using the familiar Cartesian coordinate system. In this book we use x and y for the coordinates

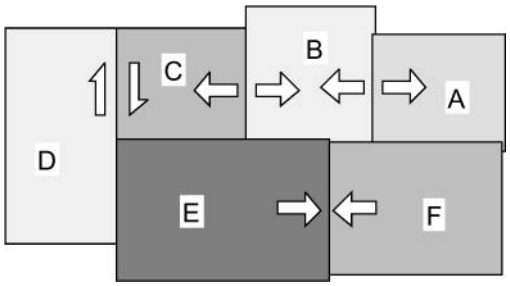


Figure 2.7. Six plates on a flat surface. The relative motions of some plates are shown by the arrows. However, the relative motions of plates C and E, D and E, B and E, B and F as well as that between A and F are completely unconstrained by the shown relative motions

oriented parallel to the surface of the earth and z for the vertical direction. In section 2.3 we discuss the relationship between those coordinates and those of other coordinate systems. In section 4.2 we discuss some confusions that may arise from mixing different reference frames.

Kinematic and *dynamic* problems on a flat earth can be described by motion and forces in the horizontal and vertical directions. Velocities and forces are described by vectors. This means that they have a direction and a magnitude and can be split up into vector components that are parallel to the axes of a Cartesian coordinate system. For combined paths we can therefore use the rues of vector algebra. Be careful to note, however, that all motions (and forces) are relative! This is important to realize as an observed motion in *one* place of the globe must not imply that this motion is elsewhere the same (for example China is moving towards Tibet, but the region *between* China and Tibet is under extension). Fig. 2.7 illustrates some more examples. All this is quite different on a spherical surface.

2.2.2 Geometry on a Sphere

On a spherical surface the position of a point is described by its longitude ϕ, and latitude λ (Fig. 2.8). As with time, spherical geometry is one of the few branches in science where the duo-decimal system is still in use: a right angle has 90 degrees and longitude and latitude around the globe are divided into 360 degrees. (The use of 100 degrees for a right angle was attempted by the introduction of "new degrees" but has not found footing in science). Every degree of longitude is described by a great circle which goes through the geographic poles. These great circles are called *Meridians*. Great circles are lines on the surface of a sphere that are defined by the intersection of a planar surface through the center of the sphere, with the surface of that sphere. Meridians are therefore a special kind of great circle, namely one that goes through the poles. Small circles are defined as intersections of all other planar surfaces with the surface of a sphere. 180 of the 360 Meridians are numbered west of Greenwich and the other 180 east of Greenwich, which has been internationally agreed upon to be the reference for longitude. Each degree of latitude is defined by a small circle parallel to the Equator and at right angles to the axis that connects the poles. 90 degrees of latitude

are north of the equator and 90 are south. Note that there is a total of 360 Meridians, but only 180 degrees of latitude. The spacing of the degrees of latitude is chosen so that they divide the Meridians into 360 sections of equal length. Thus, the distance (along the surface of the earth) between degrees of latitude is constant everywhere on the globe, while the distance between degrees of longitude is largest at the equator and zero at the poles. For more detailed description of locations on a spherical surface, every degree is divided into 60 arc minutes and every arc minute into 60 arc seconds. Just to make things worse, the duo-decimal system is often coupled with the decimal system: Geographic locations are often described by degrees and decimals. That is, tenth and hundredths of degrees are given, rather than arc minutes and arc seconds.

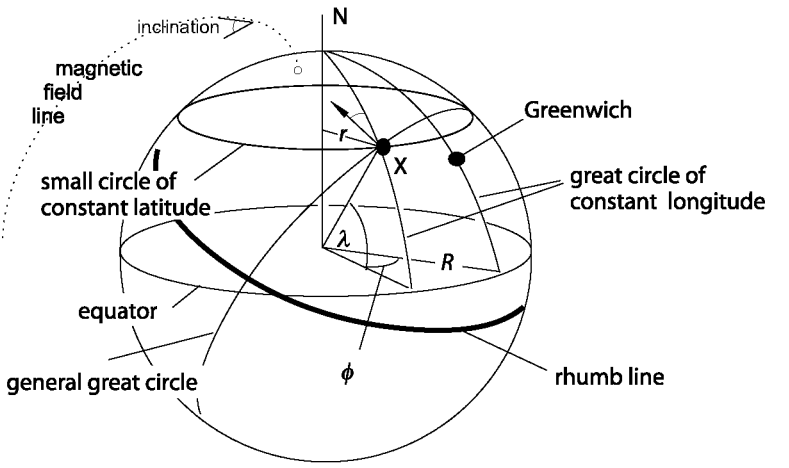


Figure 2.8. Definitions of important lines and angles on a spherical surface. The geographic longitude ϕ of point X is *west* of Greenwich. The white dot is the magnetic pole. The angle between magnetic and geographic north (labeled at point X) is the declination and the plunge angle of the magnetic field lines is the inclination

The circumference of a great circle on earth is about $2R\pi \approx 40\,000$ km. (If it were exactly 40 000 km, then the radius of the earth would have to be $R = 6\,366.2$ km; in reality the equatorial radius is 6 378.139 km and the polar radius is 6 356.75 km). In fact, one meter was long defined as the 1/40 000 000 part of the circumference of earth. One degree of longitude at the equator (and *all* degrees of latitude) is therefore about $40\,000/360 \approx 111$ km. On small circles north and south of the equator, the distance between full degrees of longitude, l, decreases with the cosine of the latitude:

$$l \approx \cos(\lambda) \cdot 111 \quad . \tag{2.2}$$

In eq. 2.2 we have used the approximate value for one degree of longitude at the equator. Correspondingly, the small circle radius of each small circle of

constant latitude shrinks with the cosine of the latitude: $r = \cos(\lambda) \times R$, where R is the radius of earth. One arc minute of latitude is defined as one nautical mile which is $\approx$ 1.8 km. Along the equator, distances between degrees of longitude and latitude are of equal length.

Other important lines on spherical surfaces are *rhumb lines* (also called *loxodromes*). These are lines that intersect degrees of latitude and longitude at constant angles. Rhumb lines are easy to follow, for example when setting constant course on a ship, but they form spiral-shaped curves on a sphere and they are *not* the shortest connection between two points (Fig. 2.8). The angle between magnetic north and the lines of longitude (geographical north) is called the magnetic declination. The vertical angle between the normal to the Geoid surface and the magnetic field lines is called the magnetic inclination.

All distances and angles on a spherical earth can be calculated with simple combinations of the trigonometric functions. Throughout the book it is assumed that the use of those, as well as some sound spatial imagination is familiar to the reader. As a reminder, some of the relationships we will need are summarized in Appendix C, Tables C.4, C.6, C.7, and Fig. C.1.

2.2.3 Kinematics on a Sphere

On a flat surface, velocity v and speed have the units of m s^{-1}. Velocity is a vector and speed is a scalar quantity. For example, the Indian Plate has a *speed* of 0.05 m s^{-1}, but a *velocity* of 0.05 m s^{-1} *moving north.* The equivalent to velocity on a spherical surface is the angular velocity w. w has the units of radian per time, which is s^{-1}. The axis that is perpendicular to the planar surface swept over by angular motion is called the pole of rotation or *Euler pole* (Fig. 2.9). The velocity that corresponds to a given angular velocity depends on the distance of the angular motion from the pole of rotation. Acceleration in a straight line is the change of velocity over time and has the units m s^{-2}. Correspondingly, the angular acceleration has the units of s^{-2}. The differences in units between linear velocity and angular velocity has lead to a lot of confusion in the literature. For example, a *constant* rate of plate motion with a constant angular velocity will cause differences in the rates of relative plate motions along the plate margin. The relative plate motion rate depends on the small circle radius of the velocity vector (Fig. 2.9). In fact, even qualitative changes from divergent plate motion to convergent plate motion may occur along a plate margin because of this (compare Fig. 2.7 and 2.9). This is spectacularly illustrate by the transform faults on the ocean floors which solve the space problem caused by the angular rotation.

2.2.4 Mechanics on a Sphere

Plate tectonic forces are often described in the literature as "torques" (e.g. Sandiford et al. 1995). For example, ridge "push" is a force, while many authors rather use the term ridge "torque". Strictly speaking, we should

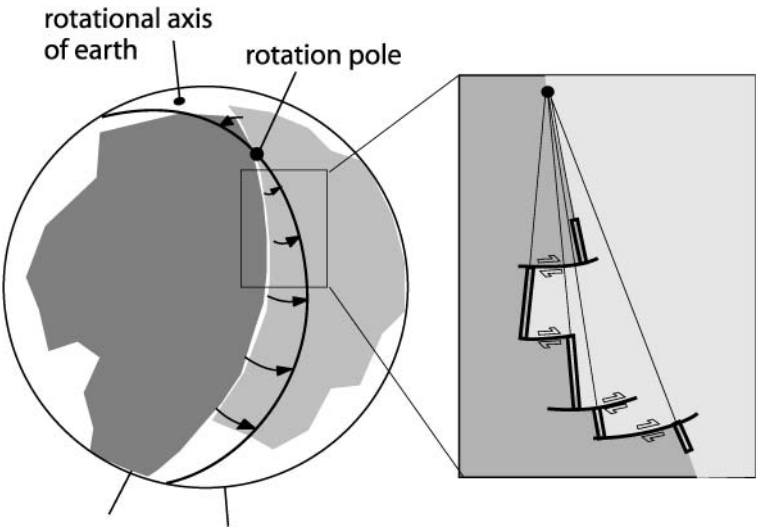


Figure 2.9. Illustration showing the meaning of rotation poles. The arrows are vectors showing the direction and magnitude of relative motion of the two plates (shaded regions). The thick line connecting the arrow heads is the new plate margin after some time. The axis of the earth is only shown to emphasize that it has nothing to do with the rotation pole of plate motion. The enlargement illustrates the origin of transform faults

only use torques to understand the mechanics that cause plate motion on the earth's surface as plates do not move in a linear direction but rather around a rotation pole (the center of earth). In calculating a force or torque balance of a mountain belt, where every point in the belt is virtually the same distance from the pole of rotation, the distance to the rotation pole cancels out and torque balances and force balances are practically equivalent (s. p. 216).

Force F is given in Newtons [N] and: $1\ \mathrm{N} = 1\ \mathrm{kg\,m\,s^{-2}}$. Force is a vector with a magnitude and direction. Horizontal forces are therefore tangential to the globe. The equivalent on a spherical surface is torque. Torque (which is different from angular momentum!) is the turning moment which is exerted by a force about an axis. It is given by the product of force and the distance from the axis about which the torque acts. Torque has the units of Nm or $\mathrm{kg\,m^2\,s^{-2}}$. A force of 10^{12} Newton that acts in direction of a great circle on the earth's surface, corresponds to a torque of $6.37 \cdot 10^{18}$ Nm. The torque changes along a plate margin, as the normal distance of the plate margin to the rotation pole changes. The units of torque can be read as "Newton times meter of leverage", where the "meters of leverage" are the normal distance to the rotation pole. In the literature, "forces" are often given in Newton per meter, meaning for example, that the force building a mountain range is normalized "per meter length of orogen". It is important not to confuse this

Table 2.1. Important kinematical and mechanical parameters and their units. Each parameter is given with both, the linear and the spherical equivalents

physical parameter	unit
velocity	$\mathrm{m\,s^{-1}}$
angular velocity	$\mathrm{s^{-1}}$
acceleration	$\mathrm{m\,s^{-2}}$
angular acceleration	$\mathrm{s^{-2}}$
force	$\mathrm{kg\,m\,s^{-2}}$
torque	$\mathrm{kg\,m^2\,s^{-2}}$
mass	kg
moment of inertia	$\mathrm{kg\,m^2}$
linear momentum	$\mathrm{kg\,m\,s^{-1}}$
angular momentum	$\mathrm{kg\,m^2\,s^{-1}}$

with torques, which have the units of Newton times meters, there the meters are the distance to the rotation pole.

Force = mass × acceleration ($F = m \times \mathrm{d}v/\mathrm{d}t$) and similarly torque = mass × angular acceleration. In plate tectonics the changes in velocity and angular velocity occur over very long time periods, so that accelerations and angular acceleration are negligible (s. p. 218). Thus, the common assumption is that the sum of the torques or the net torque acting on a plate is zero or, correspondingly, that the sum of the forces or net force acting on a smaller region such as a mountain belt is zero.

Aside from force and torque, there are some other important mechanical parameters that we will need in this book. The linear momentum I is the product of mass and velocity: $I = mv = \mathrm{kg\,m\,s^{-1}}$. Just as momentum is given by $I = mv$, the *change* of momentum is given by: $\Delta I = m\Delta v$, mass times the *change* in velocity. Considering that force has the units of the product of mass and acceleration (the change in velocity) we can write that: $F = m\mathrm{d}v/\mathrm{d}t$. Thus, the change of momentum is: $\Delta I = Ft$, force acting over a given time. On a spherical surface, the angular momentum is analogous to the linear momentum. The angular momentum D is the product of the moment of inertia J and the angular velocity w: $D = Jw$. The moment of inertia is the ratio of torque and angular velocity and has the units of $J = \mathrm{kg\,m^2}$. Angular momentum has the units of $D = \mathrm{kg\,m^2\,s^{-1}}$. In plate tectonics, changes in momentum and angular momentum are ignored because changes in velocity and angular velocity of plates takes place over very large time periods (s. sect. 5.1.1). In most problems then, the net forces or torques acting on a given plate must balance or add up to zero (s. p. 216). However, momentum

and angular momentum help to understand why forces *or* velocities can form the boundary conditions for orogenic processes.

2.3 Map Projections

Map projections are mathematical or geometrical models that portray the features of the spherical surface of our globe on the two dimensions of paper. The map projection model serves the purpose to make these features easier to look at and pays for that by accepting some distortion. Usually, the process of projection involves the conversion of latitude λ and longitude ϕ into Cartesian coordinates (Robinson et al. 1984). Map projections have become an increasingly important part of geology since global data sets are used. Map projections can be performed geometrically or they can be purely mathematical, without any apparent geometrical equivalent. Among projections *with* a geometrical equivalent we discern three important types (Fig. 2.10):

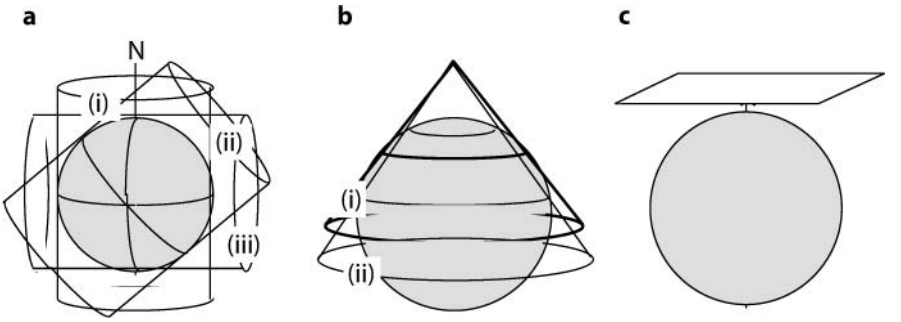


Figure 2.10. Schematic illustration of three examples of commonly used geometrical map projections. **a** *Cylindrical projection* with examples of (i) normal or equatorial, (ii) transverse and (iii) oblique orientations of the projection surface; **b** *conical projection* with examples of (i) tangential orientation (touching the sphere along the thick line) and (ii) secant orientation (intersecting the globe) of the projection surface; **c** *Azimuth* projection with polar orientation of the projection surface

- Cylindrical projections,
- conical projections,
- azimuth projections.

These three types of projections can be imagined as projections of the earth's surface from an imaginary light source (usually assumed to be located at the center of earth) onto some enveloping surface which is then rolled open to form a planar surface. In these three projections, the enveloping surface is a cylinder, a cone or a planar surface. The enveloping surface can touch the globe along one line (tangential orientation) or transect it (sectant orientation) so that it has two lines along which there is no distortion. In total,

there are several hundreds of known projections that serve a large variety of purposes. A large number of other projections are explained by Snyder (1987) and Snyder and Voxland (1989). All their projections are usable with the software PROJ by Evenden (1990) and there is great overlap with the projections incorporated in GMT and described therein (Wessel and Smith 1995). A bit more detail was also given in the first edition of this book. Most known projections are *not* geometrical but are only described by a *projection function* that relates spherical and planar coordinates. Just about all map projections are either:

- Conformal (orthomorphic) projections,
- equivalent (equal-area) projections,
- azimuthal projections.

Conformal projections represent angular relationships the way they are on the spherical surface. *Equivalent* projections render the same areas as those on the spherical surface. *Azimuthal* projections retain azimuthality, that is, the directional relationships along great circle bearings are the same on the map as they are on the sphere. No projections on two-dimensional paper can retain more than one of these relationships from a sphere at the same time. However, many projections aim to find a useful compromise between different relationships, on the expense of being neither true in angle nor true in area or great circle bearing.

Most geologists are familiar with at least one conformal and one equivalent projection, even if they have never used maps of the globe. These are the nets of Schmidt (also known as *Lambert's equivalent projection*) and Wulff (also known as *stereographic projection*) that are used by crystallographers for the presentation of angular relationships of crystal faces and by structural geologists for the equivalent illustration of planar structures, respectively (Fig. 2.11). Note that lines of constant latitude and longitude need not look like they do on Fig. 2.11. If the orientation of the projection surface is oblique to the axis of the globe, then these lines may have very different shapes as shown on Fig. 2.12.

2.3.1 The Mercator Projection

The *Mercator projection* is one of the most familiar projection and is one of several known cylindrical projections where the projection surface is a cylinder as shown in Fig. 2.10a. It was first used by G. Mercator in 1569 for navigation on the oceans. For this projection, the projection surface is a cylinder a cylinder that touches the globe along the equator and has its axis therefore parallel to the axis of earth (Fig. 2.10a). Every point of the earth's surface is projected onto this cylinder from an imaginary light source at the center of earth. The imaginary cylinder is then rolled open. The map is ready! There is two great advantages of this projection:

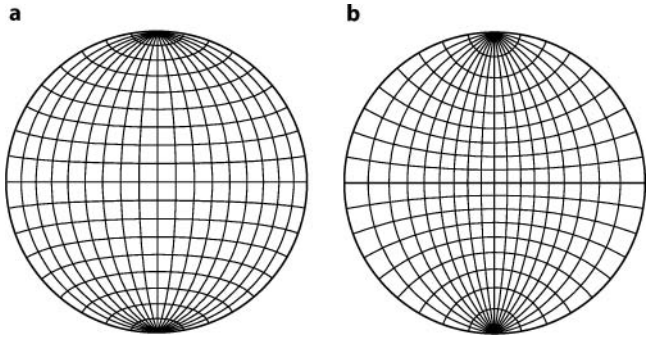


Figure 2.11. a The equal-area net by Schmidt. **b** The conformal net of Wulff

1. There is *no* distortion along the equator and the distortion in equator-near regions is negligible.
2. Lines of constant longitude, latitude and rhumb lines appear as straight lines.

Because rhumb lines are easy to navigate along, Mercator projections used to be common for shipping (before navigation software made more direct routing much easier). However, note that the shortest connection between two points on the globe – a great circle – appears on the Mercator projection as a curved line. The biggest disadvantage of the Mercator projection is that the areal distortion increases with latitude. Areas near the poles appear too large and the poles themselves cannot be represented at all. For example, Greenland appears to be larger than south America although it has only an eight part of its size.

This problem can be avoided by using the *transversal Mercator projection.* There, the cylinder touches the globe along one of the Meridians and its axis is at right angles to the axis of the earth. Accordingly, there is no distortion along this Meridian and the distortion goes towards infinity on the equatorial points of the Meridian that is 90° from the Meridian where the cylinder touches. Both lines of constant longitude *and* latitude are curved on this projection.

The UTM Projection. The Universal Transversal Mercator projection (UTM) is defined by a total of 60 different Meridians (all separated by 6 degrees of longitude) all of which are touching Meridians for 60 assumed projection cylinders (Fig. 2.13). In other words, it is a combination of 60 transversal-Mercator projections. This avoids serious distortion of any point on the surface of the globe.

2.3.2 National Grids

Using latitude and longitude is not always the best way to define a point on a map. Lines of constant latitude and longitude may be curved lines or – even

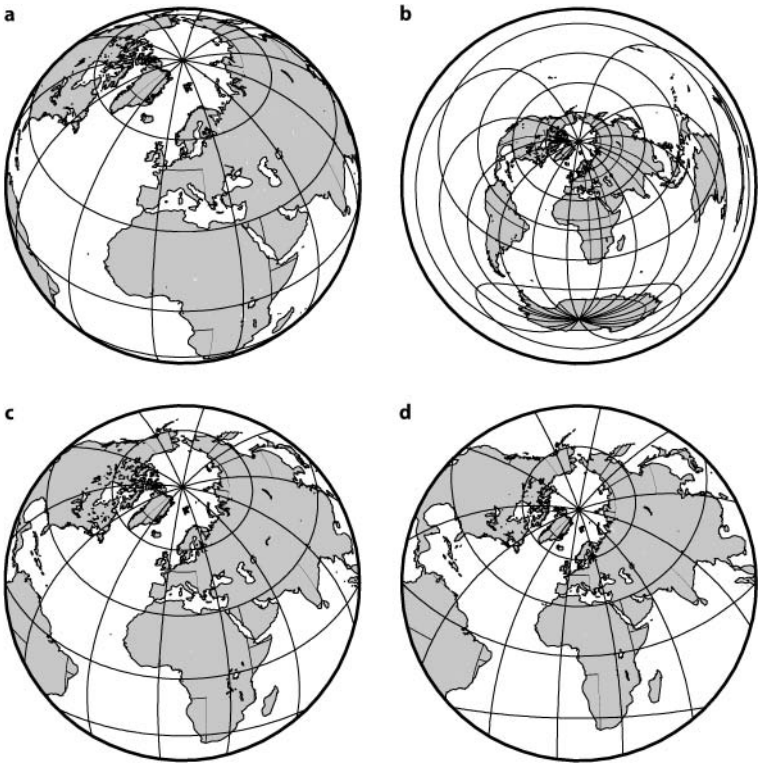


Figure 2.12. Four examples of map projections with circular outlines. **a** orthographic projection, **b** equidistant azimuthal projection, **c** Lambert's equal area projection and **d** stereographic projection. All shown examples are for inclined orientations of the rotation axis of earth relative to the projection direction. If **c** and **d** were projected parallel to the axis of earth, then lines of constant longitude and latitude in **c** would correspond to those of Fig. 2.11a and lines of constant longitude and latitude in **d** would correspond to those of Fig. 2.11b

if they are straight lines – their spacing is unlikely to be constant. Thus, in order to simplify the use of maps, *national grids* are used. National grids are orthogonal grids with equal spacing in two horizontal directions x and y that are superimposed on projected maps (Figs. 2.13; 2.14). Different countries use different grids for their national mapping but many of them are based on the UTM projection or slight modifications thereof.

For example, topographic maps in Austria are based on the Gauss Krüger projection which differs from the UTM projection by having transversally oriented cylinders touching earth (described by a Bessel ellipsoid) every 3 degrees (not 6), starting at Ferro (instead of Greenwich) and going east (for details on the Bessel ellipsoid and its difference to the WGS84 system see sect. 4.2.1). Within this projection, the axes of the superimposed orthogonal national grid are assumed to be the equator and the touching Meridian (which are called M28, M31 and M34 in Austria). x and y values on the orthogonal

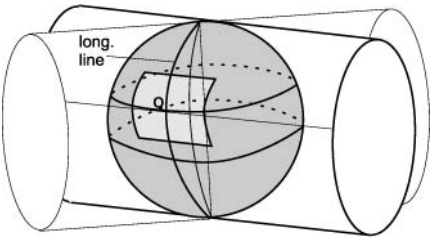


Figure 2.13. Illustration of the UTM projection. A series of transversally oriented cylinders touch the reference ellipsoid along longitudinal lines. A given point along this line is then chosen as the origin of an orthogonal national grid with northing and easting values.

grid are then counted eastwards from the touching Meridian in kilometers and northwards from the equator, respectively. Just to make it a bit more complicated surveyors have decided that – in Austria – the y values get too large if counted from the equator and that negative x values west of the touching Meridian look ugly. To resolve this, the origin at the equator is assigned an arbitrary x and y value (x=150 km and y = - 5000 km for the M28) so that the numbers at 45° have reasonable values and eastings don't get negative.

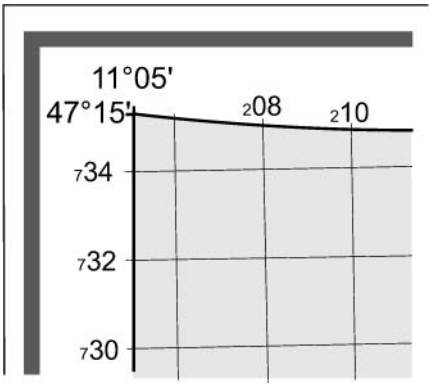


Figure 2.14. Example of the top left corner of a topographic map sheet. Typically, referenced map sheets are bound by lines of constant latitude and longitude. The "lat" and "long" values are indicated at the corners. Because these may be curved (depending on the projection), the corners are usually not parallel to the edges of the paper. The superimposed national grid is usually slightly oblique to the lat and long lines and is labelled in kilometer. The fact that the first digit is printed small has no significance other than to save some space

Similarly, the National Imagery and Mapping Agency (NIMA) of the United States has divided the United States into 10 zones (numbered from 10 through 19), each of which is 6 degrees wide and which lie between 126° west and 66° west. Within each zone, a transversally oriented cylinder is assumed to touch a central Meridian and northings and eastings on the superimposed orthogonal grid are counted in kilometers from 10 000 km south of the equator and 500 km west of the touching Meridian. The shifting of the grid origin for each zone to 10 000 km south and 500 km to the west is done to avoid negative numbers south of the equator or on western halves of zones, because the US use this system for maps world wide.

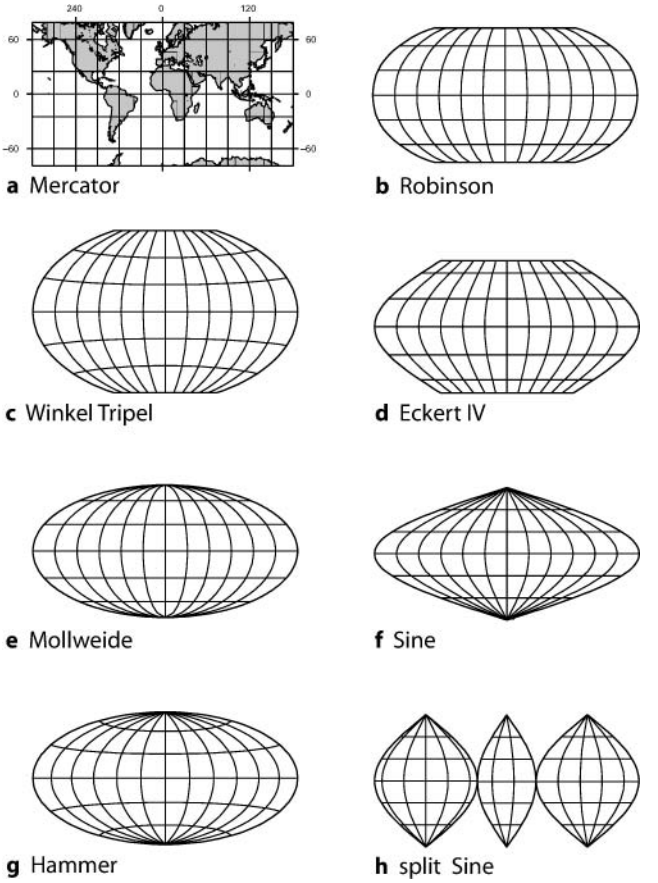


Figure 2.15. Eight examples of projections commonly used to portray the entire globe. **a** Mercator projection shown up to 75 degrees north. **b** The Robinson projection is pseudocylindrical and is neither conformal nor equivalent but "looks right". **c** Winkel tripel projection. **d** The Eckert IV projection is pseudocylindrical and equivalent. It is common in Atlases. **e** The Mollweide projection is also pseudocylindrical and equivalent with longitudes being straight lines and Meridians parts of ellipses. **f** The equivalent sine projection is often used for south America.**g** The Hammer projection is also equivalent. **h** A split version of the sine projection shown above

2.4 The Layered Structure of Earth

Earth can be divided into layers according to:

- different materials,
- different physical properties.

When considering the layers made of *different materials*, there is three: a crust, a mantle and a core (layers above the surface of solid earth like the

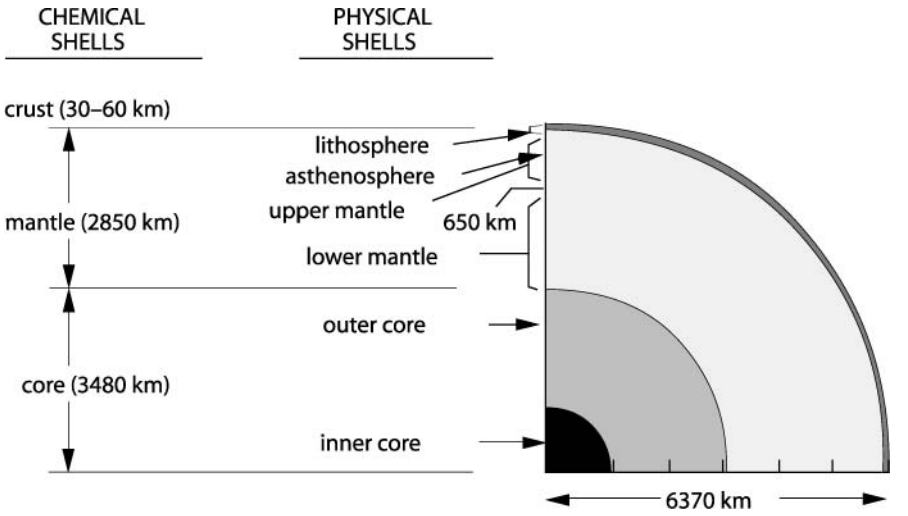


Figure 2.16. The layered structure of earth

hydrosphere, biosphere or atmosphere are not considered here). The crust is the uppermost layer. In its normal state, it is between some 5 to 7 and 30 km thick, depending on whether we deal with oceanic or continental crust. In regions of deformed continental crust – for example underneath the Tibetan Plateau or the Alps – continental crust can get up to 80 km thick. Chemically, the crust is highly-differentiated and very heterogeneous, but many of its mechanical and physical properties (e. g. density, conductivity or rheology) can be well-approximated with those of quartz. The mantle is largely made up of olivine and – at larger depths – its high-pressure breakdown products. The seismically clearly visible contact between crust and mantle is called the Mohorovičić-discontinuity (short: Moho). From the Moho the mantle reaches down to a depth of about 2900 km. The core consists mainly of iron and nickel.

When considering the *physical properties*, the layered structure is quite different. Then, the outermost layers of earth are the lithosphere and the asthenosphere. The lithosphere is solid and acts like rock on geological time scales. Therefore its name. It involves both a crustal and a mantle part. The asthenosphere consists of the soft mantle that underlies the lithosphere. Some authors call the entire upper mantle underneath the lithosphere the asthenosphere. Others use the term only for the mantle section that lies above the point where the adiabatic melting curve comes nearest to the temperature profile (Fig. 3.11). According to Ringwood (1988) the mantle can be divided into three zones:

1. The upper mantle, which reaches down to about 400 km and is characterized by a seismic p-wave velocity of about 8.1 $\mathrm{km\,s^{-1}}$.
2. A transition zone from about 400 km to the 650 km discontinuity.

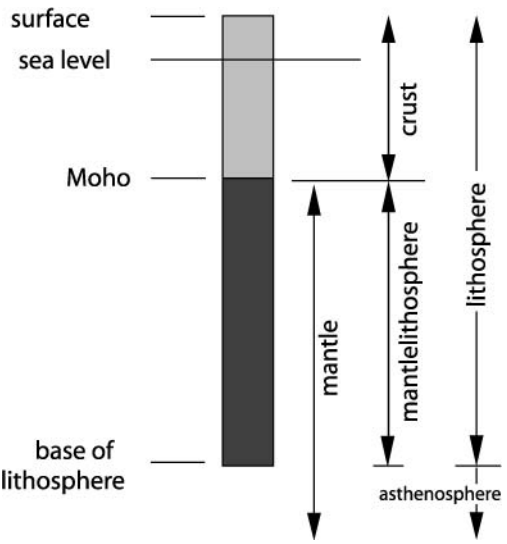


Figure 2.17. Nomenclature of different parts of the outer shells of the earth

3. The lower mantle which reaches from the 650 km discontinuity to the core-mantle boundary at 2 900 km depth.

Below that is the core. The outer core is liquid and the inner core is solid.

2.4.1 Crust and Lithosphere

The lithosphere is the outer solid layer of the earth (s. sect. 3.4). As for the whole globe, the lithosphere can be divided according to its physical properties or according to its chemical (material) properties. Because there is overlap between layers distinguished on the basis of different properties it is crucial for the understanding of this book to be familiar with the nomenclature illustrated in Fig. 2.17. However, for many geodynamic questions it is not necessary to consider these subdivisions, as the lithosphere acts as a whole.

When considering its chemical properties, the lithosphere consists of a crust and a mantle part. The crust consists of highly-differentiated partial melts from the mantle. The mantle part of the lithosphere is largely made up of the similar material to that of the underlying asthenospheric mantle, but it acts like a solid, because of its lower temperature. However, we note that chemical differences between the mantle lithosphere and the underlying asthenosphere do exist and account for example for unusually thick, but apparently mechanically stable mantle lithosphere, underneath southern Africa. Modern research has been able to document much detail of the compositional variation within the uppermost mantle, both on chemical grounds (e.g. McDonough and Rudnick 1998) and based on seismic velocities (Jordan 1981a;

1989). Nevertheless we take in this book the simple-minded view that density variations between mantle lithosphere and asthenosphere may be largely attributed to differences in temperature (e.g. on p. 165).

A schematic but characteristic thermal and density profile of the lithosphere is shown in Fig. 2.18. A very large number of the geodynamic processes discussed in this book are a function of the fundamental shape of the curves on this figure.

Within the crust the temperature profile is curved, because of radioactive heat production. Within the mantle part of the lithosphere, the thermal profile is linear (in a steady state). The base of the thermal lithosphere is defined by the point where the temperature profile intersects the 1 200°C or 1 300°C isotherm (sect. 3.4). At higher temperatures, mantle material begins to flow rapidly on geological time scales and any temperature gradients will be eliminated by convection. Thus, temperature and density are constant below the depth z_l on the scale of Fig. 2.11. Both curves of Fig. 2.18 will be useful help throughout this book.

Definition of the Lithosphere. The term "lithosphere" comes from the Greek *lithos* = rock and was introduced by Suess (1885). The term was later used by Barrell (1914) and ultimately defined by Isacks et al. (1968) as a *near surface layer of strength* of earth. Even today it remains difficult to find a more precise definition than this. Most of the physical parameters, for example temperature or density, change continuously underneath the Moho and the transition from the rigid outer shell of the earth (the mechanical boundary layer) into the more viscous hot asthenosphere (from the Greek *asthenia* = soft) is also continuous. This transition zone is called *thermal boundary layer*) (Fig. 2.19; sect. 6.3.2; Parsons and McKenzie 1978; McKenzie and Bickle 1988). However, even on the definition of the term "thermal boundary layer" there is no clear consensus in the literature. Some authors refer with this term only to the transition zone between lithosphere and asthenosphere and others to the entire lithosphere as being a thermal boundary layer to earth (s. p. 89, p. 64 and Fig. 2.19).

One thing can be said with certainty: the definition of the lithosphere depends on the question that is being asked. For example, it can be shown that the thickness of the lithosphere is a function of the observed time scale. Seismic motion, isostatic uplift and ductile deformation occur on time scales of seconds, 10^4 y and $> 10^6$ y, respectively. The larger the time scale of the process, the smaller the thickness of the lithosphere. Seismically, the lithosphere is of the order of 200 km thick, while the elastic thickness of the lithosphere is only some tens of kilometers thick. Very generally the lithosphere may be defined *mechanically* as the outer part of the earth in which stresses can be transmitted on geological time scales (s. McKenzie 1967). According to a somewhat different mechanical definition the thickness of continental lithosphere may be defined as the layer that is in isostatic equilibrium with the mid-oceanic ridges (Cochran 1982). This is meaningful, because the mid-

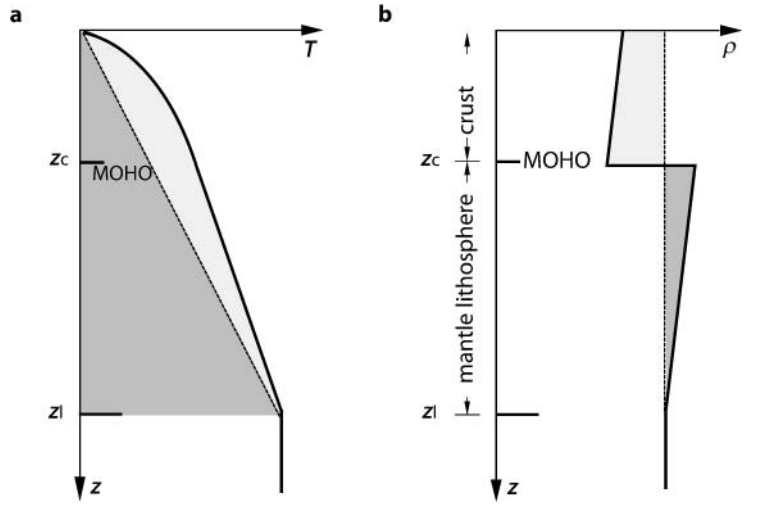


Figure 2.18. **a** Temperature and **b** density of the continental lithosphere as a function of depth. The depth of the Moho below surface is z_c, that of the whole lithosphere is z_l. **a** The curvature of the geotherm within the continental crust is caused by radioactive heat production. The light shaded area corresponds to the heat content of the lithosphere that can be attributed to radioactive decay in the crust. The dark shaded region corresponds to the heat content conducted into the lithosphere from the asthenosphere. The relative contributions of the radioactive and the mantle heat flow will be discussed in sect. 3.4.2 and sect. 6.2.1. Here it may be seen that the heat content of the crust consists to roughly equal part of mantle heat and of radioactive heat. **b** The slope of the density profile within the crust and within the mantle lithosphere is a function of the thermal expansion. A comparison of the shaded areas shows that the density deficiency in the crust (light shaded area) is comparable to the density excess in the mantle lithosphere (dark shaded area) – both relative to the asthenosphere. Within the asthenosphere convective flow equalizes all density and temperature heterogeneities. Both curve are therefore vertical

oceanic ridges may be interpreted as manometers of the upper mantle (s. p. 166, Turcotte et al. 1977).

According to a *thermal* definition the lithosphere is the part of earth in which thermal energy is largely transferred by heat conduction, in contrast to the asthenosphere, where heat is transferred by convection (s. sect. 3.4 for more detail). In some ways the thermal definition encompasses the mechanical definition because many of the mechanical properties of rocks depend on the ratio of their temperature to their melting temperature. In stable continental lithosphere, thermal and mechanical definitions indicate thicknesses of 100–150 km. The thickness of the crust and its content in radioactive minerals is crucial to the thickness of the lithosphere, because they strongly influence the Moho-temperature (s. Fig. 3.20). Many studies that are concerned with the rheology of the lithosphere avoid to use a number for lithospheric thickness. Rather, the lithosphere is defined via the Moho-temperature (s. sect. 6.2.2).

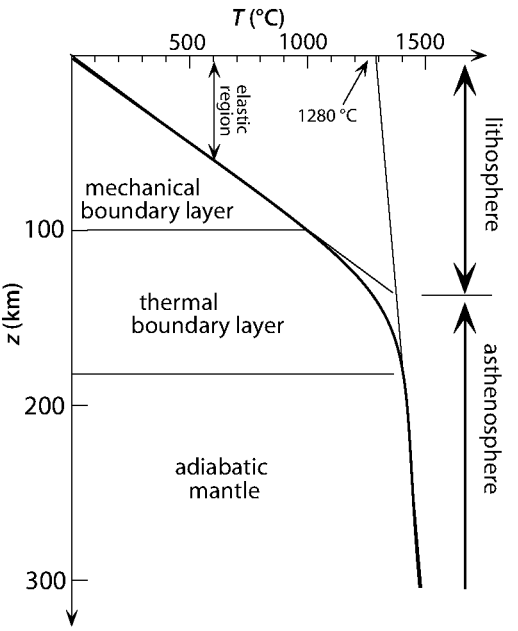


Figure 2.19. Definition of the mechanical and thermal boundary layer as well as of the lithosphere and asthenosphere according to Parsons and McKenzie (1978) and McKenzie and Bickle (1988). The thick line is the geotherm. Its curvature in the upper crust is ignored. However, note that other definitions for thermal boundary layer are also in use in the literature s. p. 89, p. 64)

Types of Lithosphere. There is two fundamentally different types of lithosphere on earth: *oceanic* and *continental* lithosphere. Despite these names, the correlation of oceanic lithosphere with the geographic area of the ocean and vice versa is only very approximate and there are substantial parts of continental lithosphere that lie under water and vice versa. A very rough indicator for the line separating continental from oceanic lithosphere is about 500 m water depth. On Fig. 2.1 most of the white colored regions within the oceans are actually continental lithosphere.

• *Oceanic lithosphere.* Oceanic lithosphere begins its life at the mid oceanic ridges. There, it consists only of an about 7 km thick oceanic crust, which is made up of crystallized partial melts from the uppermost mantle. The thickness of the mantle part of the oceanic lithosphere is zero near the mid-oceanic ridges. With increasing age – that is: with increasing distance from the ridge – the thickness of the mantle part of the oceanic lithosphere increases as the asthenosphere successively freezes to the base of the cold crust. In the oldest parts of known oceanic lithosphere the thickness of the oceanic mantle lithosphere is almost as thick as continental mantle lithosphere. However, oceanic lithosphere is being produced and consumed at all times, so that there is hardly any oceanic lithosphere on earth that is much older than about 150 my. Because of the young age (and therefore low thickness) the mean oceanic heat flow is significantly higher than that of the continents. Although there is practically no radiogenic heat production in oceanic crust

the mean oceanic heat flow is 101 $mWm^{-}2$ (Pollack et al. 1993; Wei and Sandwell 2006) (s. p. 17, Figs. 3.1, 2.4).

• *Continental lithosphere.* In contrast to the oceanic lithosphere, the total area of continental lithosphere has remained largely constant in the entire Phanerozoic. Thus, the present day continents consist largely of Proterozoic continental lithosphere, which has been reworked in many places. Nevertheless there ae enough places around the world where Archaean and Proterozoic continental crust is preserved giving us a glimpse of tectonic processes in the Precambrian (e.g. Greenland, Canadian Shield, East Antarctic Shield, Kongo Craton, Arunta Block, Yilgarn block and many others). Continental crust is chemically highly-differentiated, it has a high content in radioactive elements and in its stable state it is about 30–50 km thick. According to thermal and mechanical definitions, the mantle part of the lithosphere is 70–100 km thick, so that the total thickness of stable continental lithosphere is of the order of 100–150 km. However, in old shield areas this thickness can be much more, probably due to a different chemical composition of the mantle lithosphere. The mean heat flow of the continental lithosphere is about 65 $mWm^{-}2$ (Pollack et al. 1993; Wei and Sandwell 2006).

2.4.2 The Lithospheric Plates

The surface of the earth can be divided into seven major lithospheric plates plus a number of smaller plates (Fig. 2.20, Table 2.2). Not all major plates correspond to the seven continents and it is of some coincidence that the number of continents equals that of the major plates. Most major plates consist of both continental *and* oceanic lithosphere.

Among the seven major plates, the Antarctic Plate and the African Plate have a special position as they are surrounded on just about all sides by mid-oceanic ridges. Both plates increase therefore permanently in size and they form a good example where plates go from a compressive state into an extensional state purely as a function of their increasing age (Sandiford and Coblentz 1994; sect. 5.3.1).

2.4.3 The Plate Boundaries

Most geodynamically interesting processes occur along the plate boundaries. These boundaries can be divided according to:

- their kinematics,
- the types of plates that are in contact.

When choosing a division according to the types of bounding plates, we can discern:

- plate boundaries between two continental plates,
- plate boundaries between two oceanic plates,

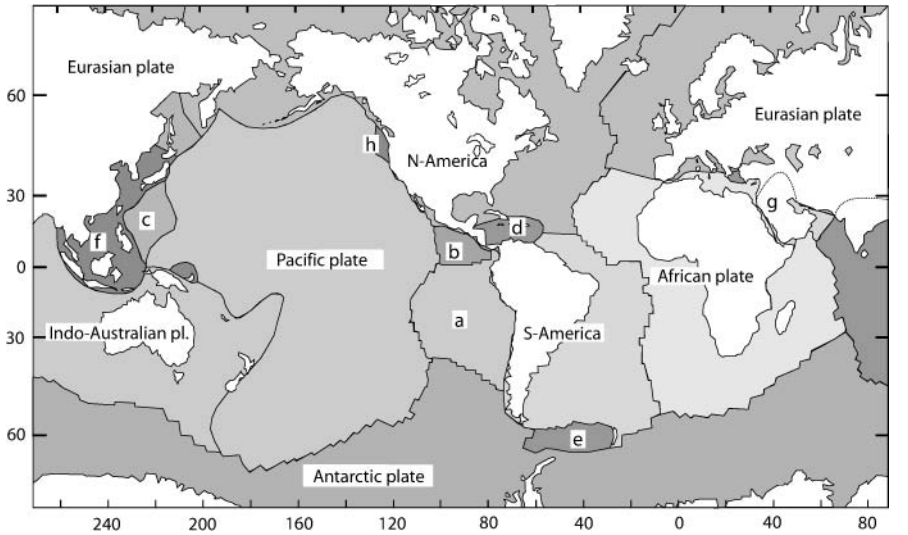


Figure 2.20. Plate tectonic division of the surface of earth. Continents are white and oceans are shaded. The difference between continental and oceanic lithosphere is not shown. For that, see Fig. 2.1. Note that the plate boundaries coincide only in a few places with the coast lines that delineate the continents (s. Table 2.2). The seven major plates are labeled with their names. The most important minor plates are labeled with letters. They are: *a* Nazca Plate; *b* Cocos Plate; *c* Philippine Plate; *d* Caribbean Plate; *e* Scotia Plate; *f* Chinese Subplate; *g* Arabic Plate; *h* Juan-de-Fuca Plate

– plate boundaries between a continental and an oceanic plate.

When choosing a division according to the kinematics, we can discern between *convergent*, *divergent* and *transform* plate boundaries (s. Tables 2.3; 2.4). Passive margins are formerly divergent plate boundaries between two continental plates which now consist of a passive contact between oceanic- and continental lithosphere. They are often listed as its own type of plate margin.

Divergent Plate Boundaries. Divergent plate boundaries are regions where two plates move into opposite directions or where one plate is splitting into two. Divergent plate margins exist only between two continental plates (e. g. central African rift system) or between two oceanic plates (e. g. Mid-Atlantic Ridge). The passive seams between continental and oceanic lithosphere are mechanically very strong and it would be a great coincidence if a divergent plate margin would form exactly along them. However, there are places on the globe where divergent plate margins *cross* passive margins. The Sheba Ridge in the Gulf of Aden and the Carlsberg Ridge in the Indian Ocean are examples. Divergent plate boundaries on the continents are called rifts. The best know examples (in order of progressive rift development) are the Rheingraben, the Central African rift system and the Red Sea. Divergent

Table 2.2. Approximate parts of oceanic and continental lithosphere of each of the major plates

plate	% oceanic lith.	% continental lith.
major plates		
Pacific Plate	100	0
North-American Plate	30	70
South-American Plate	50	50
Eurasian Plate	30	70
Antarctic Plate	50	50
African Plate	50	50
Indo-Australian Plate	40	60
important minor plates		
Nazca Plate	100	0
Cocos Plate	100	0
Juan-de-Fuca Plate	100	0
Scotia Plate	100	0
Philippine Plate	100	0
Caribbean Plate	100	0
Arabic Plate	10	90

Table 2.3. The twelve most important relative motions of plates (after DeMets et al. 1990). 1° corresponds to about 110 km

plate boundary	rotation pole longitude	latitude	angular velocity $\cdot 10^{-7\circ}/y$
Africa – Antarctica	5.6°N	39.2°W	1.3
Africa – Eurasia	21.0°N	20.6°W	1.3
Africa – North-America	78.8°N	38.3°E	2.5
Africa – South-America	62.5°N	39.4°W	3.2
Australia – Antarctica	13.2°N	38.2°E	6.8
Pacific – Antarctica	64.3°S	96.0°E	9.1
South-America – Antarctica	86.4°S	139.3°E	2.7
India – Eurasia	24.4°N	17.7°E	5.3
Eurasic – North-America	62.4°N	135.8°E	2.2
Eurasia – Pacific	61.1°N	85.8°W	9.0
Pacific – Australia	60.1°S	178.3°W	11.2
North-America – Pacific	48.7°N	78.2°W	7.8

plate margins between two oceanic plates (mid-oceanic ridges) are – in most cases – the last stage of a rift (sect. 2.4.4, Fig. 2.21).

Convergent Plate Boundaries. Convergent pate boundaries may form between two continental plates (Fig. 2.15), between two oceanic plates or between a continental and an oceanic plate. In convergent plate boundaries

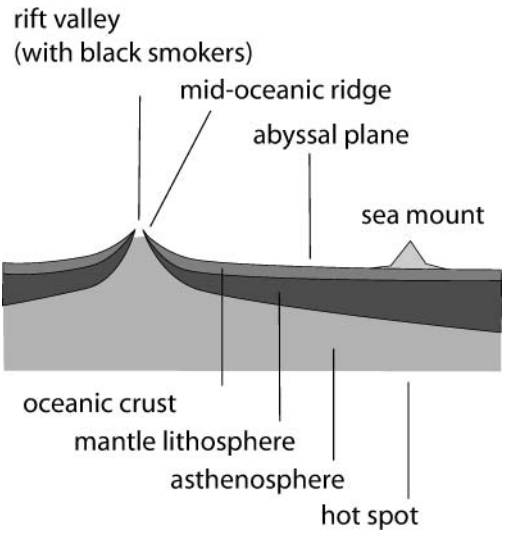


Figure 2.21. Nomenclature of some of the most important parts of oceanic lithosphere

between continental and oceanic plates, the oceanic plate dives beneath the continent, because of its higher density. This is called subduction and its surface expression is a trench (Fig. 2.23). The most famous example for subduction is the subduction of the Nazca Plate underneath the south-American continent along the Peru-Chile Trench. Subduction leads to high pressure metamorphism in the hanging wall of the subducted plate. This metamorphism is associated with dehydration and partial melting of the plate of the oceanic plate in the *Benioff zone*. Fluids that rise through the overlying mantle wedge react with the wedge material in endothermic reactions (Fig. 3.28). However, additional heat input by convection in the wedge leads to partial melting (Hoke et al. 1994) and ultimately to the development of volcanic arcs on the surface.

The kinematics of subduction zones is complicated. The forces and velocities with which subducting plates sink into the asthenosphere are comparable to the forces exerted by mid-oceanic ridges onto the plate. Subduction zones can therefore move *backwards* (towards to mid-ocean ridge) if the downward velocity of the subducting plate is *larger* than the rifting rate at the ridge (e. g. South-Georgia, Scotia Plate). They can move *forward* (towards the continent) if the rifting rate at the ridge is *larger* than the downward velocity (e. g. Pacific Plate – Alaska). In other words, the distance between the trench and the continent in the far field hinterland increases, decreases or remains constant. Depending on details of the force and velocity field in subduction zone environments, forearc- or backarc basins may develop. In some cases of collision between oceanic and continental lithosphere, parts of the oceanic plate are welded onto the continental plate or even thrust over it. This is called *obduction*.

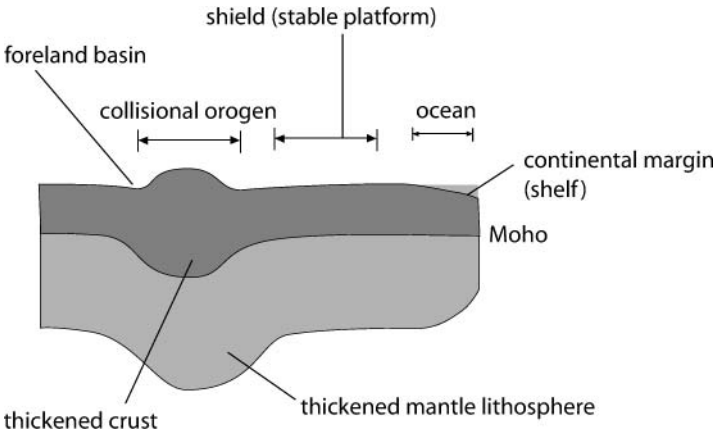


Figure 2.22. Nomenclature of important parts of continental plates

In contrast to subduction zones – where one of the two plates usually dives underneath the other – the convergence of two continental plates leads to a

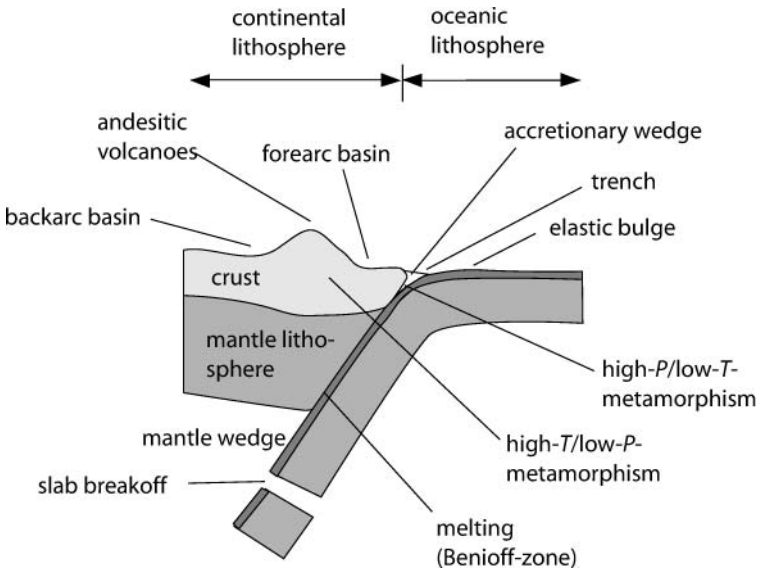


Figure 2.23. Nomenclature of important parts of subduction zones. Forearc- and backarc basins are even more clearly developed during subduction of oceanic lithosphere underneath another oceanic plate. Good descriptions of various phenomena on this figure can be found for slab break off by: Blanckenburg and Davies (1995); about the mantle wedge by: Spiegelman and McKenzie (1987) (sect. 3.5.2); about metamorphism by: Miyashiro (1973); about accretionary wedges in: sect. 6.2.3; about the elastic fore bulge: sect. 4.4.2

head-on collision of both plates. The reason for this is because continental lithosphere is (i) much thicker, (ii) less dense and (iii) much softer than oceanic lithosphere. Why the continental lithosphere is much softer than the oceanic lithosphere although it is much ticker will be discussed in sect. 5.2. This leads to the formation of the collisional mountain ranges that form most of the topographically high mountain belts of our globe (Fig. 1.1). During this head on collision the crust gets typically pervasively deformed, while the mantle parts of the lithosphere override each other (a bit like shown schematically on the cover of this book).

When two oceanic plates converge, no collision occurs and subduction zones form, similar to those that form when two plates of different kind collide. In contrast to the collision between two continental plates, *no* collision occurs between two oceanic plates because they are thinner, much stronger and because they are much denser and can therefore dive easier into the upper mantle (sect. 5.2.2). Because little internal deformation of the plates occurs, island arcs form that are clearly defined in space. Two beautiful examples for this are the subduction of the Pacific Plate underneath the Philippine Plate along the Mariana Trench or the subduction of the Pacific Plate underneath North-America along the Aleutes.



Figure 2.24. Topography of the Dragonsback Ridge segment of the San Andreas Fault in California. The shown segment is about 3 km long and the resolution is less than 0.5 m. The fault is oriented northwest - southeast (top left to bottom right). Image was obtained by the B4 Project using Airborne Laser Swath Mapping (Hudnut and Bevis, pers. comm., 2006)

Transform Plate Boundaries. When two plates glide past each other without much convergence or divergence, their contact is called a transform plate boundary. No topographic features as significant as rift valleys or mountain ranges form. However, transform plate boundaries are well-known because they form some of the most important zones of seismicity on the globe. The best known examples are the San-Andreas-fault zone (Fig. 2.24), (probably because it crosses one of the most densely populated parts of North-America) or the Alpine fault in New Zealand.

Triple Junctions. The spherical geometry of the earth requires that three or more plates touch each other in some places. This is called a triple junction.

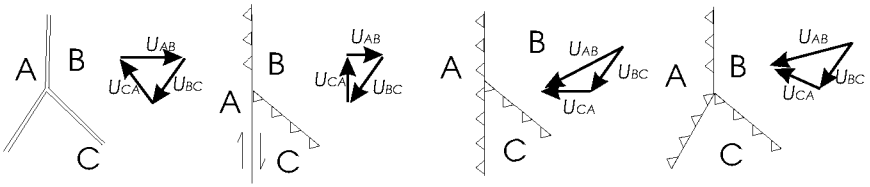


Figure 2.25. Cartoons of four types of triple junction and vector diagrams showing their relative velocities. The shown examples are all stable because the sum of the vectors results in zero velocity

Triple junctions have a *stable* configuration if the relative motions of the bounding plates can maintain the geometry through time. They have an *unstable* configuration if their geometry is transient. Places where four or more plates touch are always unstable and will quickly resolve into two or more triple junctions. Such an area occurs currently west of New Guinea, where the Philippine, Australian, Eurasian and Pacific Plates meet . However, detailed mapping shows that this area may be divided into a series of micro plates and triple junctions and that the touching of more than three plates has only occurred transiently. Depending on the kinematics of plate boundaries, we can discern between a large number of different triple junctions. Using "R" (as in Ridge) for divergent plate boundaries, "T" (as in Trench) for convergent plate boundaries and "F" (as in Fault) for transform boundaries, we can describe RRR-, TTT-, FFF-, RTF- and a number of other triple junctions (McKenzie and Morgan 1969). Four examples are shown in Fig. 2.25. The examples shown there are all stable because the sum of the velocity vectors (U) is zero in all cases.

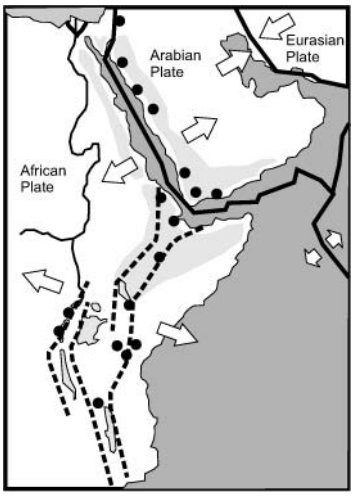


Figure 2.26. Schematic map of the most famous triple junction on earth: the meeting point of the East African rift system (dashed) with the Red Sea rift. The black dots are volcanoes, the light shaded areas show the rift flank uplift

Table 2.4. Different types of plate boundaries divided according to their kinematics and according to the plate type (O = oceanic lithosphere, C = continental lithosphere)

relative motion	type	plate tectonic feature	example
convergent	O-O	island arc	Philippines
	O-C	subduction zone trench	west coast of South-America
	C-C	collisional mountain belt	Himalaya
divergent	O-O	mid-oceanic ridge	Atlantic
	C-C	continental rift	east African rift
passive	O-C	passive plate margin	eastern Australia

One of the most famous triple junction is the RRR triple junction between the Gulf of Aden, the Red Sea and the Central African rift system (Fig. 2.26). With this, as with other RRR triple junctions, the three different branches occur in various stages of development and, in fact, one arm may stop rifting leaving behind a failed rift and a single plate boundary. A beautiful example of this is given by the failed Lambert Graben rift in Antarctica which stopped rifting in the Cretaceous while the ridge in the southern Indian Ocean continued to separate Australia from Antarctica.

2.4.4 The Wilson Cycle

The Wilson-cycle is a model that brings the individual processes that we discussed in the last sections into an imaginary cycle (Fig. 2.27). This cycle was first suggested by Wilson (1972) and supersedes the old terms of "geosynclines".

The Wilson-cycle begins with extension of a continent. This stage of the Wilson-cycle can be currently observed in the Rheingraben. In the second stage, the continent breaks up and a spreading center forms in its middle. Oceanic lithosphere begins to form. This is the stage of the Red Sea. The third stage involves the development of a classic ocean with passive continental margins on both sides and a mid-oceanic ridge in its center. This is currently being observed in the Atlantic. The following stages describe processes that occur during opposite relative plate motion. These stages may therefore occur independent of the first three stages. The fourth stage describes the commencement of subduction of an oceanic plate underneath a continental plate as it is currently observes along the west coast of South-America. The fifth stage involves the subduction of a mid-oceanic ridge underneath the continental plate. A present day example for this is the subduction of the

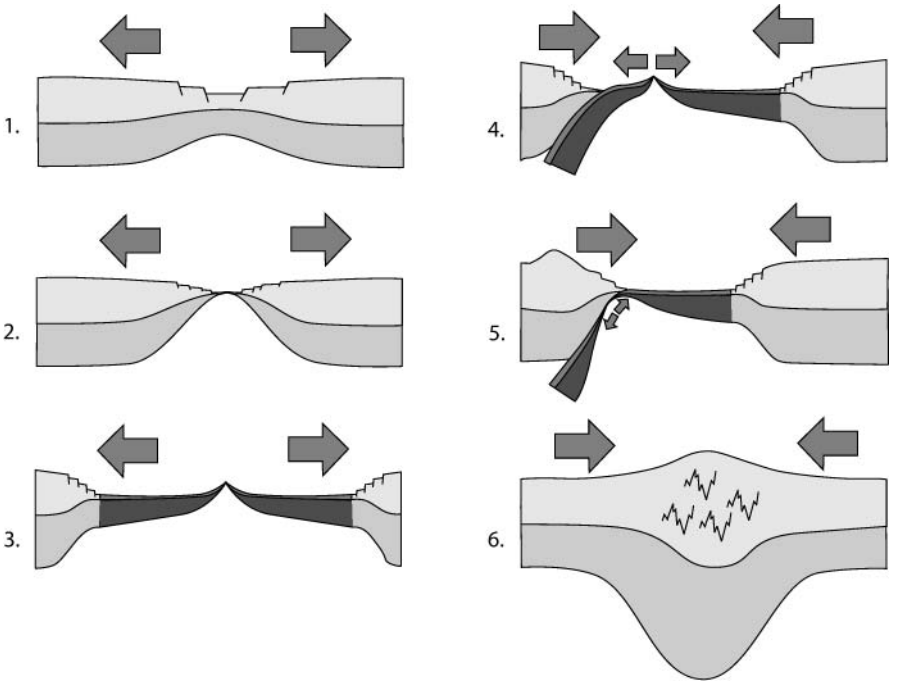


Figure 2.27. The Wilson cycle. The arrows indicate the relative plate motions

ridges bounding the Juan de Fuca Plate underneath North-America. During the sixth and final stage the collision between two continents occurs. Clearly, the India-Asia collision is the most spectacular example for that.

2.5 Problems

Problem 2.1. *Small circles and subduction zones (p. 22):*
Make a very rough estimate of the small circle radius of the Aleute arc (from Fig. 2.5 or any atlas) and estimate the subduction angle of the Pacific Plate underneath Alaska using the model of Fig. 2.6. This subduction angle is actually known. It is $\approx 45°$. Discuss possible reasons for the difference between your estimate and this value. Compare your estimate with an estimate of the subduction angle of the Indoaustralian Plate underneath the Eurasian Plate along the Java Trench.

Problem 2.2. *Understanding longitude and latitude (p. 23):*
Find a point on the surface of earth (outside the north pole!) where the following experiment is possible: You walk 1 km south then 1 km west and then 1 km north and you are where you started from. (It is said that this

question was asked by Sir Ernest Shackleton in 1908 to applicants for his south pole expedition.)

Problem 2.3. *Understanding longitude and latitude (p. 23):*
The city of Vienna is located at 16° east and 48° north; the city of Munich is located at 11° east and the same latitude as Vienna. How much earlier does the sun rise in Vienna?

Problem 2.4. *Understanding arc-minutes in decimals (p. 23):*
Improve your result of Problem 2.3 using the following values: Vienna: 16°21' east, 48°11' north; Munich: 11°39' east and the same latitude as Vienna.

Problem 2.5. *Estimating the curvature of earth (p. 24):*
a) How far is it from Vienna to Munich going directly from east to west along the surface of the globe? (use eq. 2.2). b) How far is it from Vienna to Munich going the shortest distance along a great circle? (use the formula for angular separation of 2 points in Table C.4). c) How far is it from Vienna to Munich if you could go along a tunnel connecting the two cities in a straight line through the earth? (use the result from b) and the logic illustrated in Fig. 2.6). Use the longitudes and latitudes given in Problem 2.3 and a perfectly spherical earth with radius R= 6 370 km.

Problem 2.6. *Spherical kinematics (p. 25):*
What is the west-east velocity of a plate that has rotation pole that coincides with the geographic north pole and rotates relative to another plate with an angular velocity of 10^{-15}s^{-1}; a) at the equator; b) at 48° north? (Use R= 6 370 km).

Problem 2.7. *Understanding torque (p. 25):*
A north-south striking mid-oceanic ridge has a west-directed force of 10^{12} N. At which latitude is the torque of this ridge around the axis of the earth $4 \cdot 10^{18}$ Nm? (The radius of earth is: $R = 6\,370$ km)

Problem 2.8. *The Mercator projection (p. 29):*
Derive the function that is used in the Mercator projection to convert longitude ϕ and latitude λ into the Cartesian coordinates x and y. Detailed study of Fig. 2.10 is helpful to answer this problem.

Problem 2.9. *Physics of the lithosphere (p. 37):*
Put numbers on the axes of the diagrams on Fig. 2.18. (This problem requires the use of parameter values that have not been discussed up to this chapter yet. It is meant as a self-test for the interested reader.)

Problem 2.10. *Modern and ancient plate boundaries (p. 40):*
a) Draw relative velocity vectors between the plates onto Fig. 2.20. b) Draw three future new plate boundaries and three plate boundaries that will disappear within the next 50 my on Fig. 2.20. c) Draw three ancient plate boundaries onto Fig. 2.20.

Problem 2.11. *Understanding triple junctions (p. 44):*
How many different types of triple junctions exist, using the three possible relative motions R, T and F? Not all of the triple junctions you should come up with are stable. Draw an example of a stable and an unstable triple junction.