CHAPTER ONE
Basins in their geodynamic environment

Summary

Sedimentary basins are regions of prolonged subsidence of the Earth's surface. The driving mechanisms of subsidence are related to processes originating within the relatively rigid, cooled thermal boundary layer of the Earth known as the lithosphere and from the flow of the mantle beneath. The lithosphere is composed of a number of plates that are in motion with respect to each other. Sedimentary basins therefore exist in a background environment of plate motion and mantle flow.

The Earth's interior is composed of a number of compositional and rheological zones. The main compositional zones are between crust, mantle and core, the crust containing relatively low-density rocks overlain by a discontinuous sedimentary cover. The mechanical and rheological divisions do not necessarily match the compositional zones. A fundamental rheological boundary is between the lithosphere and the underlying asthenosphere. The lithosphere is sufficiently rigid to comprise a number of relatively coherent plates. Its base is marked by a characteristic isotherm (c.1600 K) and is commonly termed the thermal lithosphere, which encloses a mechanical lithosphere. The upper portion of the thermal lithosphere is able to store elastic stresses over long time scales and is referred to as the elastic lithosphere. The continental lithosphere has a strength profile with depth that reflects its composition, temperature and water content. A weak, ductile zone exists in the lower crust below a brittle–ductile transition, but the strength of the underlying lithospheric mantle is uncertain. The oceanic lithosphere lacks this low-strength layer, its strength increasing with depth to the brittle–ductile transition in the upper mantle.

The relative motion of plates produces deformation, magmatism and seismicity concentrated along oceanic plate boundaries. Continental lithosphere is more complex, exhibiting seismicity and deformation far from plate boundaries, and with a heat flow and geotherm that is strongly influenced by radiogenic self-heating. Plate boundary forces and elevation contrasts strongly influence the state of stress of lithospheric plates.

Sedimentary basins have been classified principally in terms of the type of lithospheric substratum (i.e. continental, oceanic, transitional), their position with respect to the plate boundary (intracontinental, plate margin) and type of plate motion nearest to the basin (divergent, convergent, transform). The formative mechanisms of sedimentary basins fall into a small number of categories, although all mechanisms may operate during the evolution of a basin:

- Isostatic consequences of changes in crustal/lithospheric thickness, such as caused mechanically by lithospheric stretching, or purely thermally, as in the cooling of previously upwelled asthenosphere in regions of lithospheric stretching.
- Loading (and unloading) of the lithosphere causes a deflection or flexural deformation and therefore subsidence (and uplift), as in foreland basins.
- Viscous flow of the mantle causes non-permanent subsidence/uplift known as dynamic topography, which can most easily be recognised in the domal uplifts of the ocean floor at volcanic hotspots.

From the point of view of lithospheric processes there are two major groups of basins: (i) basins due to lithospheric stretching and subsequent cooling, belonging to the rift–drift suite; and (ii) basins formed primarily by flexure of continental and oceanic lithosphere.

1.1 Introduction and rationale

Maps of the global or plate-scale distribution of sediment thickness reveal strong variations (Fig. 1.1). It can be seen at both the global scale (Fig. 1.1) and the plate or continental scale (Fig. 1.2) that much of the area of the continental interiors is devoid of any sedimentary cover, with Precambrian crystalline rocks exposed at the surface. Elsewhere, the greatest sedimentary thicknesses are found in particular geological settings such as at extensional continental margins and fringing the world's great collisional mountain belts. These regions of large sedimentary thickness have undergone extensive and prolonged subsidence (Bally & Snelson 1980). The complexities of geological history have resulted in a patchwork of currently subsiding active basins and their ancient counterparts. Sedimentary basins, ancient and modern, are the primary archive of information on the evolution of the Earth over billions of years.

Basin Analysis: Principles and Application to Petroleum Play Assessment, Third Edition. Philip A. Allen and John R. Allen.

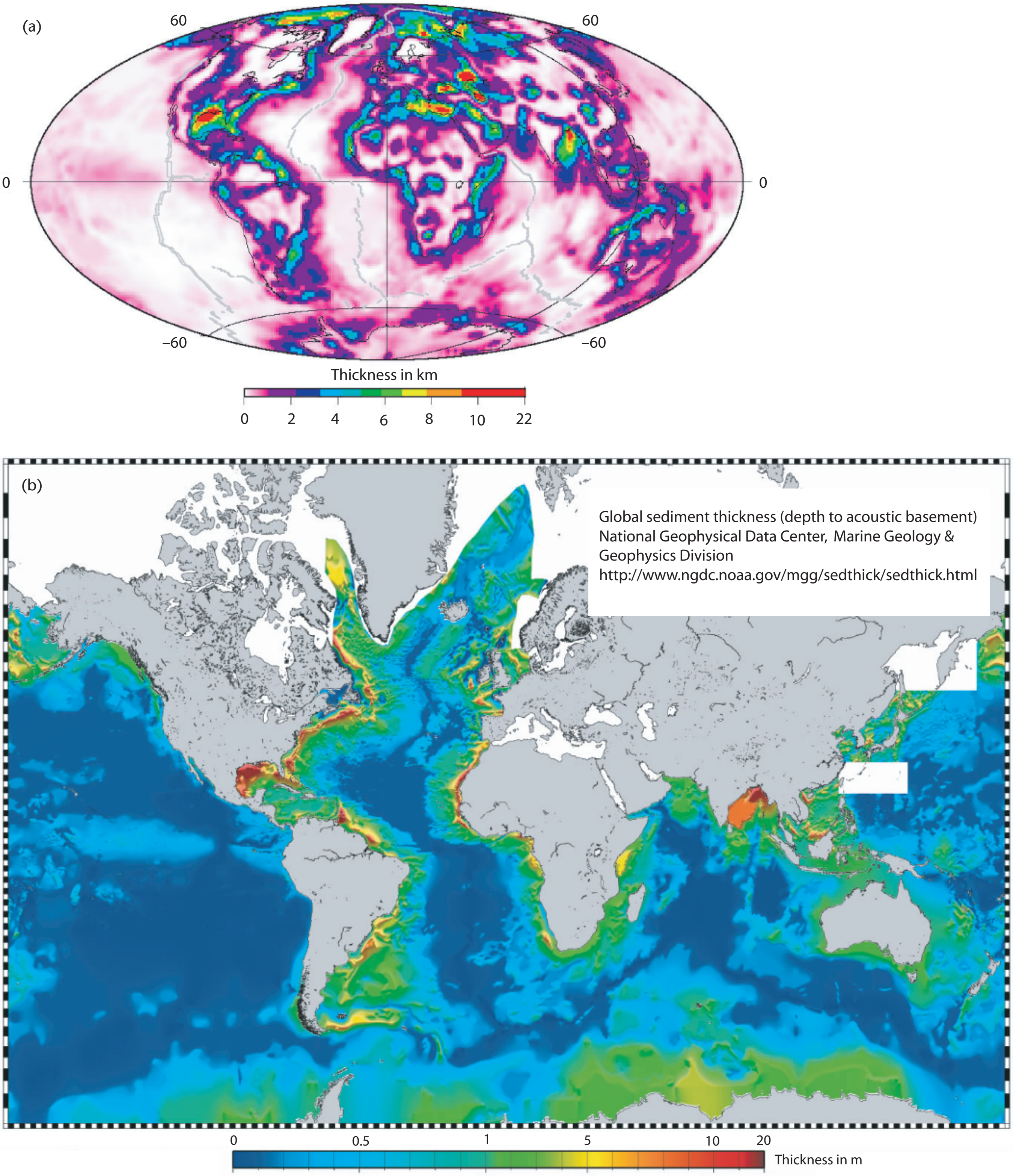

Fig. 1.1 (a) Global sediment thickness, from Laske & Masters (1997) based on the digital database of Gabi Laske at the University of San Diego, California (http://mahi.ucsd.edu/Gabi). (b) A higher-resolution map of the total sediment thickness in the ocean, from Divins (2003).

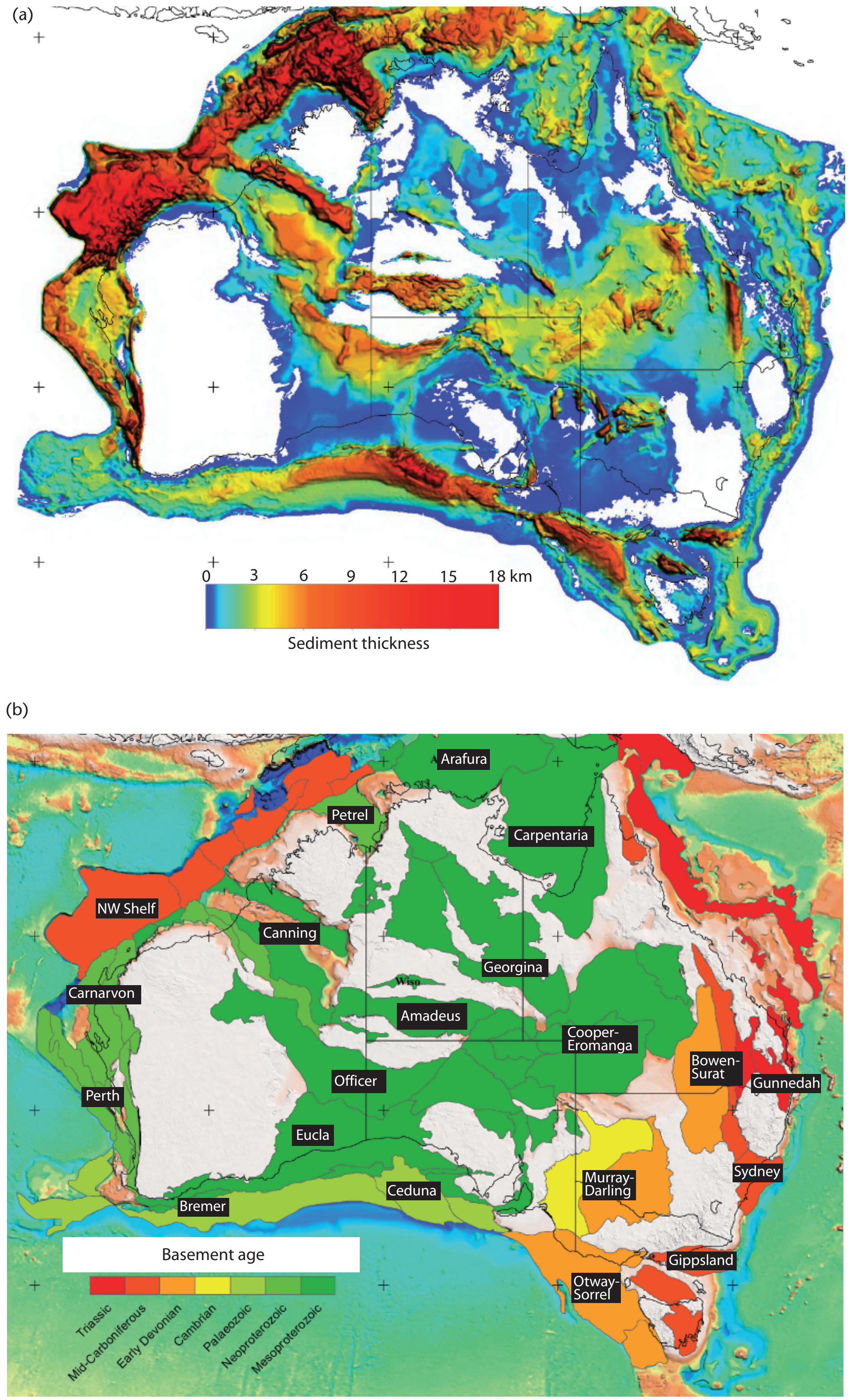

Fig. 1.2 (a) Sediment thicknesses, Australia. Note that the greatest sediment thicknesses are in continental margin basins superimposed on Paleozoic basement rocks. In addition, there are high sediment thicknesses in the continental interior related to structural inversion of intracontinental mountain belts (e.g. Petermann and Alice Springs orogenies) bordering the Officer and Georgina basins. This tectonic activity compartmentalised large cratonic basins. Note also that large areas of old cratons have no sedimentary cover. (b) Outlines of ancient and modern sedimentary basins shown according to basement age. Note that the bulk of the basement ages are Precambrian (green). Younger basement rocks underlie basins of the continental margins (e.g. NW Shelf, Ceduna Basin in south), and the terranes east of the Tasman Line. Both (a) and (b) fromFrOG Tech Pty Ltd. (2005).

The location of sedimentary basins and their driving mechanisms are intimately associated with the motion of discrete, relatively rigid slabs, which together represent the cooled thermal boundary layer of the Earth. The outer shell of the Earth comprises a relatively small number of these thin, relatively rigid plates, which are in a state of motion with respect to each other. Such motions set up plate boundary forces that may be transferred considerable distances into the interior of the plates, so that sedimentary basins exist in a background environment of stress set up by plate motion.

The lithospheric plates are the surface manifestation of a slow thermal convection in the mantle, and are subject to differential thermal stresses along their bases. The mantle and lithosphere therefore do not operate as independent systems. We see spectacular evidence for the interaction of mantle processes and the lithosphere in the volcanic and topographic expression above mantle flow structures, some of which may have risen from the core–mantle boundary. We also discern, though less spectacularly, the effects on mantle flow caused by the subduction of cold slabs of oceanic crust at ocean–continent boundaries.

Deep Earth processes involving the thermomechanical behaviour of the lithosphere and the flow of the underlying mantle are coupled to Earth surface processes of erosion, sediment and solute transport and deposition in sedimentary basins. This coupling between 'deep' and 'surface' is the fundamental basis for the practice of broad, integrative thinking in basin analysis, and underpins the understanding of sedimentary basins as geodynamical entities. It is also the framework for the study of petroleum systems in sedimentary basins (Fig. 1.3). The connectedness between deep and surface geodynamics is emphasised throughout this text, and the fruits of an improved understanding derived by studying such connections are illustrated in the application to the exploration of hydrocarbons in Chapters 11 and 12.

Two key, dovetailed concepts therefore underlie this necessity of integration in basin analysis:

1. The dynamics of the solid Earth results in tectonic processes at various scales that control the generation of space in which sediment may accumulate for long periods of time. Tectonic processes determine the bulk strain and strain rate at which the basin and its structures form, and also control thermal history, magmatism and burial history. At a smaller scale, tectonic processes generate fault arrays that provide the template for spatial patterns of uplift and subsidence, and determine the pathways of sediment transport from source to sink.
2. Sediment is derived by weathering and erosion and released into dispersal systems *en route* to long-term depositional sinks (Allen 2008). The 'erosional engine' of the sediment routing system is powered by tectonic processes causing uplift of rocks, and the deposition of sediment to form stratigraphy is strongly modulated by the spatial patterns and rates of tectonically generated subsidence. The functioning of sediment routing systems determines the gross depositional environments in a basin and its stratigraphic geometries and has important feedbacks into tectonic deformation, structural reactivation and exhumation.

1.2 Compositional zonation of the Earth

There are three main compositional units: the crust, mantle and core (Fig. 1.4).

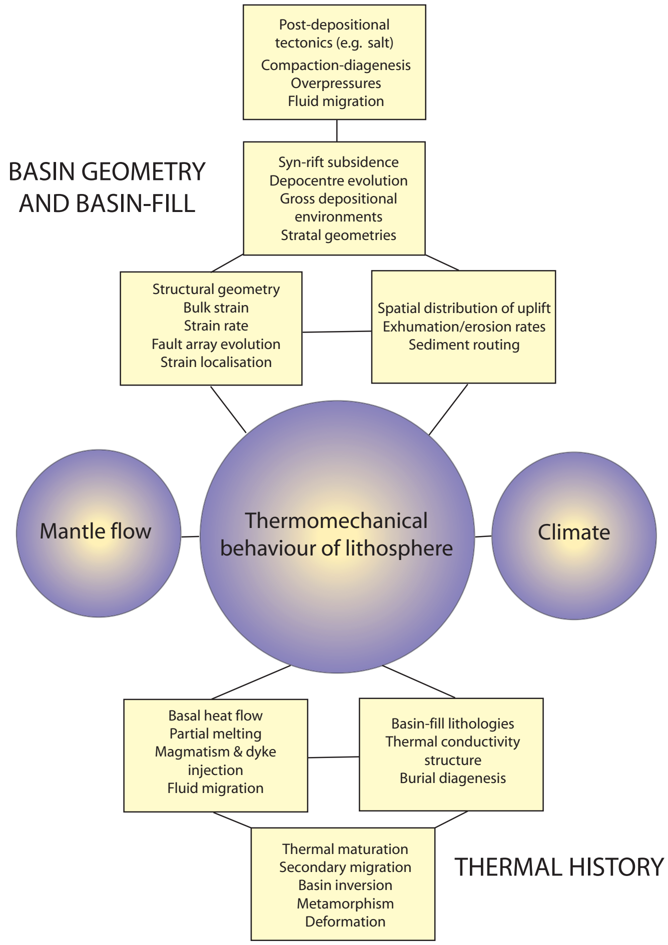

Fig. 1.3 The basis for integration in basin analysis is the coupling between deep and surface geodynamics, the effects of which cascade through the various aspects of basin analysis to the application to petroleum systems.

1.2.1 Oceanic crust

The crust is an outer shell of relatively low-density rocks. The oceanic crust is thin, ranging from approximately 4–20 km in thickness, 6–10 km being 'normal', and with an average density of about 2900 kg m^{-3}. It comprises a number of layers that reflect its mode of creation: an upper veneer (layer 1) of unconsolidated or poorly consolidated sediments, generally up to 0.5 km thick; an intermediate layer 2 of basaltic composition, consisting of pillow lavas and associated products of submarine eruptions; and a layer 3 of gabbros and peridotites that comprise the parent rocks, which upon differentiation give rise to the basalts of layer 2. The oceanic crust was formerly thought to be distinctly layered in terms of velocity of seismic waves, but more recent views are that it possesses a more gradual and continuous increase in velocity with depth.

Ocean crust is being created as new oceanic seafloor at 2.8 $km^2\,yr^{-1}$, and the current area of ocean crust is $3.2 \times 10^8\,km^2$, roughly 60% of the surface of the Earth. At this rate of production, the area of ocean crust would double in a further 100 Myr. Yet the lifetime of ocean crust is short – the oldest oceanic crust in today's oceans is as young

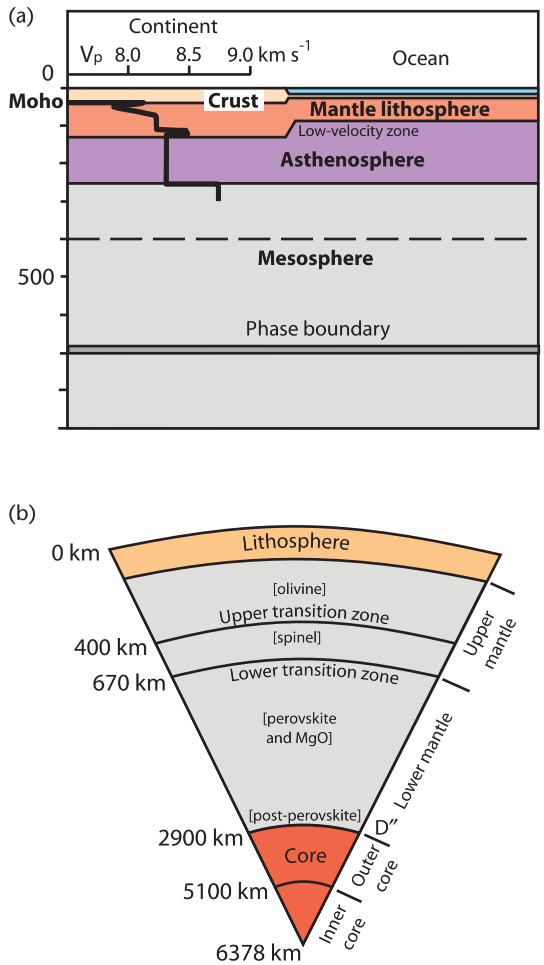

Fig. 1.4 The main compositional (a) and rheological (b) boundaries of the Earth. The most important compositional boundaries are between crust, mantle and core. There are also compositional variations within the continental crust and compositional variations caused by phase changes in the mantle. V_p is the velocity of P-waves. The main rheological boundary is between the lithosphere and the asthenosphere. P-wave velocities increase markedly beneath the Moho, but decrease in a low-velocity zone in the asthenosphere. The lithosphere is rigid enough to act as a coherent plate. The perovskite to post-perovskite transition at >2600 km is associated with the D″ seismic discontinuities.

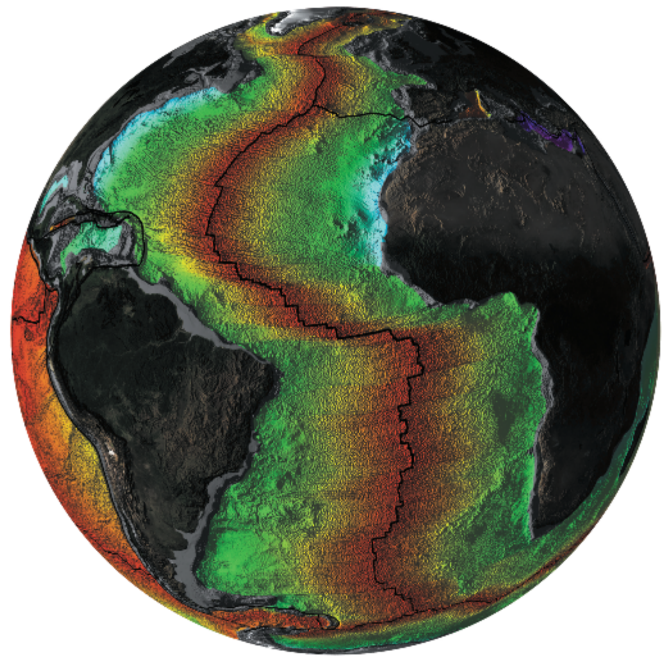

Fig. 1.5 Age of the ocean crust in the Atlantic Ocean. Colour code for the age of the oceanic crust is red for zero at the mid-ocean ridge and blue for Jurassic. Reproduced with permission from Mueller *et al*. (2008). Note that the age of the ocean floor closely mimics the depth of the ocean floor below sea level (see Chapter 2).

as Jurassic in age (c.150 Ma) (Fig. 1.5) and the average age of oceanic crust is about 100 Ma. This is because the oceanic crust caps the gravitationally unstable oceanic lithosphere; as a result, it is consumed by subduction.

1.2.2 Continental crust

The continental crust is thicker, ranging from 30 to 70 km, but with an average thickness of 35–40 km (Table 1.1). It was originally thought to be divided into two layers, each with a distinct composition and density: (i) an upper layer with physical properties similar to those of granites, granodiorites or diorites, overlain by a thin,

Table 1.1 Physical properties of the Earth and its compositional zones

Properties of the Earth	
Equatorial radius	6.378137×10^6 m
Volume	1.0832×10^{21} m^3
Mass	$5.9736\ 10^{24}$ kg
Mean density	5.515×10^3 kg m^{-3}
Surface area	5.10×10^{14} m^2
Land area	1.48×10^{14} m^2
Mean land elevation	875 m
Continental area (incl. continental shelves)	2.0×10^{14} m^2
Water area	3.62×10^{14} m^2
Mean ocean depth	3794 m
Mean surface heat flow	87 mW m^{-2}
Age of Earth	4.55 Ga
Crust	
Mass	2.36×10^{22} kg
Mean thickness of continental crust	40 km
Mean thickness of oceanic crust	6 km
Mean continental heat flow	65 mW m^{-2}
Mean oceanic heat flow	101 mW m^{-2}
Zero-pressure density (granodiorite)	2700 kg m^{-3}
Zero-pressure density (basalt and gabbro)	2950 kg m^{-3}
Mantle	
Volume	9.06×10^{20} m^{-3}
Mass	4.043×10^{24} kg
Zero-pressure density (peridotite)	3250 kg m^{-3}
Core	
Volume	1.77×10^{20} m^3
Mass	1.883×10^{24} kg

discontinuous veneer of sedimentary rocks. This so-called 'granitic layer' has a thickness of between 20–25 km and a density of 2500–2700 kg m^{-3}. The term 'granitic' is, however, misleading, since average densities are greater than that of granite. (ii) A lower layer of primarily basaltic composition, but the pressure and temperature at depths in excess of 25 km imply that the rocks are granulites, or their high-pressure, high-temperature equivalents, eclogites or amphibolites. The density of this lower layer is 2800–3100 kg m^{-3}. These layers may not in reality be well defined, and instead a more continuous variation of composition with depth may exist.

Information on the density of crustal rocks has been obtained largely by observations on seismograms of the speed of seismic waves passing through the various layers, combined with laboratory experiments on rock materials. The existence of a low-velocity crust was discovered by the geophysicist Mohorovicic. At the crust–mantle boundary, seismic **P** (longitudinal) wave velocities increase markedly; this abrupt increase in velocity may reflect a corresponding increase in rock density (Fig. 1.4). This horizon is known as the Mohorovicic discontinuity or Moho. The Moho varies in depth considerably. The continental crust thickens under orogenic belts, thins under zones of rifting and attenuates completely at continental margins (Fig. 1.6).

In some regions, particularly the attenuated margins of continents, the crust is intermediate in character and thickness between typical oceanic and continental varieties. This may be due to the injection of dense intrusions, to metamorphism, or to other processes accompanying stretching. In particular, the depth to the Moho may be abnormally great due to the igneous underplating of the crust associated with high temperatures in the asthenosphere. The presence of several kilometres of underplate emplaced in the Early Tertiary over the head of the Iceland plume has been interpreted from seismic experiments carried out on the northwest European continental margin.

Since the upper part of the continental crust is thick and gravitationally stable, it is not subducted, and contributes to the much greater average age of the continental crust (10^9 years). However, the denser lower continental crust and the underlying lithospheric mantle may undergo recycling into the Earth's interior (see Chapter 2).

1.2.3 Mantle

The *mantle* is divided into two layers: the upper and lower mantle. The upper mantle extends to about 680 km ± 20 km and is punctuated by phase transitions. The lower mantle extends to the outer limit of the core at 2900 km, with an increasing density with depth.

Although there are of course no *in situ* measurements of the composition of the mantle, it can be estimated from the chemistry of volcanic and intrusive rocks derived by melting of the mantle, from tectonically emplaced slivers of mantle rock preserved in orogenic belts known as ophiolites, from nodules preserved in volcanic rocks, from minerals brought to the surface explosively in kimberlites, and, importantly, from the remote but sophisticated measurement of the mantle using seismic waves. The main constituent of the mantle is thought to be olivine, mostly the Mg-rich variety forsterite.

Olivine is known to undergo phase changes to denser structures at pressures equivalent to depths of 390–450 km and c.700 km in the Earth. At 390–450 km olivine is thought to change to spinel via an exothermic reaction involving a 10% increase in density. At c.700 km spinel changes to perovskite (magnesium silicate) and magnesium oxide in an endothermic reaction. These phase changes can be recognised by variations in the velocity of **S**-waves (McKenzie 1983) and may determine the scale of convection in the mantle (Silver *et al*. 1988). The high-pressure form *post-perovskite* is stable at temperatures above 2500 K and pressures above 120 GPa, corresponding to a depth in excess of 2600 km in the innermost mantle. The perovskite to post-perovskite transition may be responsible for the seismic discontinuity known as D″ (Peltier 2007) (Fig. 1.4).

1.3 Rheological zonation of the Earth

The mechanical or rheological divisions of the interior of the Earth do not necessarily match the compositional zones. One of the rheological boundaries of primary interest to students of basin analysis is the differentiation between the lithosphere and the asthenosphere. This is because the vertical motions (subsidence, uplift) in sedimentary basins are principally a response to the deformation of this uppermost rheological zone of the Earth, or to the guiding through the lithosphere of stresses transmitted from the mantle.

1.3.1 Lithosphere

The *lithosphere* is the rigid outer shell of the Earth, comprising the crust and the upper part of the mantle. It is of particular importance to note the difference between the *thermal* and *elastic* thicknesses of the lithosphere. It is generally believed (e.g. Parsons & Sclater 1977; Pollack & Chapman 1977) that the base of the lithosphere is represented by a characteristic isotherm (1100–1330 °C, or 1600 K) at which mantle rocks approach their solidus temperature. The solidus temperature of peridotite in K is $1500 + 0.12p$, where the pressure p is in MPa. This defines the *thermal lithosphere* (Figs 1.7, 1.8). Heat flow in the thermal lithosphere is by conduction, whereas heat flow in the underlying asthenosphere is by convection. The temperature gradient in the convecting mantle is the adiabatic gradient, which is approximately 0.5 K km^{-1} in the shallow upper mantle.

Typical thicknesses of lithosphere under the oceans vary from c.5 km at mid-ocean ridges to c.100 km in the coolest parts of the oceans. The lower boundary of the lithosphere is poorly defined under continents, depths of 100 km to 250 km being typical. However, the precise seismological mapping of the base of the thermal lithosphere is difficult since it does not correspond to a compositional boundary. Instead, it corresponds to a rheological change in peridotite, which at about 1600 K becomes sufficiently weak to convect. A thermal boundary layer separates the convecting asthenosphere from the mechanically strong lithosphere above. Stepwise increases in velocities of **S** and **P** waves with depth through the lithosphere suggest that it also contains minor compositional boundaries within it.

It is puzzling that continental subcrustal lithosphere can extend to great depths of >200 km, since it should become gravitationally unstable relative to the underlying hot asthenosphere (see Chapter 2). This thick subcrustal lithosphere may avoid delamination by being depleted by the extraction of melts, resulting in harzburgites. Harzburgites, the Mg-rich refractory residuum of peridotite after the extraction of basaltic melts, are not only less dense, but also have a higher (~500 K) melting temperature than unmodified peridotite, allowing these cratons to retain both their long-term buoyancy and their thickness. The main period of growth of cratonic lithosphere was in the Archaean, but since then there have been cycles of enhanced

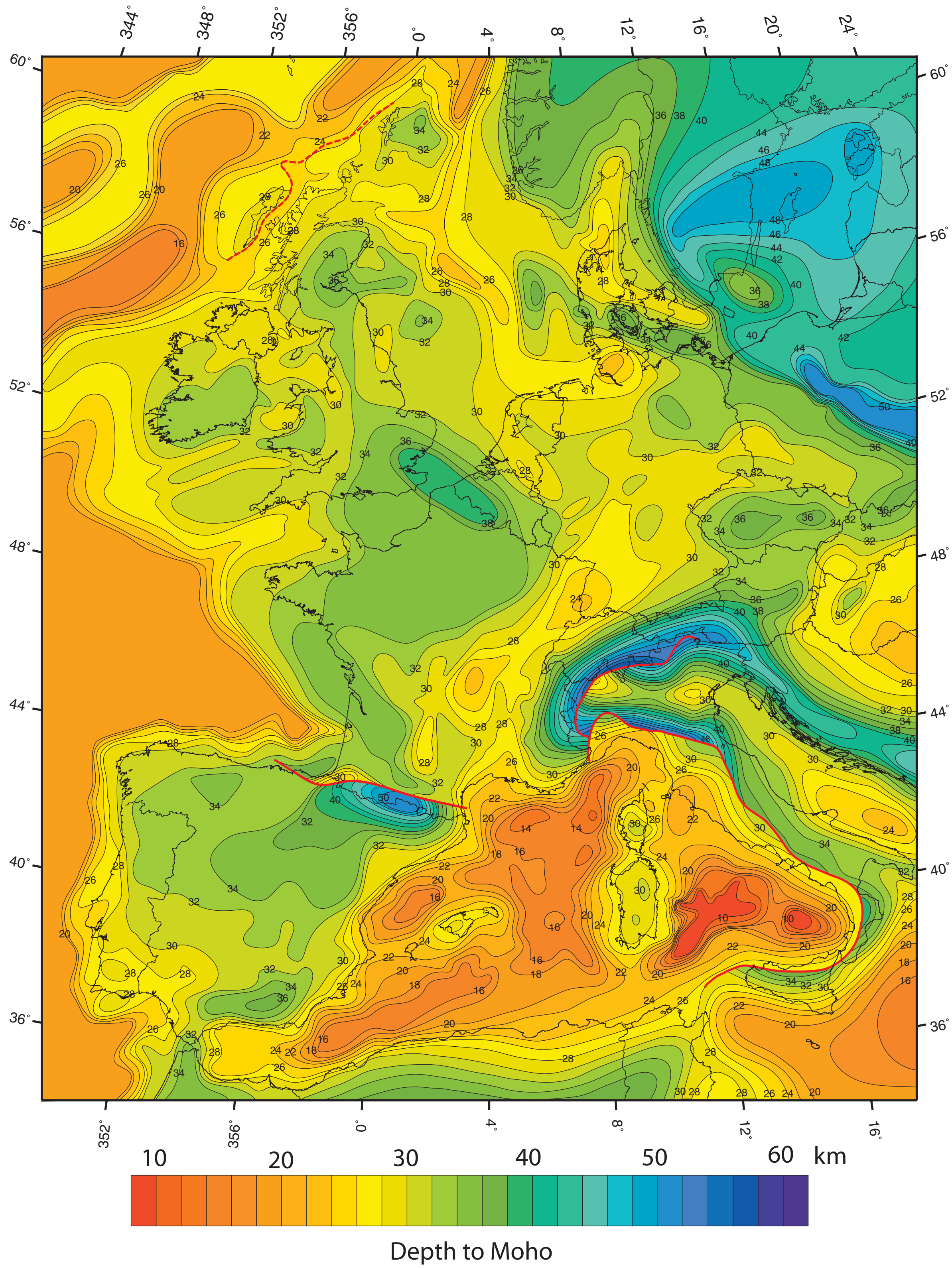

Fig. 1.6 Depth to the Moho, western Europe, from Dèzes & Ziegler (2002). Note that the Moho is deep under orogenic belts where the crust has been thickened tectonically, such as the Pyrenees (Spain), Alps (France, Switzerland, Austria), Dinarides (Croatia) and Apennines (Italy) and under the cratonic Scandinavian shield. The crust is thin in the basins floored with oceanic crust in the eastern Mediterranean and the Atlantic Ocean. The crust is also thin in regions of stretching of the continental lithosphere, as can be seen particularly in the North Sea region, and in the Rhine-Bresse-Rhone region of the West European Rift system. Free download from EUCOR-URGENT, http//comp1.geol.unibas.ch/downloads/Moho_net/euromoho1_3.pdf.

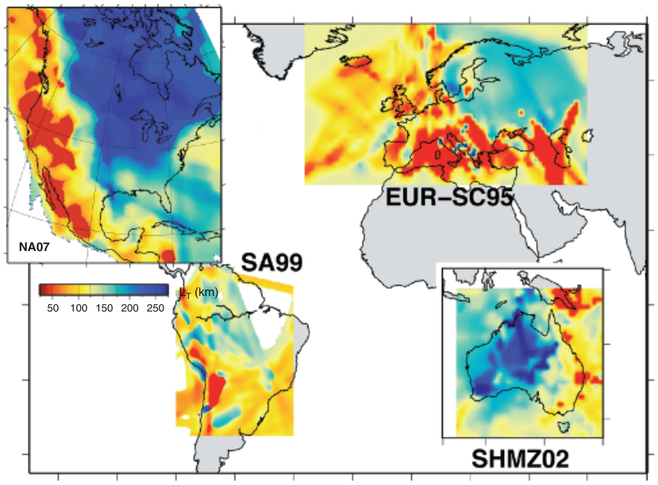

Fig. 1.7 Thickness of the thermal lithosphere (1200 °C isotherm) from a conversion of the **S**-wave velocity model of Bedle & van der Lee (2009) to temperature. The temperature conversion follows Goes and van der Lee (2002) and assumes a constant peridotitic composition for the mantle lithosphere. Note (especially for North America and Europe) that there is a thick, cold core of the plate, shown by blue colours, up to 250 km thick, surrounded by relatively thin, hot lithosphere shown by red colours (≪100 km thick), such as the Basin and Range province of west-central USA, and the Mediterranean region of southern Europe.

magmatism, judged from the frequency of age of detrital zircons (Condie 1998; Hawkesworth *et al.* 2009).

The rigidity of the lithosphere allows it to behave as a coherent plate, but only the upper part of the lithosphere is sufficiently rigid to retain elastic stresses over geological time scales (say 10^7 years). Below this upper *elastic lithosphere* creep processes efficiently relax elastic stresses, so there is a physical and conceptual difference between the elastic lithosphere and the thermal lithosphere. The lithosphere below the upper elastic portion must therefore be sufficiently weak to relax elastic stresses but sufficiently rigid to remain a coherent part of the surface plate.

The lithospheric plates can be easily deformed by bending about a horizontal axis, but are highly resistant to torsion about steeply inclined axes. This latter property of strength allows the motion of plates over the Earth's surface to be modelled, assuming no internal deformation, except at plate boundaries. But how do the oceanic and continental lithosphere compare in terms of their long-term (>1 Myr) flexural strength? Different views exist on this problem (Jackson 2002; Burov & Watts 2006).

On the one hand, oceanic plates are stronger because they consist of more mafic mineral assemblages and contain fewer intrinsic weaknesses such as old fault systems, but on the other hand are thinner and hotter and therefore bend more easily than continental plates. The strongest part of the oceanic lithosphere occurs in the mantle between 20 and 60 km depth, below which it becomes increasingly ductile (Fig. 1.9). In contrast, the continents contain quartz, which shows ductile flow at lower temperatures than olivine, and contain heterogeneities, but are thicker and cooler than oceanic plates. The strength profile is potentially complex. In one model (jelly sandwich) both the upper crust and mantle are strong, separated by a mid-lower crustal aseismic ductile zone that has been invoked as a level of detachment of major upper crustal faults (e.g. Kusznir & Park 1987). In another model (crème brûlée) the elastic strength is confined to the crust. Consequently, major earthquakes are confined to this brittle crustal layer (Jackson 2005) (Figs 1.9, 1.10).

There are heterogeneities in the mantle part of the lithosphere, although they are small compared to the crust. Seismological studies of western Europe suggest a highly stratified lithosphere beneath the Moho. In particular, a 'channel' of reduced **P**-wave velocities has been interpreted between 10 and 20 km below the Moho. This 10 km-thick layer cannot be explained in terms of partial melting since the solidus temperature is far in excess of the actual temperature; the hydration (serpentinisation) of peridotites has been postulated as a possible mechanism. Whatever the cause, this upper low-velocity channel may serve as a zone of decoupling of the upper lithosphere from the lower portion of the lithosphere when acted upon by tangential tectonic forces. There are few examples, however, where a process of decoupling can be unambiguously demonstrated at these levels.

1.3.2 Sub-lithospheric mantle

The underlying region, the *asthenosphere*, is weaker than the lithosphere and is able to undergo deformation relatively easily by flow. The upper part of the asthenosphere is known as the low-velocity zone, where **P** and **S**-wave transmission speeds drop markedly, presumably due to partial melting.

Studies of minute variations in the transmission speeds of seismic waves have allowed the structure of the deep mantle to be visualised and mapped (Nolet 2008), a topic known as *seismic tomography*. Zones of faster than average seismic velocity are attributed to propagation through denser rock, which in turn is most likely due to a cooler temperature. Zones with slower than average seismic velocity are likewise thought to be due to warmer temperatures. Variations in temperature are probably caused by large-scale convection.

Far from being inert, the mantle represents a vast volume of rock that dynamically interacts with the lithosphere. Instabilities rise from the core–mantle boundary as *plumes* of hot material that impinge on the base of the overlying lithosphere and may have a major role in continental break-up (Burke & Dewey 1973; White & McKenzie 1988). The thermal effects of the subduction of cold oceanic lithosphere and the insulating effects of supercontinental assemblies are also thought to be recognisable in the thermal structure of the mantle (Gurnis *et al.* 1996; O'Neill *et al.* 2009) (Chapter 5).

1.4 Geodynamic background

1.4.1 Plate tectonics, seismicity and deformation

Plate tectonics is a kinematic theory that describes the motion of the lithosphere as comprising a relatively small number of rigid plates that deform at their boundaries. These plate boundary regions are very narrow in comparison with the size of individual plates. In the oceans, earthquakes are strongly concentrated in narrow bands corresponding to mid-ocean ridges, subduction zones and transform fault zones (Barazangi & Dorman 1969) (Fig. 1.11). Elsewhere, the oceanic crust is essentially aseismic. Consequently, the oceans behave as described by plate tectonic theory.

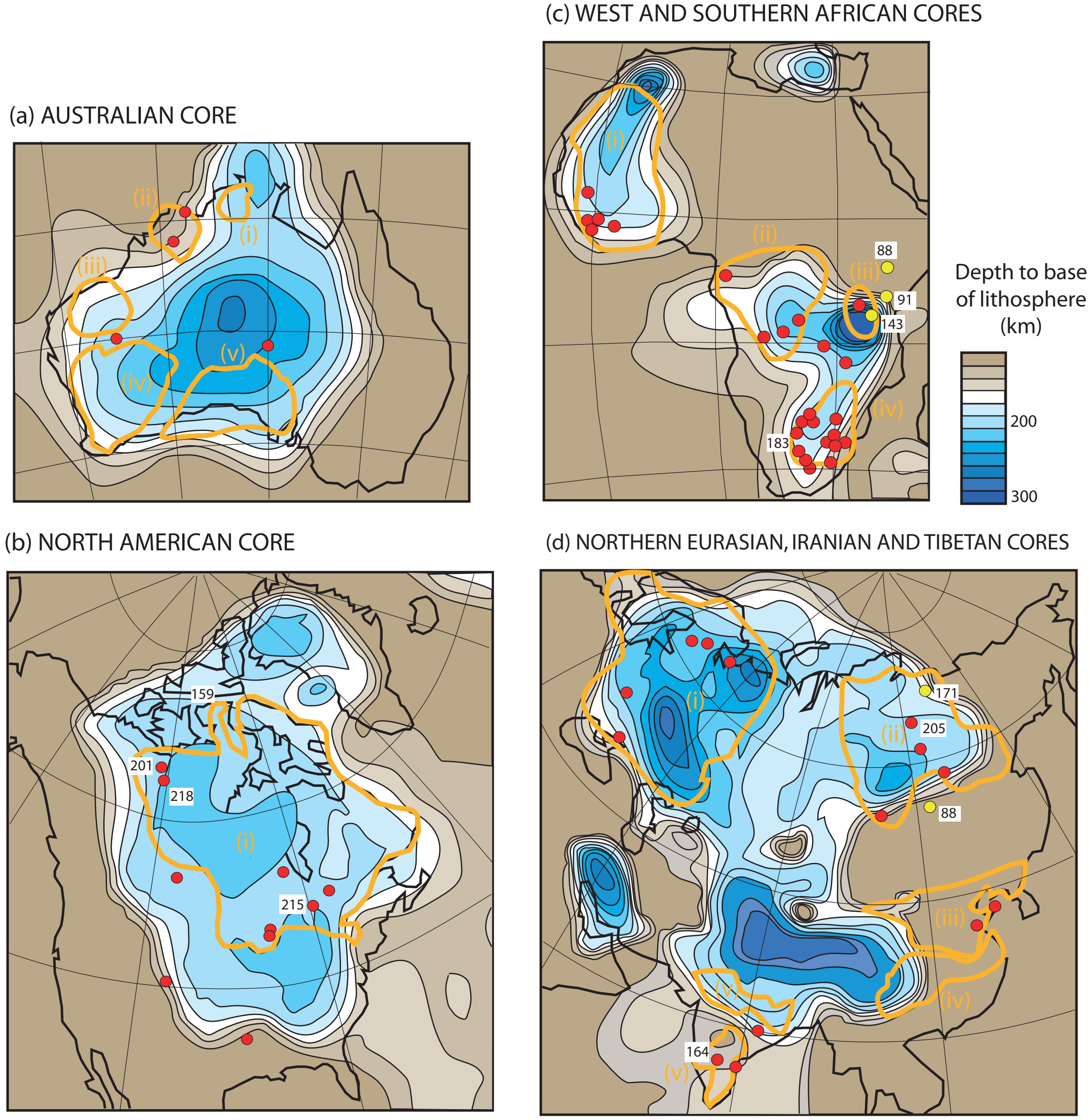

Fig. 1.8 Lithospheric thickness derived from **S**-wave velocities (Preistley & McKenzie 2006). The yellow circles show the locations of kimberlites and alkali basalts lacking diamonds, with white boxes showing lithospheric thickness derived from nodule mineralogies. The red circles show the locations of kimberlites bearing diamonds, with the lithospheric thickness where indicated. The yellow lines show the boundaries of the cratons. (a) Australia: i, Darwin craton; ii, Kimberley craton; iii, Pilbara block; iv, Yilgarn block; v, Gawler craton. (b) North America: i; North American craton. (c) Africa: i, West African craton; ii, Angolan craton; iii, Tanzanian craton; iv, Kalahari craton. (d) Eurasia: i, East European craton; ii, Siberian craton; iii, North China craton; iv, Yangtze Block of South China craton; v, Himalayan cratons; vi, Dharwar craton. Diamond occurrences show that the lithospheric thickness beneath cratons exceeds 250 km. Reprinted from Preistley & McKenzie (2006), with permission from Elsevier.

In the continents, however, earthquakes are very widely distributed (Fig. 1.11). Geodetic measurements, principally from global positioning system (GPS) networks, confirm that active deformation is taking place over very extensive areas rather than being confined to plate boundary zones (England 1987). Continental plates clearly undergo deformation a long way from plate boundaries, in the form of intracontinental orogenies causing structural reactivation and basin development in continental interiors. In such cases, the driving force may be the excess potential energy of elevated continental crust or the far-field transmission of in-plane stresses from a plate boundary. Plate tectonics theory is therefore a far less adequate description of the deformation of the lithosphere in continental regions. Instead, the continents may be better approximated by thin viscous sheets that deform in a ductile fashion; at a large scale, the deformation is continuous rather than discrete. This view of a diffuse, continuous deformation of continental lithosphere is supported by geodetic measurements of regions such as the Aegean Sea, eastern Mediterranean. Whatever the best descriptions for deformation in the oceanic and continental lithosphere, the nature and rates of relative plate motion govern many aspects of the geodynamic environment of basins.

The fact that earthquake epicentres occur at depths as great at 650 to 700 km along some plate boundaries suggests that a process exists that is capable of transferring brittle material to depths normally associated with deformation by flow. This process of plate subduction is responsible for both the relative youth of the oceanic crust and the distribution of earthquake epicentres.

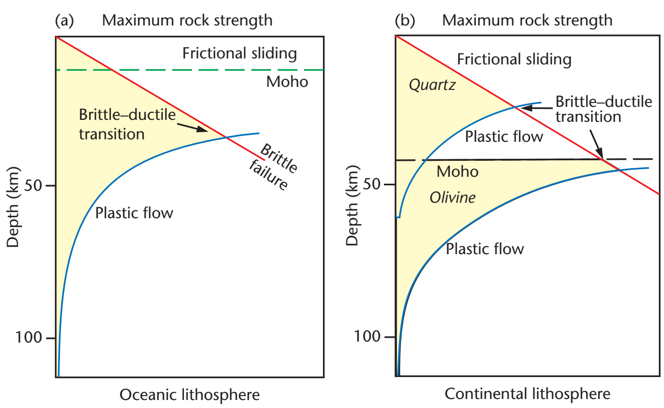

Fig. 1.9 Strength profiles of the continental and oceanic lithosphere. Strength also depends on whether the lithosphere is in tension or compression, and on the presence of volatiles such as water (Jackson 2005). The model for the continental lithosphere is called 'jelly sandwich', but an alternative view for the continental lithosphere is that the upper mantle is not strong and is aseismic, which is termed 'crème brûlée' (compare Burov & Watts 2006 and Jackson 2002).

1.4.2 The geoid

The Earth's outer fluid envelope, the oceans, flows to a level that is determined by Earth's gravitational potential field, which is controlled by the distribution of density with depth in the solid Earth. The water of the oceans accumulates to a height that is not a perfect reflection of the oblate spheroid shape of the Earth. Instead, mean sea level, representing the observed geoid height, shows very large undulation-like departures from the idealised shape (Fig. 1.12). These departures, or geoid anomalies, may result from a complicated set of contributing factors, but the fact that they vary significantly from one plate to another suggests that there must be an important long-wavelength sub-lithospheric contribution. This sub-lithospheric source of geoid anomalies presumably derives from the density heterogeneity of the mantle, which may itself be due to the history of convective flow and slab subduction.

The equivalence in terms of gravitational potential of different sections of oceanic and continental lithosphere is an important question. For example, a 125 km-thick continental lithosphere with an average density of 2750 kg m^{-3} should produce a (+6 m) geoid anomaly, consistent with an isostatically compensated continental lithospheric column with a surface elevation of 1 km. Such a column is also likely to be in potential energy balance with a mid-ocean ridge with a water depth of 2.5 km. Bearing in mind the mean elevations of the continents and ocean basins (Table 1.1 and §1.4.3), this suggests that continents achieve a mean elevation that is balanced by the adjacent ocean basins. The isostatic balance between adjacent columns of oceanic and continental lithosphere, and the horizontal

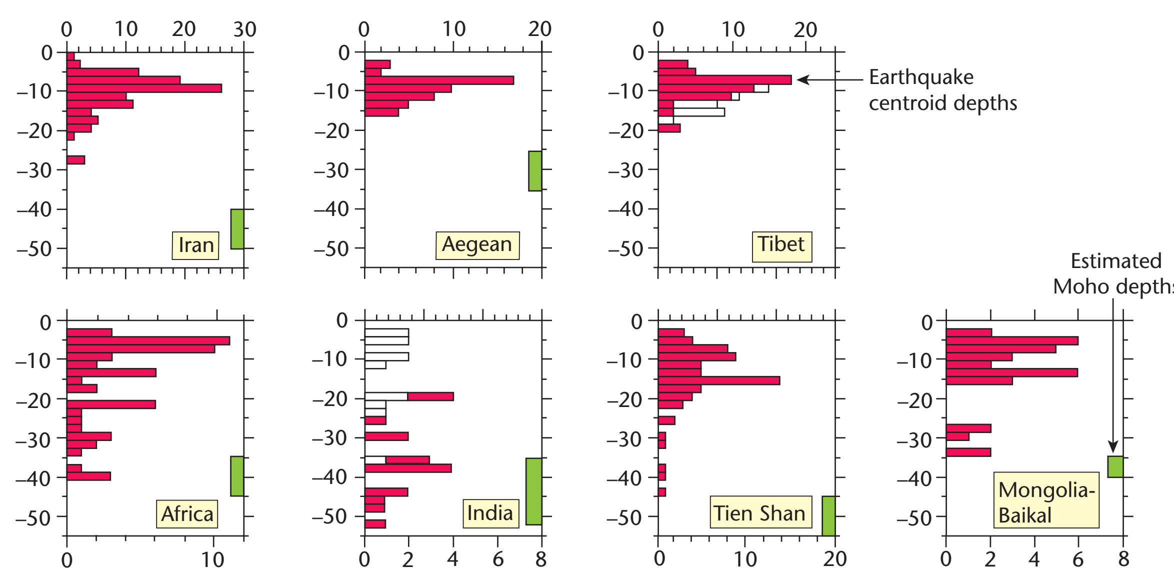

Fig. 1.10 Depth of occurrence of earthquakes in several regions of the world. In the top row, the distribution of earthquakes shows that they must originate only from the upper part of the crust, at depths of less than 20 km. In the bottom row, earthquakes occur over a larger depth range of up to 40–50 km. However, comparison with the estimates of Moho depth suggests that even these deep earthquakes may originate from the crust. In other words, there is little indication that the mantle is seismogenic. Consequently, the mantle may contribute little to the elastic strength of the continental plates (Jackson *et al*. 2008).

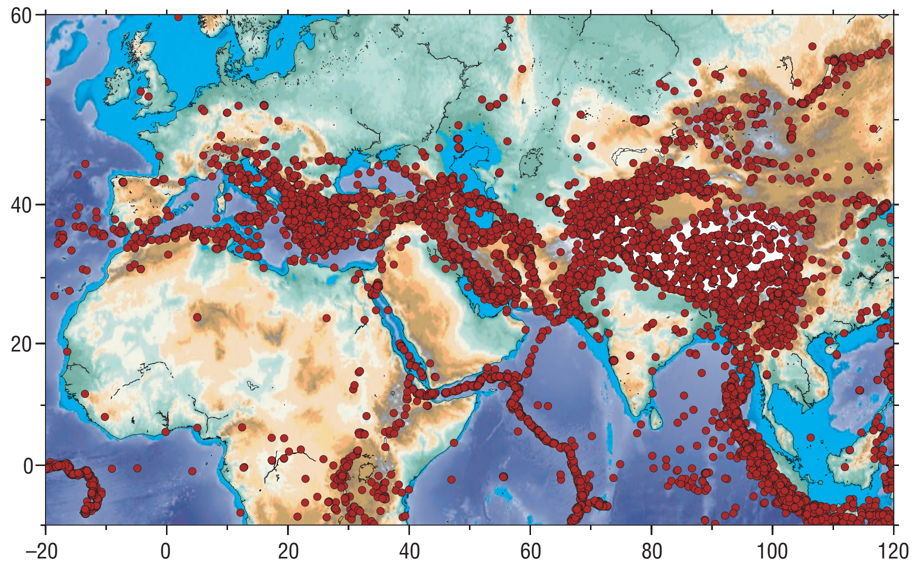

Fig. 1.11 Earthquake epicentres, shown in red circles, are distributed over a very wide mountainous region, whereas they are concentrated in narrow zones in the oceans. From Jackson (2005).

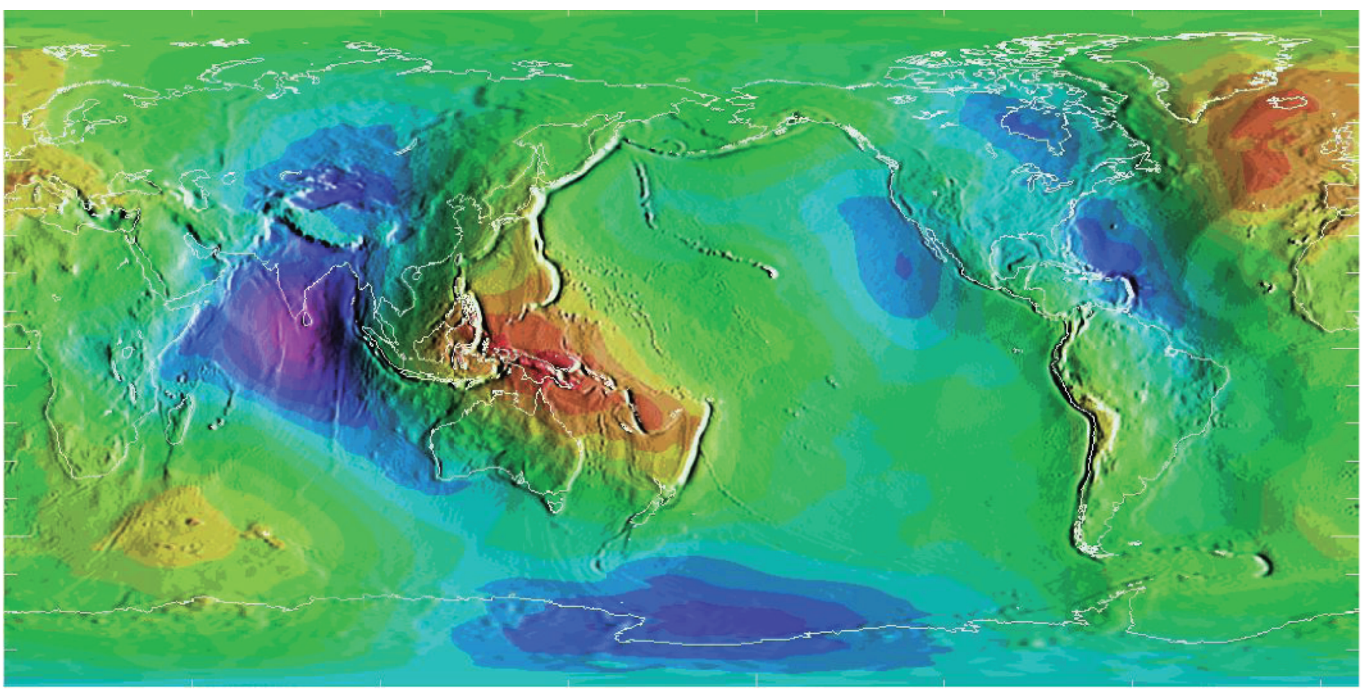

Fig. 1.12 The geoid height anomaly: the geoid relative to a reference ellipsoid, showing two major geoid highs and lows per circumference of the Earth. One geoid high is situated across much of Africa, western Eurasia and the North Atlantic. Another occupies the western Pacific area. These long-wavelength geoid height anomalies may be almost permanent features of the mantle (Torsvik *et al*. 2008).

forces between them, is given in more detail in Chapter 2 and in Appendices 2, 3 and 4.

1.4.3 Topography and isostasy

The mean elevation of the land is just 875 m, whereas the mean depth of the ocean basins is 3794 m (Table 1.1). These mean elevations reflect the buoyancy of the underlying continental and oceanic lithosphere. Where the oceanic lithosphere is extremely thin at mid-ocean ridges, the water depth is commonly between 2 and 3 km. This is the water depth at which we expect the top of a column of normal asthenosphere to lie. In some regions the mid-ocean ridge crest shallows and may become emergent, as in Iceland, suggesting that the underlying asthenosphere is hotter and therefore more buoyant than normal. Either side of the elevated mid-ocean ridges, the ocean basins deepen with a square root of crustal age relationship (see §2.2.7).

Continental topography extends to the edge of the continental shelves, at approximately 200 m present-day water depth, and up to the highest mountain peak at c.8 km. There is no relationship between the age of the continental crust and its elevation. Topography is primarily controlled by the history of deformation and erosion, with an upper limit on the height of mountain ranges provided by rock strength (Willett & Brandon 2002). That is, once a critical elevation is reached, rocks are prone to collapse under gravitational forces.

Simple calculations of isostasy underpin many aspects of basin analysis. Isostasy is the way in which hydrostatic equilibrium (Archimedes principle) influences the support of the oceanic and continental plates by the mantle. For equilibrium to be maintained, we can equate the surface forces due to the differing rock columns under continental and oceanic lithosphere (see Appendix 2). A familiar parameterisation for isostasy is of a mountain belt with a 'root', the excess mass in the elevated continental crust being compensated by the mass deficit at depth in the continental root.

The presence of continental blocks of varying thickness and topography adjacent to oceanic lithosphere of varying thickness causes pressure differences between adjacent columns of rock, which result in horizontal surface forces that act in addition to the lithostatic term (§2.1.1) (Appendices 3 and 4). These deviatoric or in-plane stresses may be compressive or tensile. For a continental crustal block adjacent to mantle-density oceanic lithosphere, the tensional horizontal deviatoric stress in the continent is of the order of −10 to −100 MPa, but if the continental block is 70 km thick, as in some mountain ranges, the tensile deviatoric stress may reach −150 MPa. Thickened continental lithosphere, as in continental plateaux such as Tibet and the Altiplano, consequently pushes against the adjacent platforms and ocean basins, leading to extension in the elevated regions.

Elevated continental plates surrounded by passive margins should therefore be in deviatoric tension. Continental plates surrounded by convergent boundaries, on the other hand, experience compressional deviatoric stresses roughly orthogonal to the orientation of the convergent boundary, as seen in the Indo-Australian plate (Hillis & Reynolds 2000) (Fig. 1.13).

1.4.4 Heat flow

A global heat flow map shows that the oceans have high surface heat flows, especially at the mid-ocean ridges (Fig. 1.14). The average surface heat flow value for the ocean crust is 101 mW m^{-2}, nearly double the continental average (Table 1.1). Fig. 1.14b demonstrates that the surface heat flow of the ocean floor shows the same pattern as the ocean bathymetry, with the same root-age dependence. This

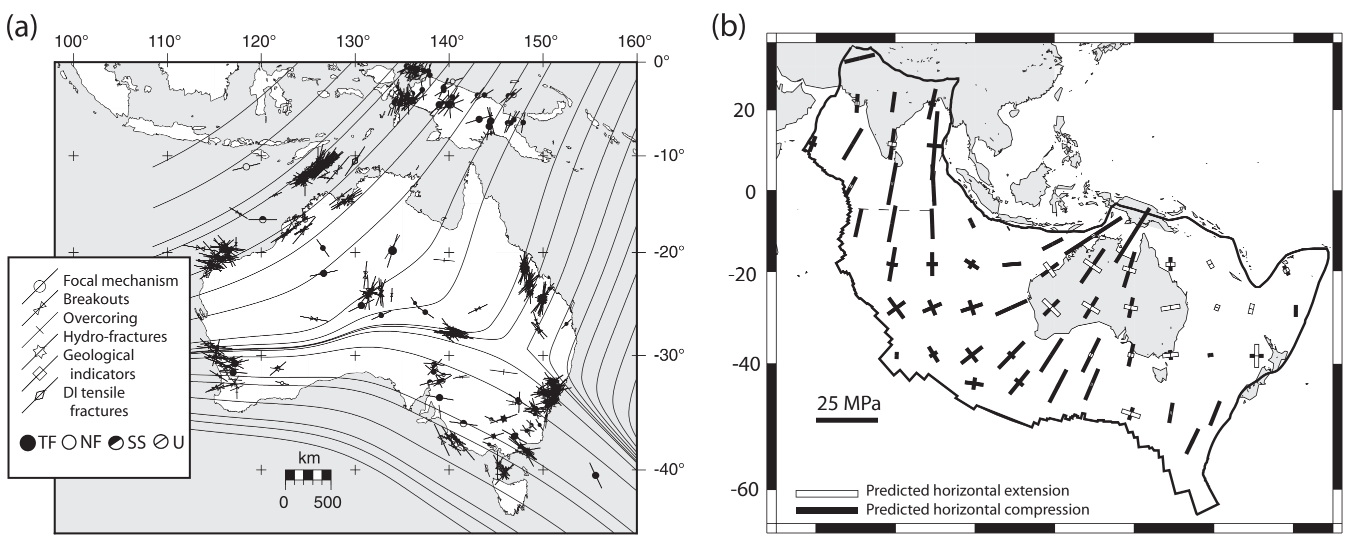

Fig. 1.13 (a) Stress trajectories based on the Australian stress map (Hillis & Reynolds 2000): NF, normal fault; SS, strike-slip; TF, thrust fault; U, unknown. (b) Stresses predicted by an elastic numerical model of the forces acting on the Indo-Australian plate, after Coblentz *et al*. (1998). Ridge push forces are balanced by fixing the collisional segments of the northeastern margin of the plate (Himalayas, New Guinea and New Zealand). Bars indicate orientation and magnitude of extensional (open) and compressive (solid) deviatoric stresses. Note that compressive deviatoric stresses align themselves normal to the convergent boundaries. Maximum deviatoric compressive forces are <50 MPa.

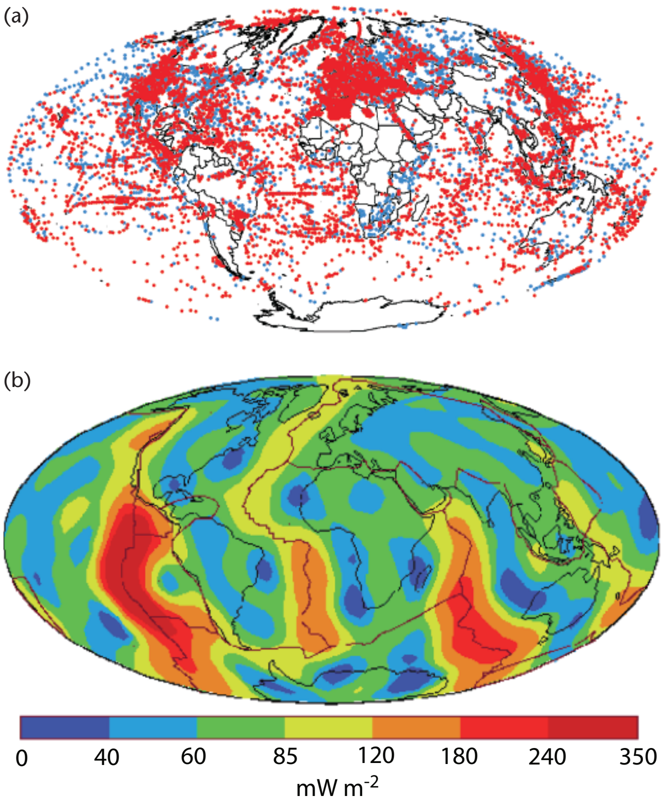

Fig. 1.14 (a) Global surface heat flow database, based on the compilation of Pollack *et al*. (1993) and additional data from Davies and Davies (2010), making over 38,000 measurements. (b) Contoured surface heat flow (degree 12 spherical harmonic), from Pollack *et al*. (1993).

shows that as the ocean lithosphere moves away from the spreading centre, it cools, thickens and sinks.

On the continents, there is considerable variability in surface heat flow values, but the variability is primarily explained by variations in the radiogenic heat production of continental rocks (§2.2.3). It is common for over 50% of the surface heat flow to be caused by radiogenic heating. Consequently, although the global average continental surface heat flow is 65 mW m^{-2}, the basal heat flows under the continents must be much smaller, typically between 20 and 30 mW m^{-2}. High concentrations of radiogenic elements in upper crustal rocks causes self-heating and the elevation of the geotherm above the non-radiogenic case. Consequently, although the oceanic geotherm may be linear due to conduction, the geotherm in continental lithosphere is typically curved (§2.2.2). Elevated geotherms may play an important role in controlling continental seismicity and deformation.

1.4.5 Cycles of plate reorganisation

The relative motion of plates with constructive, conservative and destructive plate margins creates a continually changing picture of continental splitting, ocean basin creation, ocean closure and continental collision. This cycle of plate motion involving the birth and closure of oceans is termed the Wilson cycle since it is based on early ideas of the opening and closing of the Atlantic Ocean by John Tuzo Wilson (Wilson 1966). Many sedimentary basins can be fitted into a particular phase of the Wilson cycle.

A related concept is that of the aggregation of supercontinents and their dispersal in response to convective flow in the mantle (Anderson 1982; Gurnis 1988). Rodinia, Gondwana and Pangaea are all examples of supercontinental assemblies, with a repeat time scale of the order of 300 Myr. Supercontinents are thought to act as insulators of the mantle, hindering it from losing heat as it does so effectively through the oceanic lithosphere. Over a period of time (about 10^8 yr), sub-plate temperatures would rise as the buoyant supercontinent is trapped on a geoid high between two adjacent cold downwellings. Eventually, the lateral spread of heat would allow the plate to move off the geoid high and to settle over a downwelling at a geoid low, causing the widespread flooding of the continental surface. The time scale of the supercontinental cycle therefore depends primarily on the time taken to incubate the sub-plate mantle sufficiently to inflate the supercontinent (Grigné *et al.* 2007), cause break-up, and lead to migration of continental fragments to adjacent geoid lows. The Wilson cycle is therefore driven fundamentally by mantle circulation (Gurnis, 1988).

Although very generalised, a wide range of basin types can be placed in the geodynamical context of supercontinental assembly and dispersal. Extension of the supercontinent triggers the formation of the rift-drift suite of sedimentary basins, comprising continental rifts, continental rim basins, cratonic basins, failed rifts, proto-oceanic troughs, and passive margins. The subduction phase associated with convergence triggers the formation of ocean trenches, accretionary wedge basins and fore-arc basins trenchwards of the magmatic arc, a range of basins in the retro-arc region, and strike-slip basins linked to oblique convergence. Continental collision is typified by foreland basin systems, including wedge-top deposition (De Celles & Giles 1996; Ford 2004). Consequently, certain basin types are common during particular phases of the Wilson cycle. In addition, the frequency of occurrence of basin types will depend on their long-term preservation by being shielded from tectonic recycling. This is an important consideration for Proterozoic basins, whose preservation potential is enhanced by incorporation into relatively rigid, stable, cratonic cores.

The dispersal and reassembly of plates may take place in two modes, termed introversion and extroversion (Nance & Murphy 2003; Murphy & Nance 2008). In the case of introversion, plates disperse from the supercontinental core, producing new interior oceans bordered by extensional trailing edges of the continental plates, and convergent boundaries along their external perimeter. Subsequent subduction of the interior ocean causes it to close and to result in reassembly of the plates in essentially the same orientation as in the original super-assembly. Extroversion, on the other hand, involves the continued expansion of the continental plates away from the supercontinental core, and the eventual closure of the external ocean by subduction along the advancing edges of the dispersing continental plates. Introversion and extroversion therefore control the large-scale setting for basin development at the time scale of the supercontinental cycle.

1.5 Classification schemes of sedimentary basins

Ideally, classifications are theories about the basis of natural order rather than dull catalogues compiled only to avoid chaos (Gould

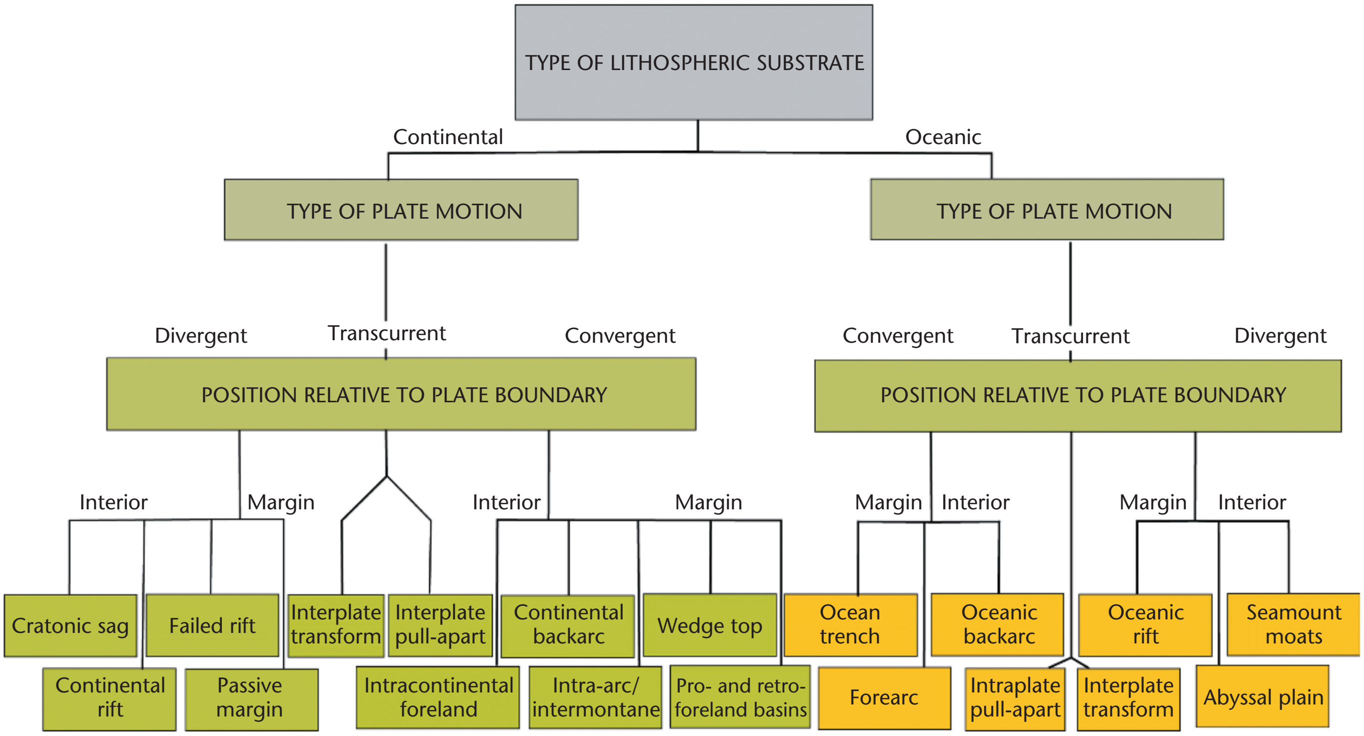

Fig. 1.15 Classification scheme for sedimentary basins based on their plate tectonic setting. Modified from scheme used by Kingston *et al*. (1983a). AAPG © 1983. Reprinted by permission of the AAPG whose permission is required for further use.

1989, p. 98, quoted in Ingersoll & Busby 1995, p. 2). In this sense, classification schemes for sedimentary basins should both reveal something of the underlying mechanisms for basin development and reflect the natural variability of the real world.

Classification schemes of sedimentary basins based on plate tectonics have much in common. Their lineage derives from Dickinson's influential work in 1974, which emphasised the position of the basin in relation to the type of lithospheric substrate, the proximity of the basin to a plate margin, and the type of plate boundary nearest to the basin (divergent, convergent, transform) (Fig. 1.15). The evolution of a basin could then be explained by changing plate settings and interactions.

Bally (1975) and Bally and Snelson (1980) differentiated three different families of sedimentary basins based on their location in relation to megasutures, which in this context can be defined to include all the products of orogenic and igneous activity associated with predominantly contractional deformation. The boundaries of megasutures are often associated with subduction, whether it be of slabs of oceanic lithosphere (Benioff or B-type subduction) or of relatively buoyant continental lithosphere (Amferer or A-type subduction) and may also be the sites of important wrench tectonism along transform faults. Ingersoll and Busby (1995) developed the classifications of Dickinson (1974) and Ingersoll (1988) to recognise 26 different types grouped into classes of divergent settings, intraplate settings, convergent settings, transform settings and hybrid settings. Within these settings, Ingersoll (2011) suggested that there were numerous variants depending on sediment supply and geological inheritance.

1.5.1 Basin-forming mechanisms

The goal of categorising a sedimentary basin and thereby gaining some predictive insights into the hydrocarbon potential of frontier basins is common to industry classifications (e.g. Huff 1978; Klemme 1980; Kingston *et al*. 1993a, b). Although such classifications undoubtedly have their uses, particularly in predicting the presence of key elements of the petroleum play (Chapter 11), they have the effect of scrambling some of the essential differences and similarities between basins from the point of view of geodynamical mechanisms.

Ingersoll and Busby (1995) and Ingersoll (2011) recognised seven subsidence mechanisms, operating to different degree in their various basin types (Fig. 1.16), which can be summarised as:

1. crustal thinning, such as caused primarily by stretching or surface erosion;
2. lithospheric thickening, such as caused by cooling following stretching or accretion of melts derived from the asthenosphere;
3. sedimentary and volcanic loading causing isostatic compensation;
4. tectonic (supracrustal) loading causing isostatic compensation;
5. subcrustal loading caused by subcrustal dense loads such as magmatic underplates or obducted mantle flakes;
6. mantle flow primarily due to the subduction of cold lithospheric slabs;
7. crustal densification due to changing P-T conditions or intrusion of high-density melts.

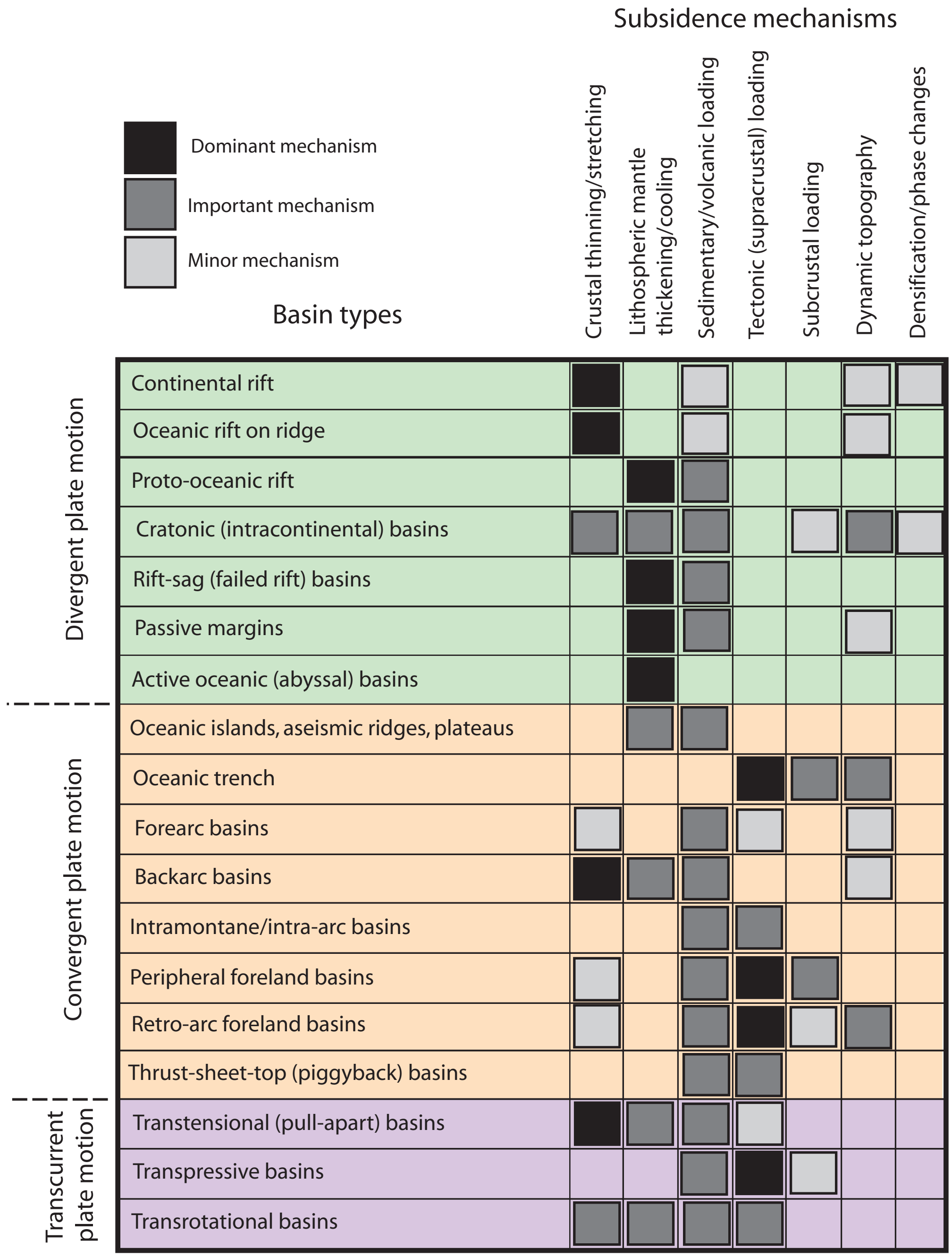

Fig. 1.16 Matrix of main basin types and the principal mechanisms of subsidence (adapted from Ingersoll & Busby 1995 and Ingersoll 2011). Note that subsidence mechanisms operate in more than one basin type, and that several mechanisms may operate in a single basin type. Some basins are 'polyhistory' and may pass through phases characterised by different sets of mechanisms.

From the point of view of fundamental lithospheric processes, the major mechanisms for regional subsidence and uplift (*isostatic*, *flexural* and *dynamic*) can be summarised as follows (Fig. 1.17):

- *Isostatic* consequences of changes in crustal and lithospheric thickness; the thickness changes may be brought about purely thermally by *cooling* of lithosphere, for example following mechanical stretching. Thickness changes causing *thinning* may be caused by mechanical stretching, subaerial erosion or at depth by removal (or delamination) of a deep lithospheric root. Mechanical *thickening* of crust and lithosphere, as in zones of continental convergence, generally causes isostatic uplift. Thickening of the lithosphere by cooling, however, causes subsidence.
- *Loading* and unloading at the surface and in the subsurface, including the far-field effects of in-plane stresses; *loading* of the lithosphere may take place on a small scale in the form of volcanoes or seamount chains, and on a large scale in the form of mountain belts, causing *flexure* and therefore subsidence. The

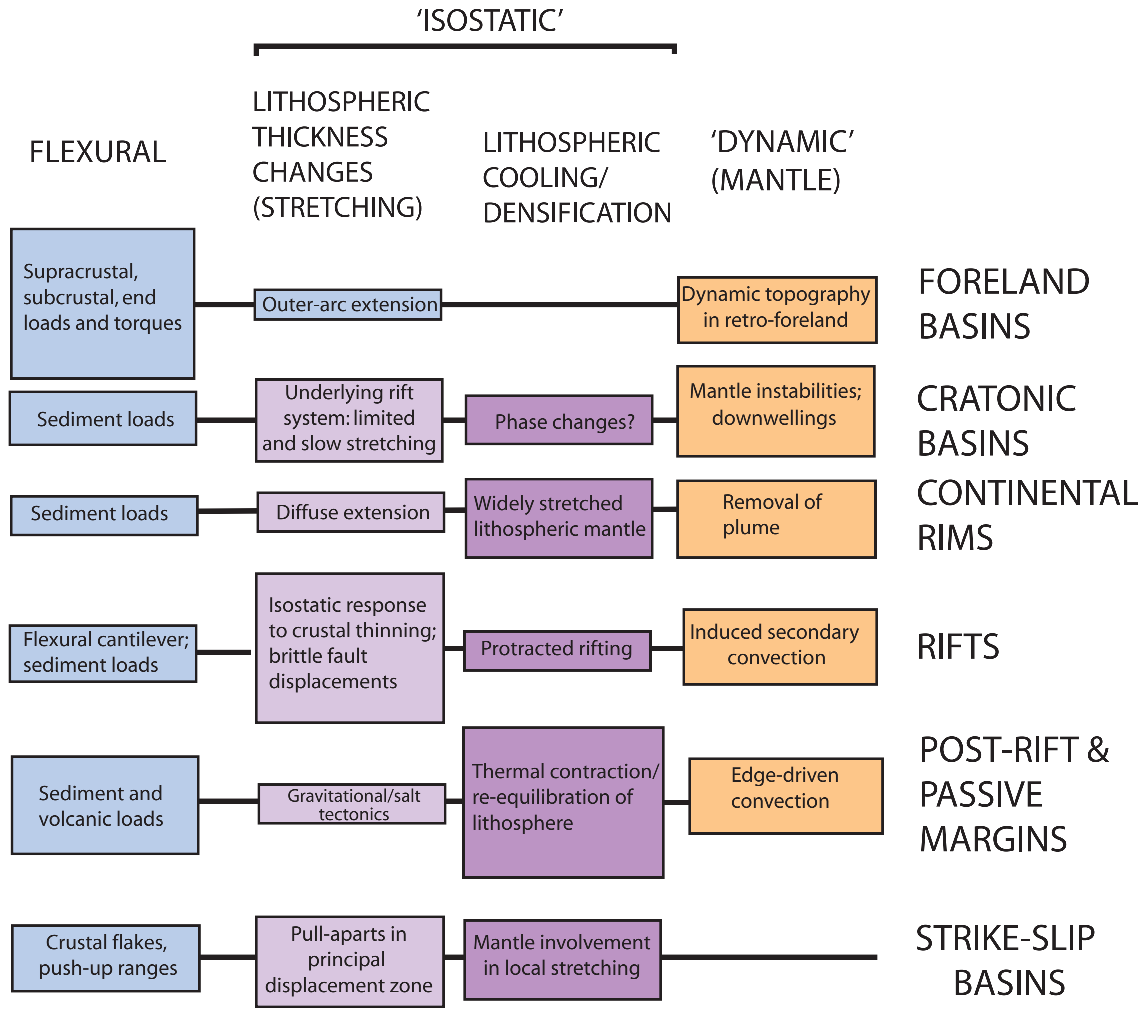

Fig. 1.17 Genetic basin classification based on flexural, isostatic and dynamic mechanisms, and their importance in a number of common basin types. Size of box reflects roughly the importance of the mechanism in the formation and evolution of the basin type. Cratonic basins, continental rims, rifts, failed rifts and passive margins belong to a suite of basins associated with continental extension. Foreland basins are associated with flexure, whereas strike-slip basins are controlled by the structural geology of zones of transcurrent deformation.

sediment infilling a basin also acts as a sedimentary load, amplifying the primary driving mechanism.

- *Dynamic* effects of asthenospheric flow, mantle convection and plumes; subsidence or uplift are caused by the buoyancy effects of changes in temperature in the mantle. Since these temperature changes are transmitted by viscous flow, the surface elevation changes may be termed *dynamic*.

For a given basin type, some or all of the mechanisms given above may have a major or minor role (Fig. 1.17). In addition, a given mechanism operates in more than one basin type. The classification scheme is therefore a matrix. Rather than attempting an encyclopaedic coverage of all basin types, we focus on the main lithospheric processes in Part 2 of this book. After an introductory chapter on the fundamentals of lithospheric mechanics (Chapter 2), basin-forming processes are considered by investigating basins primarily caused by lithospheric stretching (Chapter 3), and basins primarily caused by flexure (Chapter 4). Chapter 5 discusses the role of mantle-lithosphere interactions in basin development. Basins related to strike-slip deformation are considered in Chapter 6.

Basins formed by stretching or thinning of the continental lithosphere fall within an evolutionary sequence (Kinsman 1975; Veevers 1981) (§3.1.1). The early stages of the sequence correspond to the development of intracontinental sags (cratonic basins) and continental rim basins, which lack clear evidence of brittle stretching, and continental rifts, which comprise clear extensional fault systems that may or may not be associated with topographic doming. Such rifts may evolve into oceanic spreading centres or may be aborted to form failed rifts or aulacogens. With seafloor creation and drifting of the continental edge away from the spreading centre, passive margin basins develop. The sequence has been termed the *rift–drift suite* of sedimentary basins. The mechanisms of interest within this evolutionary sequence are therefore primarily the thermal and mechanical behaviour of the lithosphere under tension, and the thermal contraction of the lithosphere following stretching.

Basins formed by flexure are mainly linked to plate convergence (Chapter 4). Flexure of oceanic lithosphere as it approaches subduction zones is responsible for the formation of deep oceanic trenches. It was the investigation of the deflection of the oceanic lithosphere at arc–trench boundaries that provided much of the framework for the general theory of lithospheric flexure. Flexure of the continental lithosphere in continental collision zones gives rise to *foreland basin systems*. Flexure of the lower or subducting plate generates pro-foreland basins, whereas flexure of the upper or overriding plate generates retro-foreland basins (Naylor & Sinclair 2008). Where subduction zone roll-back takes place, as in the Adriatic-Apennines region of Italy, the lower plate is flexed to produce a pro-foreland basin, but the retro-foreland region is extensional. Foreland basin pairs may also be associated with intracontinental mountain building where there is no subduction, as in the late Precambrian and Paleozoic of central Australia.

Processes within the mantle have an important role in basin development (Chapter 5). Mantle flow structures, probably originating at the core–mantle boundary, impinge on the base of the lithosphere and spread out laterally over a length scale of 10^3 km. They are instrumental in continental splitting and the formation of new ocean basins, and have a major role in igneous underplating and isostatic regional uplift. Present-day *hotspots* over ascending limbs of mantle convection systems are characterised by topographic doming and commonly rifting. Basins located over downward limbs of convection systems, *cold-spot basins*, appear to be broad, gentle sags. The onset of subduction of cold oceanic slabs at ocean–continent boundaries causes far-field tilting of the continental plate towards its margin and therefore has a potentially major effect in forearc, intra-arc and retro-arc settings. In addition, it is believed that supercontinental assemblies in the geological past have experienced very long-wavelength topographic doming caused by elevated sub-lithospheric temperatures generated by an overlying insulating lid. Mantle flow instabilities associated with steps in the base of the lithosphere, or associated with the drip-like removal of gravitationally unstable mantle lithosphere, may also have important impacts on surface topography.

Different basin types have a typical time span of existence (Ingersoll & Busby 1995), after which they may be uplifted and eroded, locally inverted, or modified into another basin type to become polyhistory in nature. Time spans may range over 4 orders of magnitude (Woodcock 2004). Oceanic trench basins have the shortest life span (0.1–0.8 Myr) due to rapid subduction velocities and early incorporation into an accretionary wedge. Trench-slope, intra-arc, extensional back-arc and strike-slip basins also have relatively short life spans in the range 2–25 Myr, reflecting their tectonically active settings. Rift, forearc, back-arc and foreland basins have longer time spans in the range 4–125 Myr. The longest life spans (60–440 Myr) are found in cratonic, passive margin and ocean basins, since they are situated in regions of low strain rate, experiencing prolonged thermal subsidence.