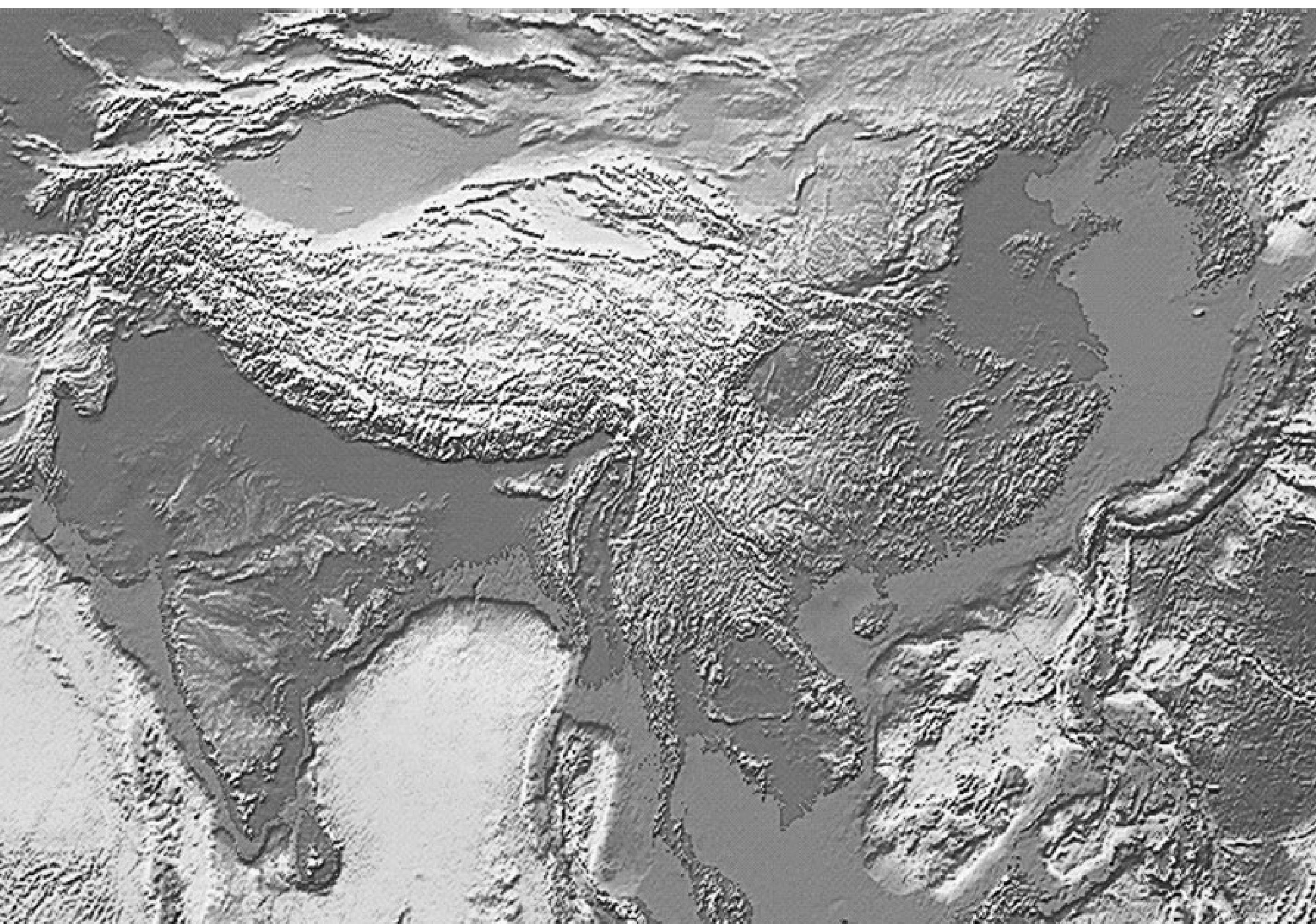
11 Precambrian tectonics and the supercontinent cycle



11.1 INTRODUCTION

The relatively flat, stable regions of the continents contain remnants of Archean crust that formed some 4.4 to 2.5 billion years ago (Plate 11.1a between pp. 244 and 245). The formation of these cratonic nuclei marks the transition from an early Earth that was so hot and energetic that no remnants of crust were preserved, to a state where crustal preservation became possible. Most of the cratons are attached to a high velocity mantle root that extends to depths of at least 200 km (King, 2005) (Plate 11.1b,c between pp. 244 and 245). These cratonic roots are composed of stiff and chemically buoyant mantle material (Section 11.3.1) whose resistant qualities have contributed to the long-term survival of the Archean continental lithosphere (Carlson *et al.*, 2005).

The beginning of the Archean Eon approximately coincides with the age of the oldest continental crust. A conventional view places this age at approximately 4.0 Ga, which coincides with the age of the oldest rocks found so far on Earth: the Acasta gneisses of the Slave craton in northwestern Canada (Bowring & Williams, 1999). However, >4.4 Ga detrital zircon minerals found in the Yilgarn craton of Western Australia (Wilde *et al.*, 2001) suggest that some continental crust may have formed as early as 4.4–4.5 Ma, although this interpretation is controversial (Harrison *et al.*, 2005, 2006; Valley *et al.*, 2006). Because evidence for continental crust and the ages of the oldest known rocks and minerals continually are being pushed back in time, the Archean has no defined lower boundary (Gradstein *et al.*, 2004). The end of the Archean, marking the beginning of the Proterozoic Eon, approximately coincides with inferred changes in the tectonic style and the petrologic characteristics of Precambrian rocks. It is these inferences that are central to a debate over the nature of tectonic activity in Precambrian times. Among the most important issues are whether some form of plate tectonics was operating in the early Earth and, if so, when it began. Current evidence (Sections 11.3.3, 11.4.3) suggests that plate tectonic mechanisms, including subduction, were occurring at least by 2.8–2.6 Ga and possibly much earlier (van der Velden *et al.*, 2006; Cawood *et al.*, 2006).

In considering the nature of Precambrian tectonic processes, three approaches have been adopted (Kröner, 1981; Cawood *et al.*, 2006). First, a strictly uniformitarian approach is taken in which the same mechanisms of plate tectonics that characterize Phanerozoic times are applied to the Precambrian cratons. This approach is common in the interpretation of Proterozoic belts, although it also has been applied to parts of the Archean cratons. Second, a modified uniformitarian approach can be postulated in which plate tectonic processes in the Precambrian were somewhat different from present because the physical conditions affecting the crust and mantle have changed throughout geologic time. This approach has been used in studies of both Archean and Early Proterozoic geology. Third, alternatives to plate tectonic mechanisms can be invoked for Precambrian times. This latter, nonuniformitarian approach most often is applied to the Early and Middle Archean. Each of these three approaches has yielded informative results.

11.2 PRECAMBRIAN HEAT FLOW

One of the most important physical parameters to have varied throughout geologic time is heat flow. The majority of terrestrial heat production comes from the decay of radioactive isotopes dispersed throughout the core, mantle, and continental crust (Section 2.13). Heat flow in the past must have been considerably greater than at present due to the exponential decay rates of these isotopes (Fig. 11.1). For an Earth model with a K/U ratio derived from measurements of crustal rocks, the heat flow in the crust at 4.0 Ga would have been three times greater than at the present day and at 2.5 Ga about two times the present value (Mareschal & Jaupart, 2006). For K/U ratios similar to those in chondritic meteorites, which are higher than those in crustal rocks, the magnitude of the decrease would have been greater.

The increased heat flow in Archean times implies that the mantle was hotter in the younger Earth than it is today. However, how much hotter and whether a hotter mantle caused young continental lithosphere to be much warmer than at present is uncertain. This uncertainty arises because there is no direct way to determine the ratio of heat loss to heat produc-

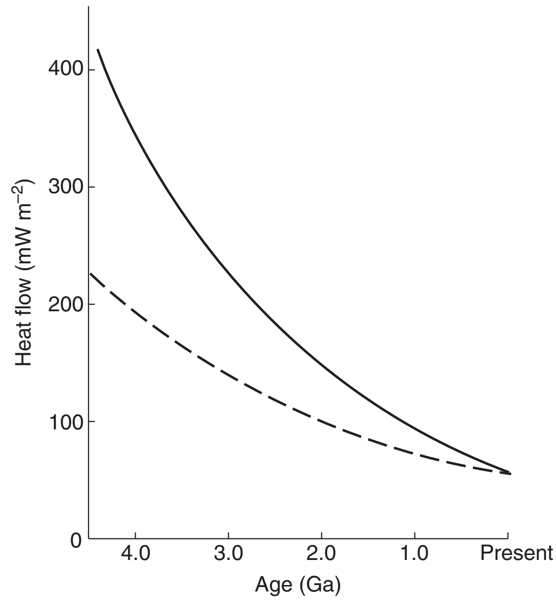


Fig. 11.1 *Variation of surface heat flow with time. Solid line, based on a chondritic model; dashed line, based on a K/U ratio derived from crustal rocks (redrawn from McKenzie & Weiss, 1975, with permission from Blackwell Publishing).*

tion in the early Earth. If the heat loss mostly occurred by the relatively inefficient mechanism of conduction then the lithosphere would have been warmer. However, if the main mechanism of heat loss was convection beneath oceanic lithosphere, which is very effective at dissipating heat, then the continental lithosphere need not have been much hotter (Lenardic, 1998). Clarifying these aspects of the Archean thermal regime is essential in order to reconstruct tectonic processes in the ancient Earth and to assess whether they were different than they are today.

Another part of the challenge of determining the Precambrian thermal regime is to resolve an apparent inconsistency that comes from observations in the crust and mantle parts of Archean lithosphere. Geologic evidence from many of the cratons, including an abundance of high temperature/low pressure metamorphic mineral assemblages and the intrusion of large volumes of granitoids (Section 11.3.2), suggest relatively high (500–700 or 800°C) temperatures in the crust during Archean times, roughly similar to those which occur presently in regions of elevated geotherms. By contrast, geophysical surveys and isotopic studies of mantle nodules suggest that the cratonic mantle is strong and cool and that the geotherm has been relatively low since the Archean (Section 11.3.1). Some of the most compelling evidence of cool mantle lithosphere comes from thermobarometric studies of silicate inclusions in Archean diamonds, which suggest that temperatures at depths of 150–200 km during the Late Archean were similar to the present-day temperatures at those depths (Boyd *et al.*, 1985; Richardson *et al.*, 2001). Although geoscientists have not yet reconciled this apparent inconsistency, the relationship provides important boundary conditions for thermal models of Archean and Proterozoic tectonic processes.

In addition to allowing estimates of ancient mantle geotherms, the evidence from mantle xenoliths indicates that the cool mantle roots beneath the cratons quickly reached their current thickness of ≥200 km during Archean times (Pearson *et al.*, 2002; Carlson *et al.*, 2005). This thickness is greater than that of old oceanic lithosphere but much thinner than it would be if the lithosphere simply had cooled from above by conduction since the Archean (Sleep, 2005). Progressive thickening by conductive cooling also can be ruled out because the mantle roots do not display an age progression with depth (James & Fouch, 2002; Pearson *et al.*, 2002). Instead, the relatively small thickness and long-term preservation of the cratonic roots indicate that they must have been kept thin by convective heat transfer from the underlying mantle (Sleep, 2003). Once the cratonic roots stabilized, the heat supplied to the base of the lithosphere from the rest of the mantle must have been balanced by the heat that flows upward to the surface. In this model, a chemically buoyant layer of lithosphere forms a highly resistant lid above the convecting mantle, allowing it to maintain nearly constant thickness over time. These considerations illustrate how the formation and long-term survival of the cool mantle roots beneath the cratons has helped geoscientists constrain the mechanisms of heat transfer during Precambrian times.

Differences in the inferred mechanism of heat loss from the Earth's interior have resulted in contrasting views about the style of tectonics that may have operated during Precambrian times (e.g. Hargraves, 1986; Lowman *et al.*, 2001; van Thienen *et al.*, 2005). A conventional view suggests that an increased heat supply in the Archean mantle could be dissipated by increasing the length of ocean ridge systems or by increasing the rate of plate production with respect to the present (Bickle, 1978). Hargraves (1986) concluded that heat loss through the oceanic lithosphere is proportional to the cube root of the total length of the mid-ocean ridge. Assuming a nonexpanding Earth (Section 12.3), the increased rate of plate production implies a similar increase in plate subduction rate. These computations suggest that some form of plate tectonics was taking place during the Precambrian at a much greater rate than today. The fast rates suggest an image of the solid surface of the early Earth where the lithosphere was broken up into many small plates that contrasts with the relatively few large plates that exist presently. This interpretation is consistent with the results of numerical models of mantle convection, which show that small plates are capable of releasing more heat from the Earth's interior than large plates (Lowman *et al.*, 2001).

More recent calculations have disputed this conventional view, at least for the Late Archean. Van Thienen *et al.* (2005) suggested that the increased heat flux from the Archean mantle could have been dissipated by thinning the lithosphere and thereby increasing the heat flow through the lithosphere. These authors concluded that for a steadily (exponentially) cooling Earth, plate tectonics is capable of removing all the required heat at a plate tectonic rate comparable to or lower than the current rate of operation. This result is contrary to the notion that faster spreading would be required in a hotter Earth to be able to remove the extra heat (e.g. Bickle, 1978). It also suggests that reduced slab pull and ridge push forces in a hotter mantle would result in a slower rate of plate tectonics compared to the modern Earth. Korenaga (2006) showed that a more sluggish style of plate tectonics during Archean times satisfies all the geochemical constraints on the abundance of heat-producing elements in the crust and mantle and the evidence for a gradual cooling of the mantle with time in the framework of whole mantle convection. This result removes the thermal necessity of having extensive ocean ridges and/or rapid spreading and subduction. It must be appreciated, however, that thermal conditions during Archean times are quite conjectural, so that these and other alternative interpretations remain speculative.

11.3 ARCHEAN TECTONICS

11.3.1 General characteristics of cratonic mantle lithosphere

A defining characteristic of the cratonic mantle lithosphere is a seismic velocity that is faster than normal subcontinental mantle to depths of at least 200 km and locally to depths of 250–300 km (Plate 11.1b,c between pp. 244 and 245). Many Proterozoic belts lack these fast velocity anomalies at similar depths. In addition, Archean cratons are characterized by the lowest surface heat flow of any region on Earth, with a heat flux that is lower than adjacent Proterozoic and Phanerozoic crust by some 20 $mW\,m^{-2}$ (Jaupart & Mareschal, 1999; Artemieva & Mooney, 2002). Isotopic age determinations and Re-Os studies of mantle nodules (Pearson *et al.*, 2002; Carlson *et al.*, 2005) confirm that the mantle roots are Archean in age and indicate that most have remained thermally and mechanically stable over the past 2–3 Ga. These observations indicate that the roots of the Archean cratons are cool, strong, and compositionally distinct from the surrounding mantle.

In the ocean basins, the base of the oceanic lithosphere is marked by a strong decrease in the velocity of shear waves at depths generally less than 100 km beneath the crust (Sections 2.8.2, 2.12). Similar low velocity zones occur under tectonically active continental regions, such as the Basin and Range Province, but beneath the stable cratons low velocity zones are either extremely weak or entirely absent (Carlson *et al.*, 2005). Consequently the base of the continental lithosphere is not well defined by seismological data. With increasing depth the high seismic velocities of the stable lithosphere gradually approach those of the convecting mantle across a broad, ill-defined transition zone below 200 km. Thermal modeling and geochemical data from mantle xenoliths have helped to define the location of the lithosphere–asthenosphere boundary. The results suggest that the base of the continental lithosphere is deepest (~250 km) in undisturbed cratonic areas and shallowest (~180 km) beneath Phanerozoic rifts and orogens (O'Reilly *et al.*, 2001). This determination is in general agreement with seismic observations.

In addition to being cool and strong, studies of mantle xenoliths indicate that the Archean mantle roots are chemically buoyant and highly depleted in incompatible elements (O'Reilly *et al.*, 2001; Pearson *et al.*, 2002). When mantle melting occurs elements such as calcium, aluminum and certain radiogenic elements are concentrated into and extracted by the melt whereas other elements, particularly magnesium, selectively remain behind in the solid residue. Those elements that concentrate into the melt are known as *incompatible* (Section 2.4.1). Both buoyancy and chemical depletion are achieved simultaneously by partial melting and melt extraction, which, in the case of the mantle lithosphere, has left behind a residue composed of Mg-rich harzburgites, lherzolites, and peridotite (O'Reilly *et al.*, 2001). Eclogite also appears to be present in the cratonic lithosphere. However, high velocity bodies consistent with large, dense masses of eclogite have not been observed in the continental mantle (James *et al.*, 2001; Gao *et al.*, 2002). An inventory of mantle xenoliths from the Kaapvaal craton suggests that eclogite reaches abundances of only 1% by volume in the continental mantle (Schulze, 1989). These characteristics have resulted in the mechanical and thermal stability of the cratons for up to three billion years (Section 11.4.2).

11.3.2 General geology of Archean cratons

Archean cratons expose two broad groups of rocks that are distinguished on the basis of their metamorphic grade: *greenstone belts* and *high grade gneiss terrains* (Windley, 1981). Both groups are intruded by large volumes of granitoids. Together these rocks form the Archean *granite-greenstone belts*. The structure and composition of these belts provide information on the origin of Archean crust and the evolution of the early Earth.

The greenstones consist of metavolcanic and metasedimentary rocks that exhibit a low pressure (200–500 MPa), low temperature (350–500°C) regional metamorphism of the greenschist facies. Their dark green color comes from the presence of minerals that typically occur in altered mafic (i.e. Mg- and Fe-rich) igneous rock, including chlorite, actinolite, and epidote. Three main stratigraphic groups are recognized within greenstone belts (Windley, 1981). The lower group is composed of tholeiitic and komatiitic lavas. *Komatiites*, named after the Komati Formation in the Barberton Greenstone belt of the Kaapvaal craton, South Africa (Viljoen & Viljoen, 1969), are varieties of Mg-rich basalt and ultramafic lava that occur almost exclusively in Archean crust. The high Mg content (>18 wt% MgO) of these rocks (Nisbet *et al.*, 1993; Arndt *et al.*, 1997) commonly is used to infer melting temperatures that are higher than those of modern basaltic magmas (Section 11.3.3). The central group contains intermediate and felsic volcanic rocks whose trace and rare earth elements are similar to those found in some island arc rocks. The upper group is composed of clastic sediments, such as graywackes, sandstones, conglomerates, and banded iron formations (BIFs). These latter rocks are chemical-sedimentary units consisting of iron oxide layers that alternate with chert, limestone, and silica-rich layers (see also Section 13.2.2).

High-grade gneiss terrains typically exhibit a low pressure, high temperature (>500°C) regional metamorphism of the amphibolite or granulite facies (Section 9.9). These belts form the majority of the area of Archean cratons. A variety of types commonly are displayed, including quartzofeldspathic gneiss of mostly granodiorite and tonalite composition, layered peridotite-gabbro-anorthosite or leucogabbro-anorthosite complexes, and metavolcanic amphibolites and metasediment (Windley, 1981). Peridotite (Sections 2.4.7, 2.5) is an ultramafic rock rich in olivine and pyroxene minerals. Leucogabbro refers to the unusually light color of the gabbroic rock due to the presence of plagioclase. *Anorthosites* are plutonic rocks consisting of >90% plagioclase and have no known volcanic equivalents. These latter rocks occur exclusively in Archean and Proterozoic crust. Most authors view Archean anorthosites as having differentiated from a primitive magma, such as a basalt rich in Fe, Al and Ca elements or, possibly, a komatiite (Winter, 2001). High-grade gneiss terrains are highly deformed and may form either contemporaneously with, structurally below, or adjacent to the low-grade greenstone belts (Percival *et al.*, 1997).

The granitoids that intrude the greenstones and high-grade gneisses form a compositionally distinctive group known as *tonalite-trondhjemite-granodiorite*, or *TTG*, suites (Barker & Arth, 1976). Tonalites (Section 9.8) and trondhjemites are varieties of quartz diorite that typically are deficient in potassium feldspar. These igneous suites form the most voluminous rock associations in Archean crust and represent an important step in the formation of felsic continental crust from the primordial mantle (Section 11.3.3).

11.3.3 The formation of Archean lithosphere

The distinctive composition and physical properties of the stiff, buoyant mantle roots beneath the cratons (Section 11.3.1) result from the chemical depletion and extraction of melts from the primitive mantle. These two processes lowered the density and increased the viscosity of the residue left over from partial melting and resulted in a keel that consists mostly of high-Mg olivine and high-Mg orthopyroxene (O'Reilly *et al.*, 2001; Arndt *et al.*, 2002). Both of these components are absent in fertile (undepleted) mantle peridotite and are rare in the residues of normal mantle melting, such as that which produces modern oceanic crust and oceanic islands. Consequently, most workers have concluded that the distinctive composition is related to unusually high degrees (30–40%) of mantle melting over a range (4–10 GPa) of mantle pressures (Pearson *et al.*, 2002; Arndt *et al.*, 2002). High-degree partial melting of mantle peridotite produces magma of komatiitic (Section 11.3.2) composition and a solid residue that is very similar to the composition of Archean lithospheric mantle (Herzberg, 1999; Arndt *et al.*, 2002).

One radiometric system that has been of considerable use in determining when melt extraction and Archean root formation occurred involves the decay of ^{187}Re to ^{188}Os (Walker *et al.*, 1989; Carlson *et al.*, 2005). The key feature of this isotopic system is that Os is compatible during mantle melting and Re is moderately incompatible. Consequently, any residue left behind after melt extraction will have a lower Re and a higher Os concentration than in either the mantle melts or the fertile mantle. This characteristic allows Re-Os isotope analyses of mantle xenoliths to yield information on the age of melt extraction. The data from mantle xenoliths show that the oldest melting events are Early–Middle Archean in age. Significant amounts of lithospheric mantle also formed in Late Archean times and are associated with voluminous mafic magmatism (Pearson *et al.*, 2002).

Although high-degree partial melting undoubtedly occurred, this process alone cannot explain the origin of the Archean mantle lithosphere. The main reason for this conclusion is that the abundance of komatiite found in the Archean crustal record is much too low to balance the amount of highly depleted peridotite found in the cratonic mantle (Carlson *et al.*, 2005). This imbalance suggests that either a large proportion of komatiitic magma never reached the surface or other processes must have contributed to root formation. One of these processes is an efficient sorting mechanism that concentrated the unusual components of the Archean mantle lithosphere at the expense of all others (Arndt *et al.*, 2002). The most likely driving force of the sorting is the buoyancy and high viscosity of high-Mg olivine and orthopyroxene, although exactly how this happened is uncertain. The density and viscosity of these minerals depends upon their Mg–Fe ratios and water content, respectively; which are lower in Archean mantle lithosphere compared to normal asthenospheric mantle. Arndt *et al.* (2002) considered three processes that could have resulted in the mechanical segregation and accumulation of a layer of buoyant, viscous mantle near the Earth's surface during Archean times. First, upwelling buoyant residue in the core of a mantle plume could have separated from the cooler, denser exterior and accumulated during ascent (Fig. 11.2a). In this model, melting begins at high pressure (~200 km depth) and continues to shallow depths, by which point melt volumes are high and the dense residue of early, low-degree melting is swept away by mantle flow. Second, buoyant residue could have segregated slowly as material was transported down subduction zones and recycled through the mantle in convection cells (Fig. 11.2b). Third, some subcontinental lithosphere could be the remnants of an initial crust that crystallized in an Archean magma ocean that formed during the final stages of Earth accretion (Fig. 11.2c). In all three cases, buoyant, viscous material rises and is separated from higher density residue during mantle flow. Whether some combination of these or other processes helped to form the cratonic keels is highly speculative. Nevertheless, they illustrate how several different mechanisms could have concentrated part of the residue of mantle melting into a near-surface layer.

In addition to high-degree partial melting and efficient sorting, most authors also have concluded that the formation and evolution of mantle lithosphere involved a multi-stage history involving many tectonic and magmatic events (James & Fouch, 2002). However, opinions are divided over whether root construction preferentially involved the underthrusting and stacking of subducted oceanic slabs (Carlson *et al.*, 2005), the accretion and thickening of arc material (Lee, 2006), or the extraction of melt from hot mantle plumes (Wyman & Kerrich, 2002). By applying a range of criteria some geologic studies have made compelling cases that ancient mantle plumes played a key role in the

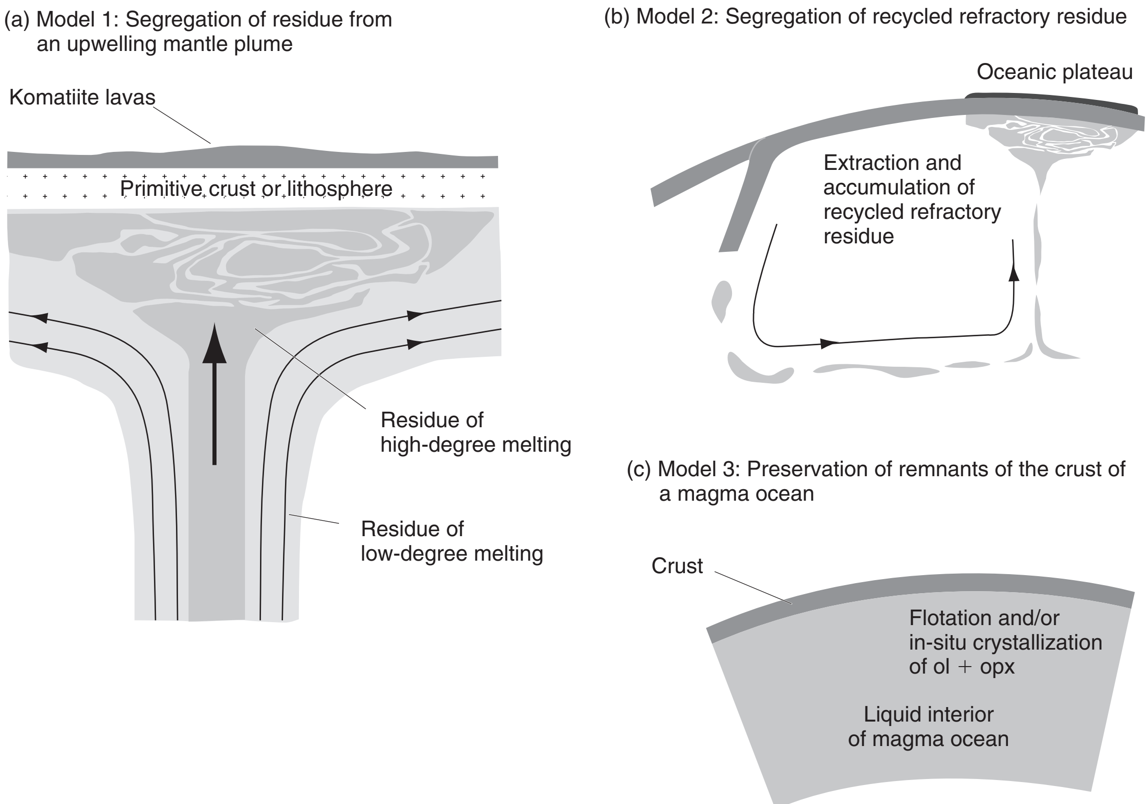


Fig. 11.2 *(a–c) Three possible mechanisms that could allow the segregation and accumulation of high-Mg olivine and orthopyroxene near the surface of the Earth (after Arndt et al., 2002, with permission from the Geological Society of London).*

evolution of Archean lithosphere (Tomlinson & Condie, 2001; Ernst *et al.*, 2005). Data from seismic profiles, geochronologic studies, and isotopic analyses indicate that many roots were affected by large pulses of mafic magmatism during the Late Archean (Wyman & Kerrich, 2002; James & Fouch, 2002). Other studies, however, have emphasized a subduction zone setting to explain the evolution of Archean mantle lithosphere. Most of the cratons display evidence for the significant modification of cratonic roots by terrane collisions and thickening during at least some stage in their history (James & Fouch, 2002; Schmitz *et al.*, 2004). In support of a subduction zone mechanism, a Late Archean (2.8–2.6 Ga) fossil subduction zone (Fig. 11.3) has been found within the Abitibi craton in northern Canada using seismic data (Calvert & Ludden, 1999; van der Velden *et al.*, 2006). Nevertheless, it is important to recognize that Archean mantle roots probably resulted from more than one tectonic environment and that no single setting or event is applicable to all cases.

The distinctive rock associations that comprise granite-greenstone belts (Section 11.3.2) provide another important means of evaluating the mechanisms that contributed to the formation and evolution of Archean lithosphere. One of the key questions to answer is whether the komatiitic and tholeiitic lavas that form the majority of the greenstones formed in environments that were broadly similar to modern tectonic environments. For example, if these lavas loosely represent the Archean equivalent of modern mid-ocean ridge basalts, as is commonly believed, then they might be used to infer that much of the volcanism in Archean times involved the creation and destruction of ocean crust (Arndt *et al.*, 1997). However, one of the problems

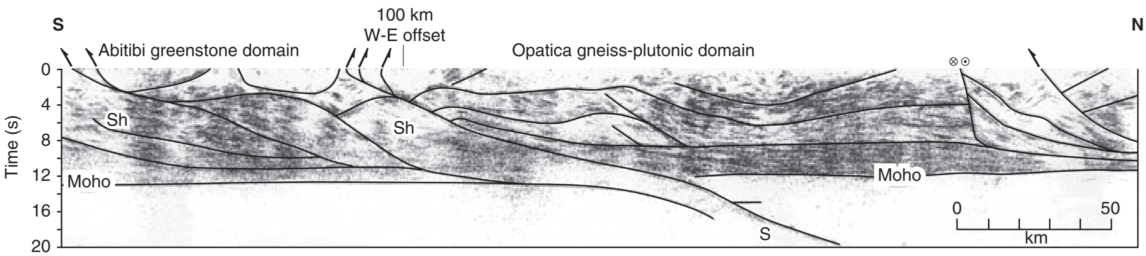


Fig. 11.3 *Seismic reflection profile of the Opatica–Abitibi belt in the Superior Province of northern Canada (modified from van der Velden et al., 2006, by permission of the American Geophysical Union. Copyright © 2006 American Geophysical Union). Interpretation is modified from Calvert et al. (1995), Lacroix & Sawyer (1995), and Calvert & Ludden (1999). S, fossil subduction zone; Sh, shingle reflections suggesting imbricated material in the middle crust.*

with these comparisons is that no chemically unaltered, complete example of Archean ocean crust is preserved. In addition, the Archean mantle was hotter by some amount than the modern mantle (Section 11.2), which undoubtedly influenced the compositions, source depths, and patterns of the volcanism (Nisbet *et al*., 1993). These problems have complicated interpretations of the processes that produced and recycled Archean crust and how they may differ from those in modern environments.

Most authors have concluded that the high magnesium contents and high degrees of melting associated with the formation of komatiites reflect melting temperatures (1400–1600°C) that are higher than those of modern basaltic magmas (Nisbet *et al*., 1993). Exactly how much hotter, however, is problematic. Parman *et al*. (2004) proposed a subduction-related origin for these rocks similar to that which produced boninites in the Izu-Bonin-Mariana island arc (Fig. 11.4). *Boninites* are high-Mg andesites that are thought to result from the melting of hydrous mantle in anomalously hot forearc regions above young subduction zones (Crawford *et al*., 1989; Falloon & Danyushevsky, 2000). If the komatiites were produced by the melting of hydrous mantle, then the depth of melting could have been relatively shallow, as in subduction zones, and the Archean mantle need only be slightly hotter (~100°C) than at present (Grove & Parman, 2004). In this interpretation, shallow melting and subduction result in the formation and thickening of highly depleted mantle lithosphere that some time later is incorporated into the cratonic mantle below a continent.

Alternatively, if the source rocks of komatiites were dry then high ambient temperatures in the Archean

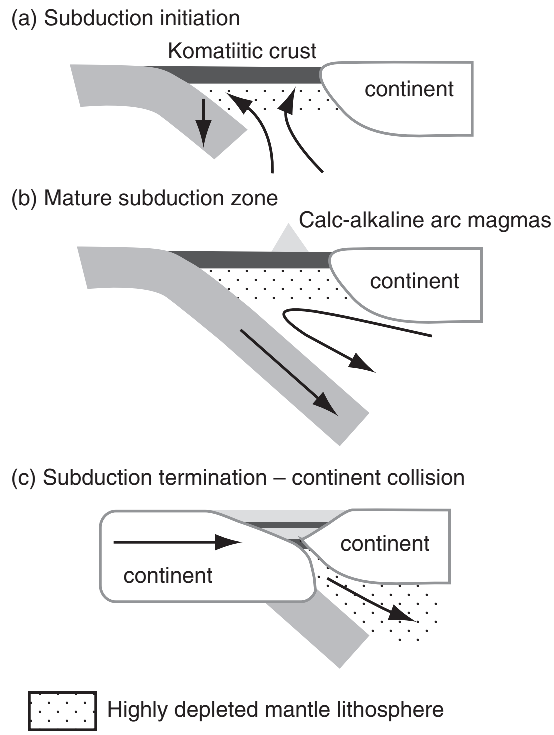


Fig. 11.4 *Conceptual model for the generation of komatiites and cratonic mantle by partial melting in a subduction zone (after Parman et al., 2004. Copyright © 2004 Geological Society of South Africa). (a) Partial melting produces komatiitic magma in a forearc setting. (b) Mature subduction cools and hydrates residual mantle. (c) Obduction of komatiitic crust occurs during collision.*

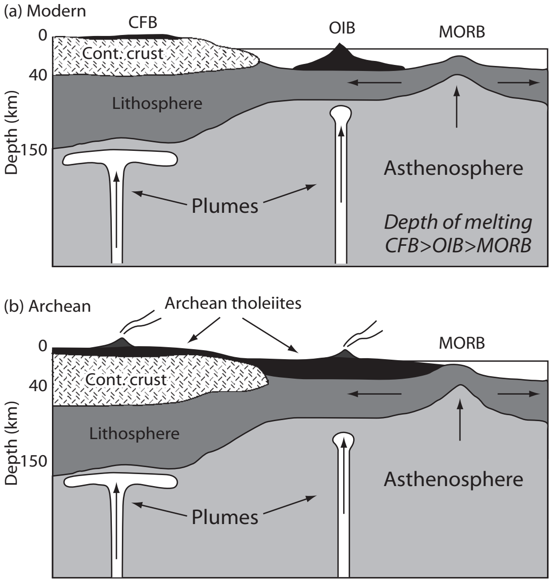


Fig. 11.5 *Model of komatiitic and tholeiitic basalt formation involving mantle plumes (after Arndt et al., 1997, by permission of Oxford University Press). Model shows the influence of lithospheric thickness on depth of melting where CFB is continental flood basalt, OIB oceanic island basalt, and MORB mid-ocean ridge basalt.*

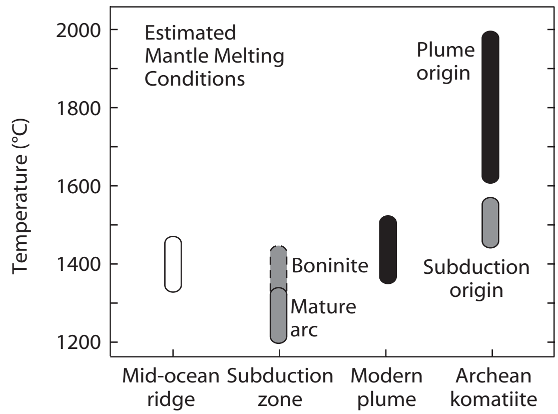


Fig. 11.6 *The range of mantle melt generation temperatures estimated for various modern tectonic settings compared to temperatures inferred for komatiite melt generation by a plume model (black filled oval) and a subduction model (gray filled oval) (after Grove & Parman, 2004, with permission from Elsevier).*

mantle would have caused melting to begin at depths that were much greater than occurs in subduction zones, possibly in upwelling mantle plumes or at unusually deep levels (~200 km) beneath mid-ocean ridges. The greater depths of melting would produce large volumes of basalt and oceanic crust that was much thicker (20–40 km) than it is today (Bickle *et al.*, 1994). Evidence of large volumes of mafic magma and high eruption rates have suggested that oceanic plateaux and continental flood basalts are the best modern analogues for such thick mafic crust and invites comparisons with Phanerozoic LIPs (Section 7.4.1) (Arndt *et al.*, 1997, 2001). In this latter context, the differences between the modern and ancient rocks are explained by variations in the depth of melting and in the effects of a thick overriding lithosphere (Fig. 11.5). These and a variety of other models (Fig. 11.6) illustrate how information on the depth and source of the melting that produced komatiites has important consequences for both the tectonic setting and the thermal evolution of the early Earth.

A variety of tectonic models also have been postulated for the origin of Archean continental crust. Windley (1981) noted the geologic and geochemical similarities between Archean tonalite-trondhjemite-granodiorite (TTG) suites and exhumed granitoids associated with Andean-type subduction zones (Section 9.8). He considered this to be an environment in which voluminous quantities of tonalite can be produced, and concluded that this represents a reasonable analogue for the formation of these rocks in Archean times. Subsequent work has led to a general consensus that these subduction models are applicable to the Late Archean. However, their applicability to Early and Middle Archean times when thick oceanic crust may have inhibited subduction is more controversial. As an alternative to subduction, Zegers & van Keken (2001) postulated that TTG suites formed by the removal and sinking of the dense, lower part of thick oceanic plateaux. The peeling away, or *delamination*, of a dense eclogite root results in uplift, extension, and partial melting to produce TTG suite magmas. This process could have returned some oceanic material into the mantle and may have accompanied collisions among oceanic terranes in Early–Middle Archean times. However, the possible absence of subduction creates a problem in that, assuming a nonexpanding Earth (Section 12.3), a high rate of formation of oceanic lithosphere during these times must have been accompanied by a mechanism by which oceanic lithosphere also was destroyed at high rates.

An important aspect of the origin of TTG suites is the type of source material that melted to produce the magma. Early petrologic studies suggested that these magmas could result from the partial melting of subducted oceanic crust in the presence of water (Martin, 1986). However, more recent work has emphasized other sources, including the lower crust of arcs and the base of thick oceanic plateaux (Smithies, 2000; Condie, 2005a). The importance of the source material is illustrated by a two-stage model proposed by Foley *et al.* (2003). This model envisages that during the Early Archean, oceanic crust was too thick to be subducted as a unit, and so its lowermost parts delaminated and melted (Fig. 11.7a). These lower roots are inferred to have been pyroxenites that were produced by the metamorphism of ultramafic cumulate layers. The melting of the pyroxenite did not favor the generation of TTG melts but produced basaltic melts instead. As the oceanic crust cooled and became thinner, a point was reached in the Late Archean when the entire crust could subduct (Fig. 11.7b). At this time hydrothermally altered crust, such as amphibolite, was introduced into subduction zones and led to the widespread formation of TTG suites. This model supports the view that the formation of the earliest continental crust requires subduction and the melting of a hydrous mafic source.

11.3.4 Crustal structure

Granite-greenstone belts display a variety of structural styles and outcrop patterns, many of which also occur in Phanerozoic orogens (Kusky & Vearncombe, 1997). Those common to both Archean and Phanerozoic belts include large tracts of metamorphosed igneous and

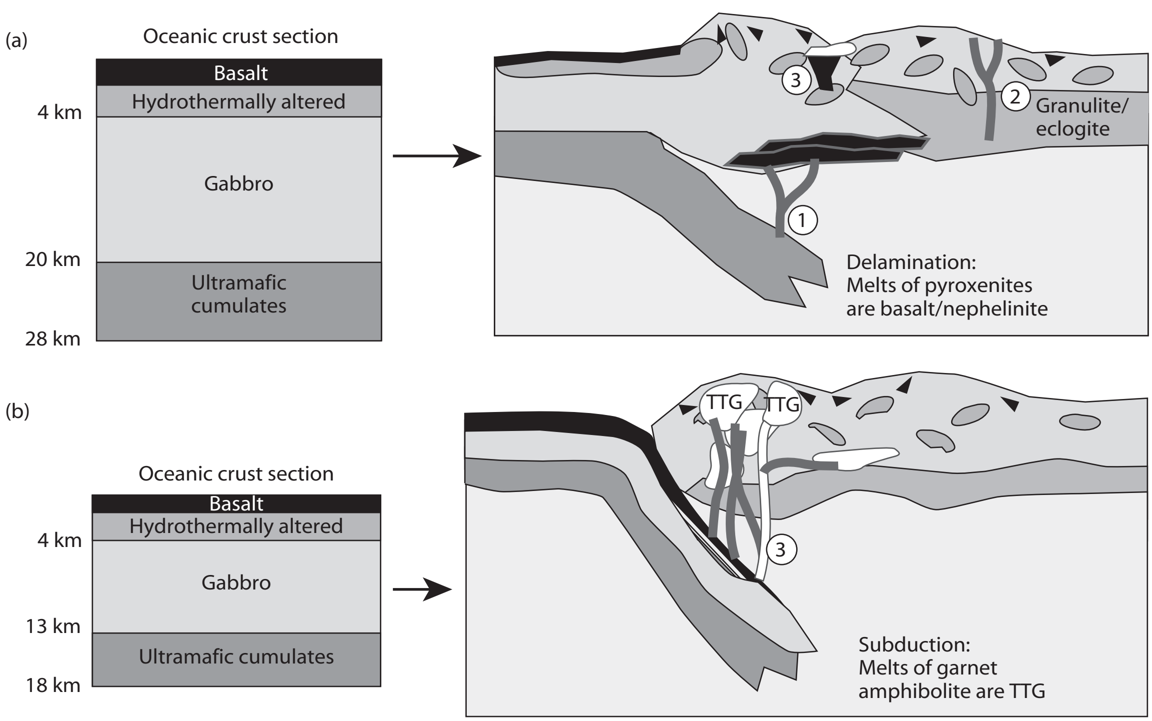


Fig. 11.7 *Two-stage model of Archean crustal evolution (after Foley et al., 2003, with permission from Nature* **421***, 249–52. Copyright © 2003 Macmillan Publishers Ltd). (a) Early Archean delamination of thick oceanic crust and the melting of pyroxenites (1). Local melting of lower crust (2) and garnet amphibolite (3) may also occur to produce small volumes of felsic magma. (b) Late Archean whole-crust subduction and the large-scale melting of garnet amphibolite to produce TTG suites (3).*

sedimentary rock that are deformed by thrust faults and strike-slip shear zones (e.g. Figs 10.13, 10.19). Another pattern, commonly referred to as a dome-and-keel architecture, occurs exclusively in Archean crust. This latter structural style forms the focus of the discussion in this section.

Dome-and-keel provinces consist of trough-shaped or synclinal keels composed of greenstone that surround ellipsoidal and ovoid-shaped domes composed of gneiss, granitoid, and migmatite (Section 9.8). The contacts between domes and keels commonly are high-grade ductile shear zones. Marshak *et al.* (1997) distinguished between two types of these provinces. One type has keels composed of greenstones and their associated metasedimentary strata (Section 11.3.2) and domes composed of granitoid rock that is similar in age or slightly younger than the greenstones. The other type has domes of mostly gneissic and migmatitic basement rock and keels composed of greenstones that are younger than the dome rocks.

The Eastern Pilbara craton of northwestern Australia (Fig. 11.8) is one of the oldest and best preserved examples of a granite-greenstone belt and dome-and-keel province with a history spanning 3.72–2.83 Ga (Collins *et al.*, 1998; Van Kranendonk *et al.*, 2002, 2007). The craton exposes nine granitoid domes with diameters ranging from 35 km to 120 km. Studies of seismic refraction data and gravity and magnetic anomalies (Wellman, 2000) indicate that the margins of the domes are generally steep and extend to mid-crustal depths of ~14 km. Despite their simple outlines, the internal structure of the domes is complex. Each contains remnants of 3.50–3.43 Ga TTG suite granitoids (Section 11.3.2) that are intruded by younger (3.33–2.83 Ga) more potassic igneous suites (Fig. 11.9a). The domes display compositional zonations and variable degrees of deformation. In many cases, the youngest bodies are located in the cores of the domes with older, more deformed granitoids at the margins. This internal structure indicates that each dome was constructed through the emplacement of many intrusions over hundreds of millions of years and that deformation accompanied the magmatism.

Between the granitoid domes are synclinal tracts of greenstone composed of dipping volcanic and sedimentary sequences up to 23 km thick (Van Kranendonk

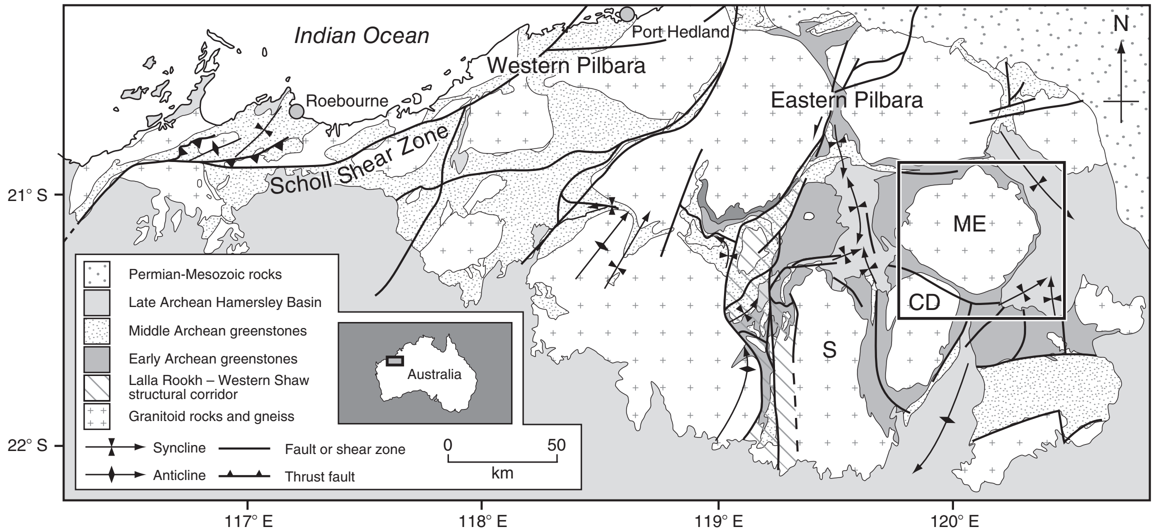


Fig. 11.8 *Geologic map of the Pilbara granite-greenstone belt (modified from Zegers & van Keken, 2001, with permission from the Geological Society of America, with additional structural information from Van Kranendonk et al., 2007) showing the typical ovoid pattern of TTG suite granitoids surrounded by greenstones belts. ME, Mt. Edgar Dome; CD, Corunna Downs Dome; S, Shaw Dome. Black box shows location of Fig. 11.9.*

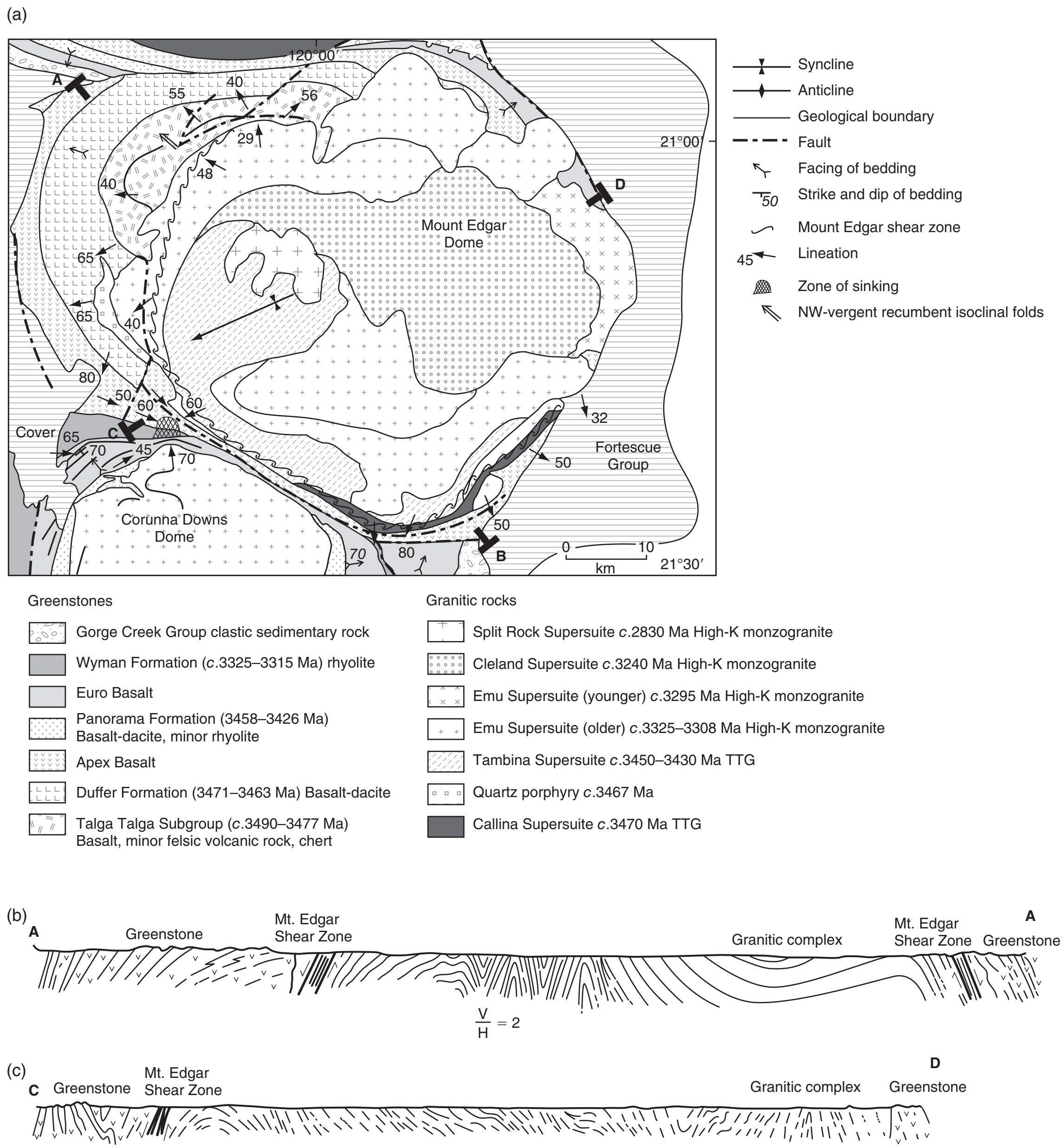


Fig. 11.9 *(a) Geologic map of the Mt. Edgar Dome and granitic complex in the Eastern Pilbara (after Van Kranendonk* et al.*, 2007, with permission from Blackwell Publishing). Map shows the internal structure of the domes and the radial distribution of 3.32–3.30 Ga metamorphic mineral elongation lineations (arrows) that converge in a vertical zone of sinking between the Mt. Edgar, Carunna Downs and Shaw (not shown) domes. These features are contemporaneous with the arcuate shear zones that formed along granite-greenstone contacts (Fig. 11.8). (b,c) Cross-sections showing the trends of foliation surfaces and shear zones within the Mt. Edgar Dome (after Collins, 1989, with permission from Elsevier).*

et al., 2002). Successive groups of these strata were deposited in autochthonous basins that developed on synclines of older greenstones. Episodes of felsic volcanism in these belts accompanied emplacement of the granitoid domes. The degree of metamorphism and the age of the strata gradually decrease away from the deformed margins of the domes and toward the cores of synclines where the greenstones are only weakly deformed. These weakly deformed areas preserve the delicate Archean *stromatolites* and other evidence of early life (Buick, 2001). The geometry of the synclines between the domes creates a high amplitude (~15 km), long wavelength (120 km) dome-and-keel structure that developed throughout the entire history of the Eastern Pilbara.

The contacts between the granitoid domes and the greenstones in the Eastern Pilbara vary from being intrusive to unconformities, ring faults and high grade shear zones. The shear zones and ring faults are concentric about the domes and generally display steep to subvertical orientations (Figs 11.8, 11.9b,c). Many of these shear zones, including the Mt. Edgar Shear Zone, formed during the period 3.32–3.30 Ga (Van Kranendonk *et al.*, 2007). The central part of the craton contains a 5- to 15-km-wide zone of ductile deformation called the Lalla Rookh–Western Shaw structural corridor (Fig. 11.8). This zone formed during a period (~2.94 Ga) of contraction and is characterized by multiple generations of folds and ductile rock fabrics (Van Kranendonk & Collins, 1998).

11.3.5 Horizontal and vertical tectonics

The origin of the unique dome-and-keel architecture of the Archean cratons (Section 11.3.4) is important for understanding the nature of Archean tectonics. In general, interpretations can be divided into contrasting views about the relative roles of vertical and horizontal displacements in producing this pattern. The Eastern Pilbara craton in western Australia illustrates how vertical and horizontal tectonic models have been applied to explain the dome-and-keel structural style. During this discussion, it is important to keep in mind that the crustal structure, as illustrated by the Pilbara example, is the product of multiple episodes of deformation, metamorphism, and pluton emplacement rather than a single tectonic episode.

Vertical tectonic models describe the diapiric rise of hot granitoid domes as the result of a partial convective overturning of the middle and upper crust. Collins *et al.* (1998) and Van Kranendonk *et al.* (2004) used strain patterns, a dome-side-up/greenstone-side-down sense of displacement in shear zones, and other features to link the formation of dome-and-keel structures to a sinking of the greenstones. The process begins with the emplacement of hot TTG suite (Section 11.3.2) granitoids into an older greenstone succession (Fig. 11.10a). Domes are initiated at felsic volcanic centers due to a laterally uneven emplacement of TTG magma. After a hiatus of several tens of millions of years, the emplacement of thick piles of basalt on top of less dense granitoids creates an inverted crustal density profile (Fig. 11.10b). The magmatism also buries the granitoids to mid-crustal depths where they partially melt due to the build up of radiogenic heat and, possibly, the advection of heat from mantle plume activity. Thermal softening and a reduction in mid-crustal viscosity facilitates the sinking of the greenstones, which then squeezes out the underlying partial melts into rising, high-amplitude granitoid domes (Fig. 11.10c). The convective overturning depresses geotherms in the greenstone tracts, resulting in local cooling and the preservation of kyanite-bearing metamorphic rocks, which equilibrate at moderate-low pressures (~600 MPa) and temperatures (500°C). This model explains the formation of the dome-and-keel structure without rigid plates or plate boundary forces and is similar to the sinking or *sagduction* models proposed for the formation of dome-and-keel structures in the Dharwar craton of India (Chardon *et al.*, 1996).

Horizontal tectonic models for the Eastern Pilbara propose that the greenstones were affected by one or more periods of horizontal contraction and extension (Blewitt, 2002). In these interpretations, the contraction results from episodes of Early Archean collision (Sections 10.4, 10.5) and terrane accretion (Section 10.6). Periods of horizontal extension result in the formation of crustal detachments and the emplacement of the granitoid domes. Kloppenburg *et al.* (2001) used observations of multiple cross cutting fabrics and unidirectional patterns of stretching lineations to suggest that the Mt. Edgar Dome initially formed as an extensional metamorphic core complex (Sections 7.3, 7.6.3, 7.6.6). An initial period of terrane collision and thrusting prior to 3.32 Ga thickens the Early Archean Warrawoona Greenstone Belt and buries granitoid basement to mid-crustal levels where it partially melts. Partial melting

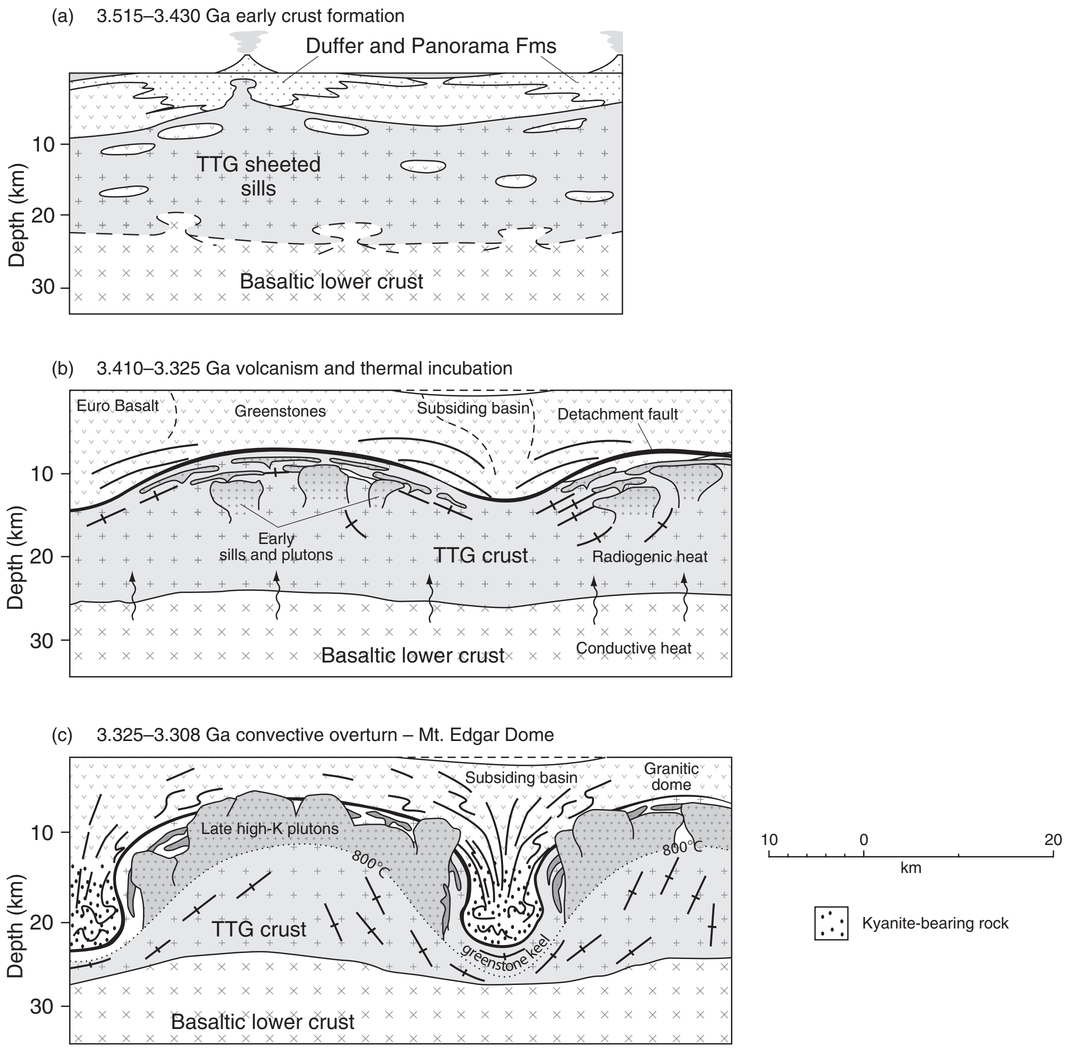


Fig. 11.10 *Three-stage diapiric model of dome-and-keel provinces in the Eastern Pilbara craton (after Collins* et al.*, 1998, with permission from Pergamon Press, Copyright Elsevier 1998; the age-spans of the stages are from Van Kranendonk* et al.*, 2007).*

facilitates the extensional collapse of the thickened crust, forming detachment faults (i.e. Mt Edgar shear Zone, Fig. 11.11a) similar to those found in Phanerozoic core complexes (e.g. Figs. 7.14b, 7.39c). The density inversion created by dense greenstones overlying buoyant, partially molten basement triggers the rise of granitoid domes at 3.31 Ga by solid state flow during extension (Fig. 11.11b). This extension is accommodated by displacement on the Mt. Edgar shear zone and by lateral strike-slip motion in a transfer zone within the gneissic basement. Normal-sense displacements drop greenstones down between the rising domes. Emplacement of the domes steepens the detachments and changes the geometry of the system so that its structure no longer resembles that of Phanerozoic core complexes (Fig. 11.11c). Steepening during periods of shortening provides an alternative explanation of the near vertical sides of the granitoid domes.

The application of both vertical and horizontal models to Archean cratons involves numerous uncer-

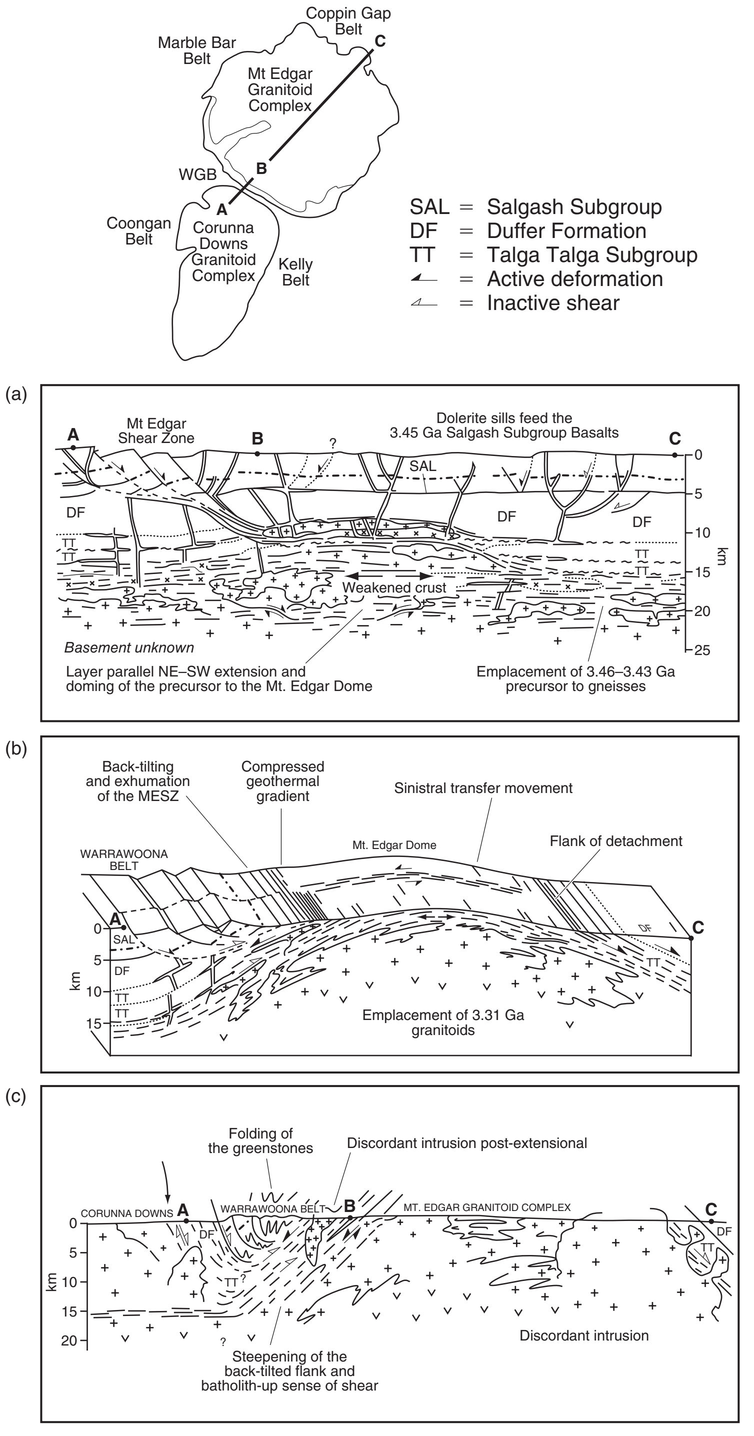


Fig. 11.11 *Cartoon summarizing the tectonic and magmatic development of the Warrawoona Greenstone Belt (WGB) and Mt. Edgar Dome by horizontal extension (after Kloppenburg* et al.*, 2001, with permission from Elsevier). (a) Pre 3.33 Ga gabbro/diorite and dolerite intrusions, NE–SW extension, and doming of the Mt. Edgar Granitoid Complex. (b) Differential extension on the Mt. Edgar Shear Zone (MESZ) and lateral motion along a transfer zone within the granitoid complex at 3.31 Ga. (c) Final localized normal displacements and steepening of the MESZ followed by discordant intrusions of post-extensional plutons.*

tainties. Problems with diapiric models commonly include uncertainties surrounding the timing of convective overturn and how an inverted density profile could be maintained or periodically established over a 750 million year history without thrust faulting (Van Kranendonk *et al.*, 2004). How the stiff rheology of granitoids allows diapirism also is unclear. Problems with horizontal tectonic models may include a lack of evidence of large-scale tectonic duplication of the greenstones by thrusts in some areas and uncertainties surrounding how the formation of metamorphic core complexes could produce the distinctive ovoid patterns of the granitoids. Horizontal tectonic models also commonly encounter difficulty explaining the kinematics and horseshoe-shaped geometry of shear zones that border many granitoid domes (Marshak, 1999).

A comparison of the evolution of various Archean cratons has suggested that aspects of both horizontal and vertical tectonic processes occurred in different places and at different times. Hickman (2004) highlighted numerous tectonic and metamorphic differences between the Eastern and Western parts of the Pilbara craton prior to ~2.95 Ga. He showed that, unlike the more or less autochthonous dome-and-keel structure of the Eastern Pilbara, the Western Pilbara preserves a series of amalgamated terranes (Section 10.6.1) that are separated by a series of thrusts and strike-slip shear zones (Fig. 11.8) and involved episodes of horizontal compression that resemble a Phanerozoic style of plate tectonics. These differences suggest that both vertical and horizontal tectonics played an important role during the formation of the Pilbara craton.

11.4 PROTEROZOIC TECTONICS

11.4.1 General geology of Proterozoic crust

Proterozoic belts display two groups of rocks that are distinguished on the basis of their metamorphic grade and deformation history. The first group consists of thick sequences of weakly deformed, unmetamorphosed sedimentary and volcanic rocks that were deposited in large basins on top of Archean cratons. The second group is composed of highly deformed, high-grade metamorphic rocks that define large orogenic belts. Both these groups contain distinctive suites of igneous rocks.

The most common lithologic assemblage in the weakly deformed parts of Early–Middle Proterozoic crust are quartzite-carbonate-shale sequences that reach thicknesses of some 10 km (Condie, 1982b). Quartz-pebble conglomerates and massive, cross-bedded sandstones also are common. Many of these sequences are intercalated with banded iron formations, cherts, and volcanic rocks. Other rock types that are either rare or absent in Archean belts appeared at this time, including extensive evaporites, phosphorous-rich sedimentary sequences, and red bed deposits (Section 3.4). These latter rocks generally are interpreted to have accumulated in stable, shallow water environments after 2.0 Ga. The appearance and the preservation of such thick sequences of sedimentary rock has been interpreted to reflect the stabilization of Precambrian continental crust during Proterozoic times (Eriksson *et al.*, 2001, 2005) (Section 11.4.2). In the Pilbara region of northwest Australia (Fig. 11.8) the deposition of 2.78–2.45 Ga coarse clastic sedimentary rocks and volcanic sequences in a shallow platform environment in the Hamersley Basin (Trendall *et al.*, 1991) reflects this stabilization. By 1.8 Ga, the existence of large, stable landmasses and free oxygen in the Earth's atmosphere allowed all of the well-known sedimentary environments that characterize Phanerozoic times to develop (Eriksson *et al.*, 2005).

The highly deformed regions of Proterozoic crust are divisible into two types (Kusky & Vearncombe, 1997). The first type consists of thick sedimentary sequences that were deformed into linear fold-and-thrust belts similar to those in Phanerozoic orogens (Figs 10.5, 10.19). The second type consists of high-grade gneisses of the granulite and upper amphibolite facies. Some of the largest and best known of these latter belts form the ~1.0 Ga Grenville provinces of North America, South America, Africa, Antarctica, India, and Australia (Fig. 11.19). Other belts (Fig. 11.12) evolved during the period 2.1–1.8 Ga (Zhao *et al.*, 2002). These orogens contain large ductile thrust zones that separate distinctive terranes. Some contain ophiolites (Section 2.5) that resemble Phanerozoic examples except for the lack of highly deformed mantle-derived rocks at their bases in ophiolites older than ~1 Ga (Moores, 2002). The presence of these features reflects the importance of subduction, collision, and terrane accretion along Proterozoic continental margins (Carr *et al.*, 2000; Karlstrom *et al.*, 2002).

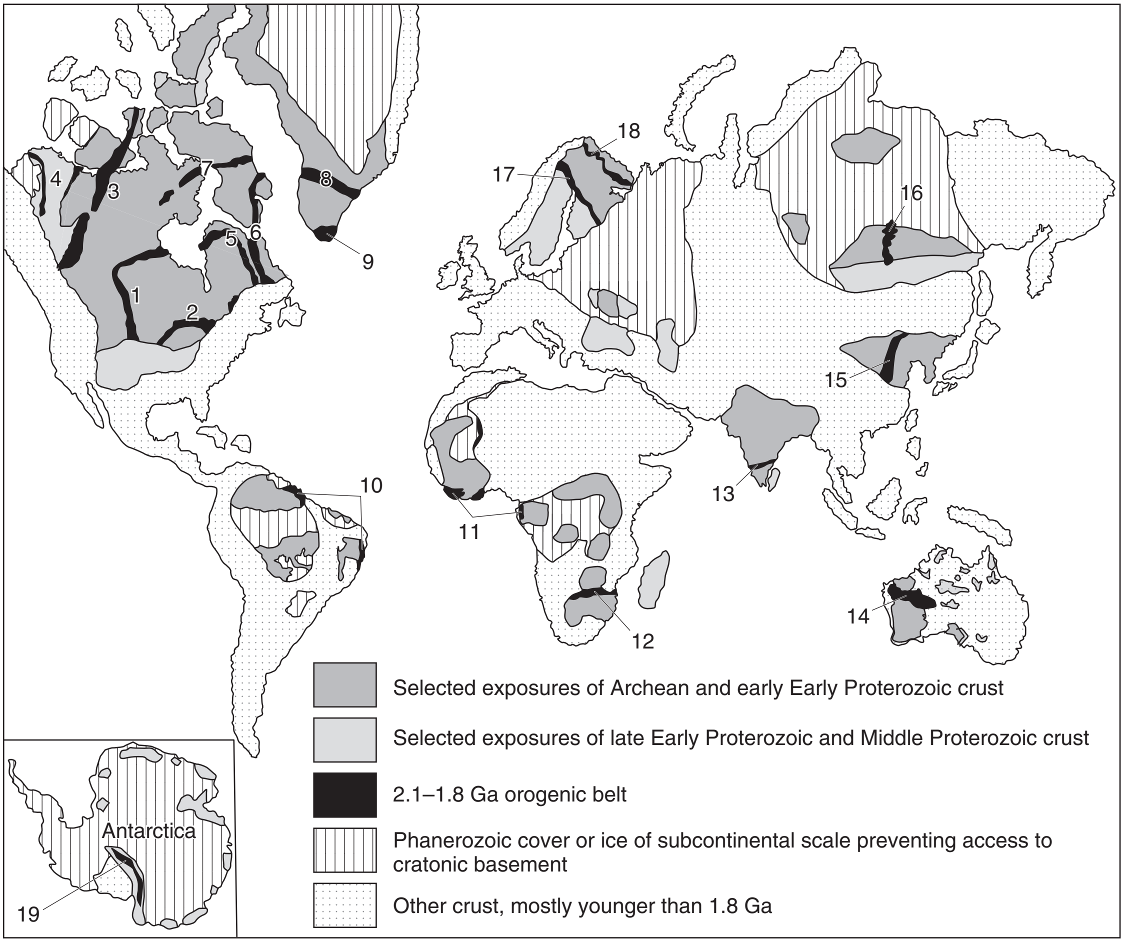


Fig. 11.12 *Global distribution of 2.1–1.8 Ga orogenic belts showing selected areas of Archean and Early Proterozoic basement (after Zhao et al., 2002, with permission from Elsevier). Orogens labeled as follows: 1, Trans-Hudson; 2, Penokean; 3, Taltson-Thelon; 4, Wopmay; 5, Cape Smith–New Quebec; 6, Torngat; 7, Foxe; 8, Nagssugtoqidian; 9, Makkovikian–Ketilidian; 10, Transamazonian; 11, Eburnian; 12, Limpopo; 13, Moyar; 14, Capricorn; 15, Trans-North China; 16, Central Aldan; 17, Svecofennian; 18, Kola-Karelian; 19, Transantarctic.*

A comparison of igneous rocks in Archean and Proterozoic belts indicates a progressive change in the bulk composition of the crust through time (Condie, 2005b). During the Early Archean, basaltic rocks were most abundant (Section 11.3.2). Later, the partial melting of these rocks in subduction zones or at the base of oceanic plateaux produced large volumes of TTG suite granitoids (Sections 11.3.2, 11.3.3). By 3.2 Ga granites first appeared in the geologic record and were produced in large quantities after 2.6 Ga. This compositional trend from basalt to tonalite to granite generally is attributed to an increase in the importance of subduction and crustal recycling during the transition from Late Archean to Early Proterozoic times.

Large swarms of mafic dikes were emplaced into Archean cratons and their cover rocks during the Late Archean–Early Proterozoic and onwards. One of the best exposed examples of these is the 1.27 Ga MacKenzie dike swarm of the Canadian Shield, which exhibits dikes that fan out over a 100° arc and extend for more than 2300 km (Ernst *et al.*, 2001). Some of these shield regions also contain huge sills and layered intrusions of mafic and ultramafic rock that occupy hundreds to thousands of square kilometers. These intrusions provide information on the deep plumbing systems of Precambrian magma chambers and on crust–mantle interactions. Three of the best known examples are the ~1.27 Ga Muskox intrusion in northern Canada (Le Cheminant & Heaman, 1989; Stewart & DePaolo, 1996), the ~2.0 Ga Bushveld complex in South Africa (Hall, 1932; Eales & Cawthorn, 1996), and the ~2.7 Ga Stillwater complex in Montana, USA (Raedeke & McCallum, 1984; McCallum, 1996). Unlike the layered igneous suites of the Archean high-grade gneiss terrains (Section 11.3.2), these intrusions are virtually undeformed.

Anorthosite massifs (Section 11.3.2) emplaced during Proterozoic times also differ from the Archean examples. Proterozoic anorthosites are associated with granites and contain less plagioclase than the Archean anorthosites (Wiebe, 1992). These rocks form part of an association known as anorthosite-mangerite-charnockite-granite (AMCG) suites. *Charnockites* are high temperature, nearly anhydrous rocks that can be of either igneous or high-grade metamorphic origin (Winter, 2001). The source of magma and the setting of the anorthosites are controversial. Most studies interpret them as having crystallized either from mantle-derived melts that were contaminated by continental crust (Musacchio & Mooney, 2002) or as primary melts derived from the lower continental crust (Schiellerup *et al.*, 2000). Current evidence favors the former model. Some authors also have suggested that these rocks were emplaced in rifts or backarc environments following periods of orogenesis, others have argued that they are closely related to the orogenic process (Rivers, 1997). Their emplacement represents an important mechanism of Proterozoic continental growth and crustal recycling.

11.4.2 Continental growth and craton stabilization

Many of the geologic features that comprise Proterozoic belts (Section 11.4.1) indicate that the continental lithosphere achieved widespread tectonic stability during this Eon. Tectonic stability refers to the general resistance of the cratons to large-scale lithospheric recycling processes. The results of seismic and petrologic studies (Sections 11.3.1, 11.3.3) and numerical modeling (Lenardic *et al.*, 2000; King, 2005) all suggest that compositional buoyancy and a highly viscous cratonic mantle explain why the cratons have been preserved for billions of years. These properties, and isolation from the deeper convecting mantle, have allowed the mantle lithosphere to maintain its mechanical integrity and to resist large-scale subduction, delamination and/or erosion from below. Phanerozoic tectonic processes have resulted in some recycling of continental lithosphere (e.g. Sections 10.2.4, 10.4.5, 10.6.2), however the scale of this process relative to the size of the cratons is small.

The cores of the first continents appear to have reached a sufficient size and thickness to resist being returned back into the mantle by subduction or delamination some 3 billion years ago. Collerson & Kamber (1999) used measurements of Nb/Th and Nb/U ratios to infer the net production rate of continental crust since 3.8 Ga. This method exploits differences in the behavior of these elements during the partial melting and chemical depletion of the mantle. The different ratios potentially provide information on the extent of the chemical depletion and the amount of continental crust that was present on Earth at different times. This work and the results of isotopic age determinations (Fig. 11.13) suggest that crust production was episodic with rapid net growth at 2.7, 1.9, and 1.2 Ga and slower growth afterward (Condie, 2000; Rino *et al.*, 2004). Each of these pulses may have been short, lasting ≤100 Ma

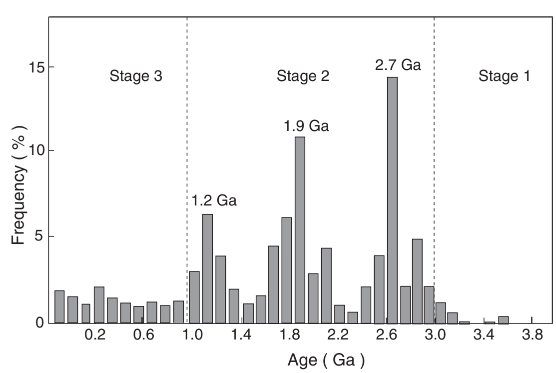


Fig. 11.13 *Plot showing the distribution of U-Pb zircon ages in continental crust (after Condie, 1998, with permission from Elsevier).*

(Condie, 2000). On the basis of available data, Condie (2005b) concluded that 39% of the continental crust formed in the Archean, 31% in the Early Proterozoic, 12% in the Middle–Late Proterozoic, and 18% in the Phanerozoic.

Two of the most important mechanisms of Late Archean and Early Proterozoic continental growth and cratonic root evolution were magma addition (Section 9.8) and terrane accretion (Section 10.6). Several authors (e.g. Condie, 1998; Wyman & Kerrich, 2002) have suggested that the ascent of buoyant mafic material in mantle plumes may have initiated crust formation and may have either initiated or modified root formation during periods of rapid net growth (Section 11.3.3). Schmitz *et al.* (2004) linked the formation and stabilization of the Archean Kaapvaal craton in South Africa to subduction, arc magmatism, and terrane accretion at 2.9 Ga. In this and most of the other cratons, isotopic ages from mantle xenoliths and various crustal assemblages indicate that chemical depletion in the mantle lithosphere was coupled to accretionary processes in the overlying crust (Pearson *et al.*, 2002). This broad correspondence is strong evidence that the crust and the underlying lithospheric mantle formed more or less contemporaneously and have remained mechanically coupled since at least the Late Archean. A progressive decrease in the degree of depletion in the lithospheric mantle since the Archean (Fig. 11.14) indicates that the Archean–Proterozoic boundary represents a major shift in the nature of lithosphere-forming processes, with more gradual changes occurring during the Phanerozoic (O'Reilly *et al.*, 2001). The most obvious driving mechanism of this change is the secular cooling of the Earth (Section 11.2). In addition, processes related to subduction, collision, terrane accretion, and magma addition also helped to form and stabilize the continental lithosphere.

Whereas these and many other investigations have identified some of the processes that contributed to the formation and stabilization of Archean cratons, numerous questions still remain. Reconciling the composition of craton roots determined from petrologic studies with the results of seismic velocity studies is problematic (King, 2005). There are many uncertainties about how stability can be achieved for billions of years without suffering mechanical erosion and recycling in the presence of subduction and mantle convection. Another problem is that the strength of mantle materials required to stabilize craton roots in numerical experiments exceeds the strength estimates of these materials

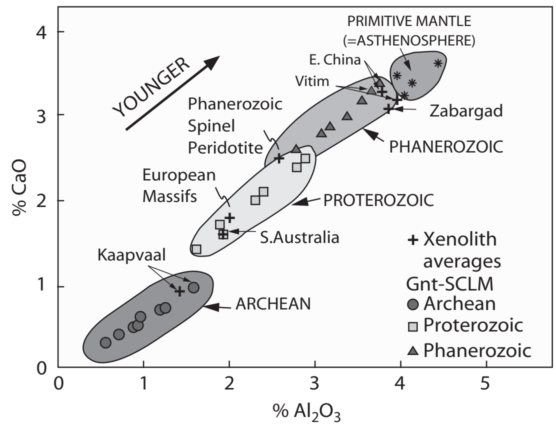


Fig. 11.14 *CaO-Al_2O_3 plot showing the range of subcontinental lithospheric mantle (SCLM) compositions for selected cratons that have been matched with ages of the youngest tectonothermal events in the overlying crust (after O'Reilly* et al.*, 2001, with permission from the Geological Society of America). Compositions have been calculated from garnet xenocrysts (Gnt). Xenolith averages shown for comparison. Plot shows that newly formed subcontinental lithospheric mantle has become progressively less depleted in Al and Ca contents from Archean through Proterozoic and Phanerozoic time. Garnet peridotite xenoliths from young extensional areas (e.g. eastern China, Vitim in the Baikal region of Russia, and Zabargad Island in the Red Sea) are geochemically similar to primitive mantle, indicating very low degrees of melt depletion.*

derived from laboratory measurements (Lenardic *et al.*, 2003). These issues, and the extent to which the cratonic mantle interacts with and influences the pattern of mantle convection, presently are unsolved. Improved resolution of the structure, age and geochemical evolution of the continental crust and lithospheric mantle promise to help geoscientists resolve these problems in the future.

11.4.3 Proterozoic plate tectonics

Early tectonic models of Proterozoic lithosphere envisaged that the Archean cratons were subdivided by mobile belts in which deformation was wholly ensialic, with no rock associations that could be equated with ancient ocean basins. These interpretations where Proterozoic orogenies occurred far from continental margins have since fallen out of favor. Most studies now indicate that Proterozoic orogens (Fig. 11.12) evolved along the margins of lithospheric plates by processes that were similar to those of modern plate tectonics.

One of the best-studied examples of an Early Proterozoic orogen that formed by plate tectonic processes lies between the Slave craton and the Phanerozoic Canadian Cordillera in northwestern Canada. This region provides a record of nearly 4 billion years of lithospheric development (Clowes *et al.*, 2005). Deep seismic reflection data collected as part of the Lithoprobe SNORCLE (**S**lave-**Nor**thern **C**ordillera **L**ithospheric **E**volution) transect of the Canadian Shield (see also Section 10.6.2) provide evidence of a modern, plate tectonic-style of arc–continent collision, terrane accretion, and subduction along the margin of the Archean Slave craton between 2.1 and 1.84 Ga (Cook *et al.*, 1999). These processes formed the Early Proterozoic Wopmay Orogen (Fig. 11.15a) and resulted in continental growth through the addition of a series of magmatic arcs, including the Hottah and Fort Simpson terranes and the Great Bear magmatic arc.

Final assembly of the Slave craton occurred by ~2.5 Ma. Cook *et al.* (1999) suggested that low-angle seismic reflections beneath the Yellowknife Basin (Fig. 11.15b) represent surfaces that accommodated shortening during this assembly. Some of these reflections project into the upper mantle and represent the remnants of an east-dipping Late Archean subduction zone. Following amalgamation of the craton, the Hottah terrane formed as a magmatic arc some distance outboard of the ancient continental margin between 1.92 and 1.90 Ga. During the Calderian phase (1.90–1.88 Ga) of the Wopmay Orogen this arc terrane collided with the Slave craton, causing compression, shortening, and the eastward translation of exotic material (Fig. 11.15c). In the seismic profile (Fig. 11.15b), the accreted Proterozoic crust displays gently folded upper crustal layers overlying reflectors that appear to be thrust slices above detachment faults that flatten downward into the Moho. Remnants of the old, east-dipping subduction zone associated with the collision are still visible today as reflections that project to 200 km or more beneath the Slave craton.

Once accretion of the Hottah terrane terminated, the subduction of oceanic lithosphere to the east beneath the continental margin created the 1.88–1.84 Ga Great Bear Magmatic arc and eventually led to the collision of the Fort Simpson terrane some time before 1.71 Ga (Fig. 11.15c,d). Mantle reflections that record subduction and shortening during this arc–continent collision dip eastward beneath the Great Bear magmatic arc from the lower crust to depths of 100 km (Fig. 11.15b,c,d). Where the mantle reflections flatten into the lower crust, they merge with west-dipping crustal reflections, producing a lithospheric-scale accretionary wedge that displays imbricated thrust slices. This faulted material and the underthrust lower crust represent part of a Early Proterozoic subduction zone that bears a remarkable resemblance to structures that record subduction and accretion within the Canadian Cordillera (Fig. 10.33) and along the Paleozoic margin of Laurentia (Fig. 10.34). Seismic refraction and wide-angle reflection data (Fernández-Viejo & Clowes, 2003) indicate the presence of unusually high velocity (7.1 km s^{-1}) lower crust and unusually low velocity (7.5 km s^{-1}) upper mantle in this zone (Fig. 11.16c) compared to other parts of this section (Fig. 11.16a,b,d). This observation indicates that the effects of collision, subduction, and the accompanying physical changes in rocks of the mantle wedge remain identifiable 1.84 billion years after they formed.

In western Australia, distinctive patterns of magnetic anomalies provide direct evidence for the collision and suturing of the Archean Yilgarn and Pilbara cratons beginning by ~2.2 Ga (Cawood & Tyler, 2004). The Capricorn Orogen (Fig. 11.17a,b) is composed of Early Proterozoic plutonic suites, medium- to high-grade metamorphic rocks, a series of volcano-sedimentary basins, and the deformed margins of the two Archean cratons. Late Archean rifting and the deposition of passive margin sequences at the southern margin of the Pilbara craton is recorded by the basal sequences of the Hamersley Basin. Following rifting between the cratons, several major pulses of contractional deformation and metamorphism took place during the intervals 2.00–1.96 Ga, 1.83–1.78 Ga, and 1.67–1.62 Ga. These events resulted in basin deformation and the juxtaposition of cratons of different age and structural trends (Fig. 11.17b,c). The episodic history of rifting followed by multiple episodes of contraction and collision corresponds to at least one and probably two Wilson cycles (Section 7.9) involving the opening and closing of Late Archean–Early Proterozoic ocean basins (Cawood & Tyler, 2004). The presence of similar collisional orogens in Laurentia, Baltica, Siberia, China, and India suggests that the early to mid-Early Proterozoic marks a period

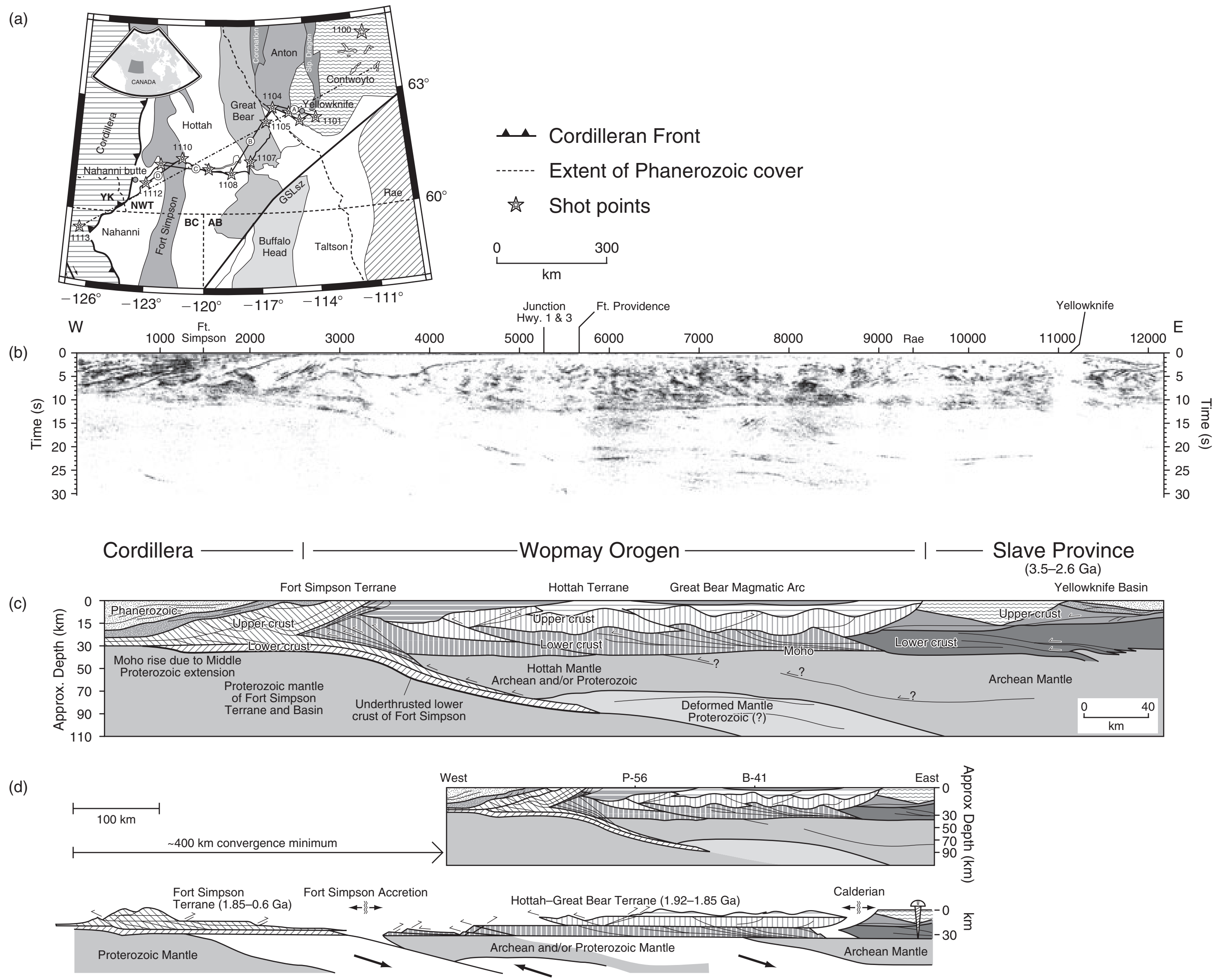


Fig. 11.15 *(a) Map showing tectonic elements of the Wopmay Orogen and location of the SNORCLE transect (after Fernández-Viejo & Clowes, 2003, with permission from Blackwell Publishing). Straight black lines (circled letters) show division used for interpretation of crustal structure. Orogen includes Coronation, Great Bear, Hottah, Fort Simpson and Nahanni domains. BC, British Colombia; AB, Alberta; YK, Yukon; NWT, Northwest Territories; GSLsz, Great Slave Lake shear zone. (b) Seismic profile, (c) interpretation and (d) reconstruction of the Wopmay Orogen and Slave Province (modified from Cook* et al.*, 1999, by permission of the American Geophysical Union. Copyright © 1999 American Geophysical Union). Reconstruction is made along major faults (bold black lines) and shows a minimum estimate.*

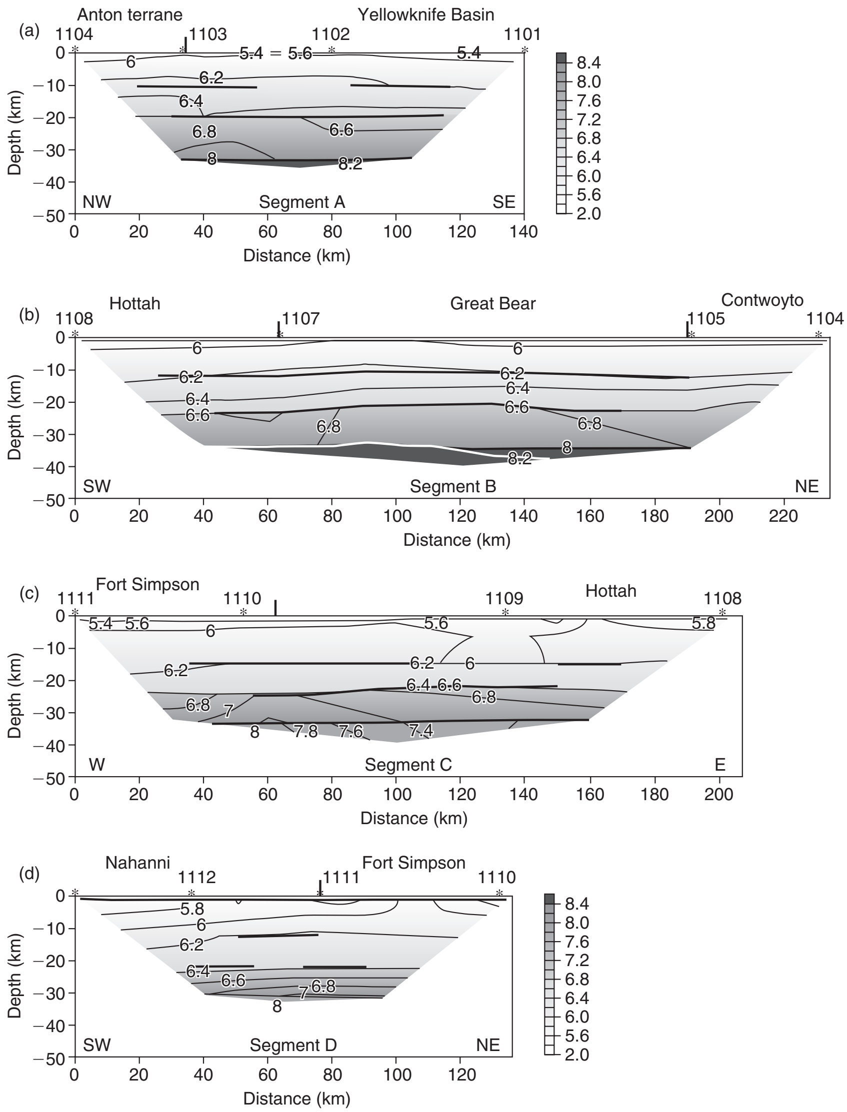


Fig. 11.16 *P-wave velocity model for (a) Slave province, (b) eastern part of the Wopmay Orogen and transition to Slave Province, (c) Hottah and Fort Simpson terranes, and (d) Fort Simpson terrane and Nahanni domain (after Fernández-Viejo & Clowes, 2003, with permission from Blackwell Publishing). Line segments shown in Fig. 11.15a. Contours are drawn at 0.2 km s^{-1} increments. Heavy black lines are locations from which wide-angle reflections were observed.*

of supercontinent assembly by plate tectonic processes (Fig. 11.12) (Section 11.5.4).

The appearance of the metamorphic products of subduction and continental collision during the Early Proterozoic, including eclogites and other >1 GPa high-pressure metamorphic assemblages (e.g. Sections 9.9, 10.4.2), represents an important marker of the onset of tectonic processes similar to those seen in the Phanerozoic Earth (O'Brien & Rötzler, 2003; Collins *et al.*, 2004; Brown, 2006). Phanerozoic eclogites commonly preserve evidence of having been partially subducted to depths greater than ~50 km and then rapidly exhumed

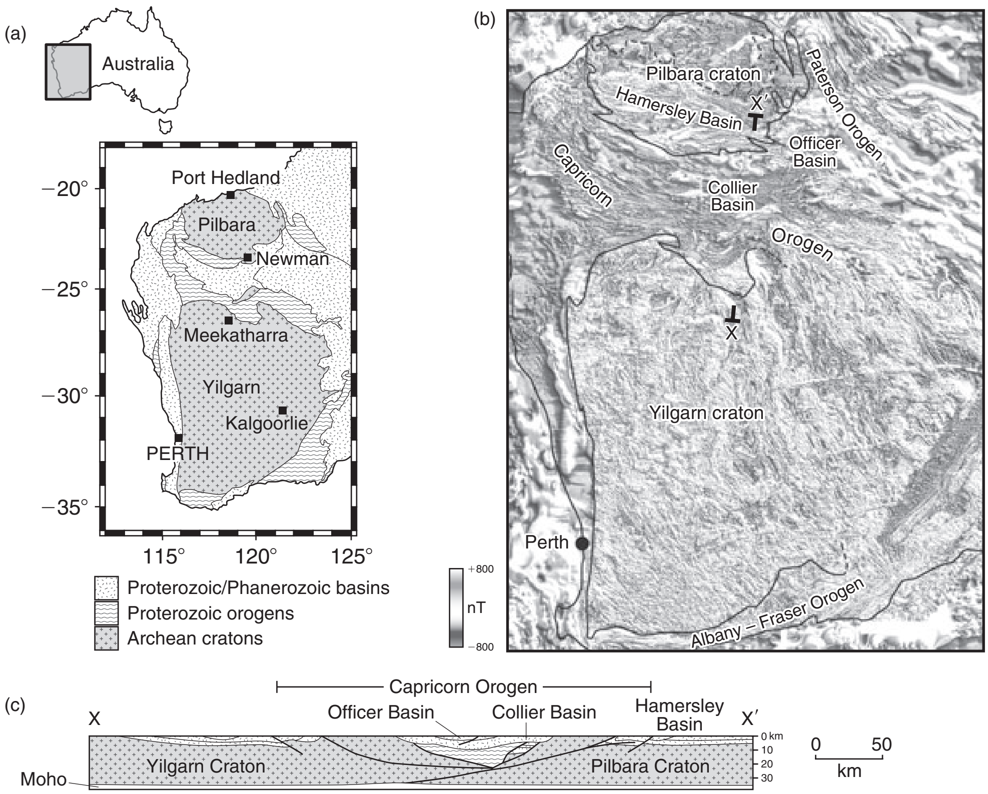


Fig. 11.17 *(a) Tectonic map (after Hackney, 2004, with permission from Elsevier) and (b) magnetic anomaly image emphasizing gradients in total magnetic intensity (after Kilgour & Hatch, 2002, with permission from Geoscience Australia, image provided by M. Van Kranendonk, Geological Survey of Western Australia). Magnetic anomaly image shows total magnetic intensity measured in nanoteslas (nT) compiled from airborne, marine and land-based geophysical surveys. The distinctive magnetic anomaly patterns from the Pilbara and Yilgarn cratons reflect different structural trends that resulted from Precambrian plate tectonic processes. (c) Interpretive cross-section of the Capricorn Orogen showing the sutured cratons (after Cawood & Tyler, 2004, with permission from Elsevier).*

(Fig. 11.18) (Collins *et al.*, 2004). In the modern Earth, these unusual conditions are met in subduction–accretion complexes and at the sites of continental collision where relatively cold crust is buried to subcrustal depths. For these subducted rocks to return to the Earth's surface with preserved eclogite facies mineral assemblages, they must be exhumed rapidly before tectonically depressed isotherms can re-equilibrate and overprint the assemblages with higher temperature granulite facies minerals.

The oldest examples of in-situ eclogites (i.e. rocks other than xenoliths) include 1.80 Ga and 2.00 Ga varieties (<1.8 GPa, 750°C) from the North China craton and the Proterozoic orogens surrounding the Tanzanian craton (Zhao *et al.*, 2001; Möller *et al.*, 1995). Eclogites recording conditions of ~1.2 GPa and 650–700°C at 1.9–1.88 Ga also occur in the Lapland Granulite Belt of Finland (Tuisku & Huhma, 1998). The Aldan Shield in Siberia (Smelov & Beryozkin, 1993) and the Snowbird tectonic zone between the Rae and Hearne cratons of Canada (Baldwin *et al.*, 2003) preserve retrogressed eclogites of 1.90 Ga. The absence of Archean eclogite facies rocks suggests that before Early Proterozoic times either the conditions to produce such rocks did not exist, the processes to exhume them at a sufficient rate to preserve eclogite-facies mineral assemblages did not exist, or all pre-existing examples have been obliterated by subsequent tectonic events.

The presence of ophiolitic assemblages in Precambrian orogens provides another possible marker of tec-

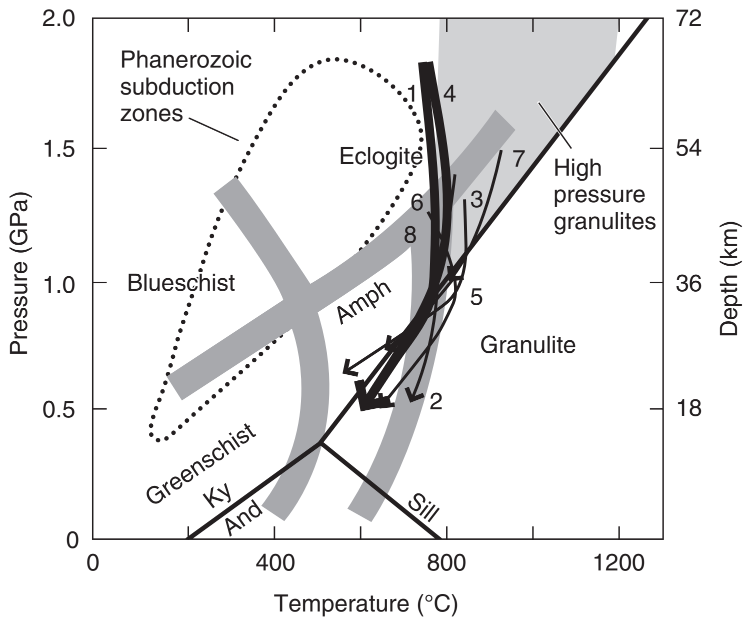


Fig. 11.18 *Pressure–temperature plot showing the published paths of the moderate temperature eclogite and high-pressure granulite facies metamorphic rocks older than 1.5 Ga (after Collins et al., 2004, with permission from Elsevier). 1, Usagaran Belt, Tanzania; 2, Hengshan Belt, China; 3, Sanggan Belt, China; 4, Ubendian Belt, Tanzania; 5, Jianping Belt, China; 6, Sare Sang, Badakhshan Block, Afghanistan; 7, Snowbird tectonic zone, Canada; 8, Lapland Granulite Belt. Thick arrows refer to Usagaran/Ubende eclogites. Field for Phanerozoic subduction zone metamorphism and metamorphic facies indicated by thick gray curves. Aluminosilicate polymorph fields plotted for reference. And, andalusite; Ky, kyanite; Sill, sillimanite; Amph, amphibolite.*

tonic processes similar to those operating during the Phanerozoic (de Wit, 2004; Parman *et al.*, 2004). Some Archean greenstone belts have been interpreted as ophiolites, although these interpretations typically are controversial. The oldest unequivocal examples are Early Proterozoic in age and support interpretations that seafloor spreading and associated ocean crust formation was an established mechanism of plate tectonics by this time. One of the best preserved and least equivocal of the Early Proterozoic examples is the Purtuniq Complex (Scott *et al.*, 1992; Stern *et al.*, 1995) of the Trans-Hudson Orogen between the Hearne and Superior cratons of northern Canada (Plate 11.1a between pp. 244 and 245). Other examples occur in the Arabian-Nubian Shield (Kröner, 1985) and the Yavapai-Mazatzal orogens of the southwestern USA (Condie, 1986). The presence of these features suggests that complete Wilson cycles (Section 7.9) of ocean opening and closure and ophiolite obduction were occurring by at least 2.0 Ga.

Together these observations strongly suggest that plate tectonic mechanisms became increasingly important after the Late Archean. Most authors link this development to the increased stability of continental crust during this Eon (Section 11.4.2). Nevertheless, it is important to realize that many uncertainties about Proterozoic tectonics remain. The age of formation of the rock units and the timing of regional metamorphism, deformation, and cooling are poorly known in many regions. The sources of magmatism and the amounts and mechanisms of crustal recycling also commonly are unclear. In addition, there continues to be a need for high resolution paleomagnetic and geochronologic data to enable accurate reconstructions of the continents and oceans during Proterozoic as well as Archean times (Section 11.5.3). These data are crucial for determining the tectonic evolution of Proterozoic lithosphere and for detailed comparisons between Proterozoic and Phanerozoic orogens.

11.5 THE SUPERCONTINENT CYCLE

11.5.1 Introduction

Geologic evidence for the repeated occurrence of continental collision and rifting since the Archean has led to the hypothesis that the continents periodically coalesced into large landmasses called supercontinents. The best known of the supercontinents include Gondwana (Fig. 3.4) and Pangea (Fig. 11.27), which formed in the latest Proterozoic and late Paleozoic times, respectively. Other supercontinents, such as Rodinia and Laurussia, also have been proposed for Late Proterozoic and late Paleozoic times, respectively. Processes in the mantle that may have led to their assembly and dispersal are discussed in Section 12.11.

11.5.2 Pre-Mesozoic reconstructions

Paleogeographic maps for the Mesozoic and Cenozoic can be computed by the fitting together of continental margins or oceanic lineations of the same age on either side of an ocean ridge (Chapters 3, 4). The location of the paleopoles can be determined from paleomagnetic measurements (Section 3.6) and so the only unknown in these reconstructions is the zero meridian of longitude. These combined techniques cannot be used for reconstructions prior to the Mesozoic because *in situ* oceanic crust is lacking.

Methods of quantifying plate motions in pre-Mesozoic times involve the use of paleomagnetic data coupled with high-precision geochronology. Ancient plate edges, although somewhat distorted, are marked by orogenic belts and ophiolite assemblages (Sections 2.5, 11.4.3), which indicate ancient sutures between welded continents and accreted terranes. Evidence provided from the past distributions of flora and fauna and indicators of paleoclimate also aid these plate reconstructions (Sections 3.4, 3.5). For a particular time the paleomagnetic pole for each ancient plate is rotated to an arbitrary single magnetic pole and the continents on the plate are re-projected using the same Eulerian rotation. The continents are then successively shifted along fixed latitudes, that is, rotated about the magnetic pole, until the overlap of continental margins is minimized. Although the paleomagnetic data do not provide a unique sequence of reconstructions, they clearly indicate the gross trends of plate movements during ancient times. More detailed inferences on the evolution of particular regions are then made from their geology viewed in terms of plate tectonic mechanisms.

The application of paleomagnetic methods for the Precambrian is less straightforward than for Phanerozoic times for three main reasons (Dunlop, 1981). First, the error limits of isotopic ages typically are larger. Second, isotopic and magnetic records may be partially reset during metamorphism to different degrees, and the distinction between pre- and post-orogenic isotopic and magnetic overprints can be difficult. Third, overprints occur during post-orogenic cooling and uplift, and the temperatures at which isotopic systems close and magnetizations stabilize are different, so that the dates may be younger or older than the magnetizations by intervals of tens of millions of years. However, even given these uncertainties and the gaps in the paleomagnetic record arising from the lack of suitable samples of certain ages, the data allow investigators to test the validity of paleogeographic reconstructions for pre-Mesozoic times based on the geologic record on the continents.

11.5.3 A Late Proterozoic supercontinent

Similarities between the Late Proterozoic geologic record in western Canada and eastern Australia (Bell & Jefferson, 1987; Young, 1992) and between the southwestern USA and East Antarctica suggest that these areas were juxtaposed during Late Proterozoic times (Dalziel, 1991, 1995; Moores, 1991; Hoffman, 1991) (Fig. 11.19a). This seemingly radical suggestion was referred to as the SWEAT (**S**outh **W**est US and **E**ast **AnT**arctica) hypothesis. The widespread Grenville orogenic belts, that immediately pre-date the Late Proterozoic, suggest that many other continental fragments can be added to this reconstruction to form a Late Proterozoic supercontinent called *Rodinia* (Fig. 11.19a). Laurentia (North America and Greenland) forms the core of the supercontinent and is flanked to the north by East Antarctica. The reconstruction shows that the North

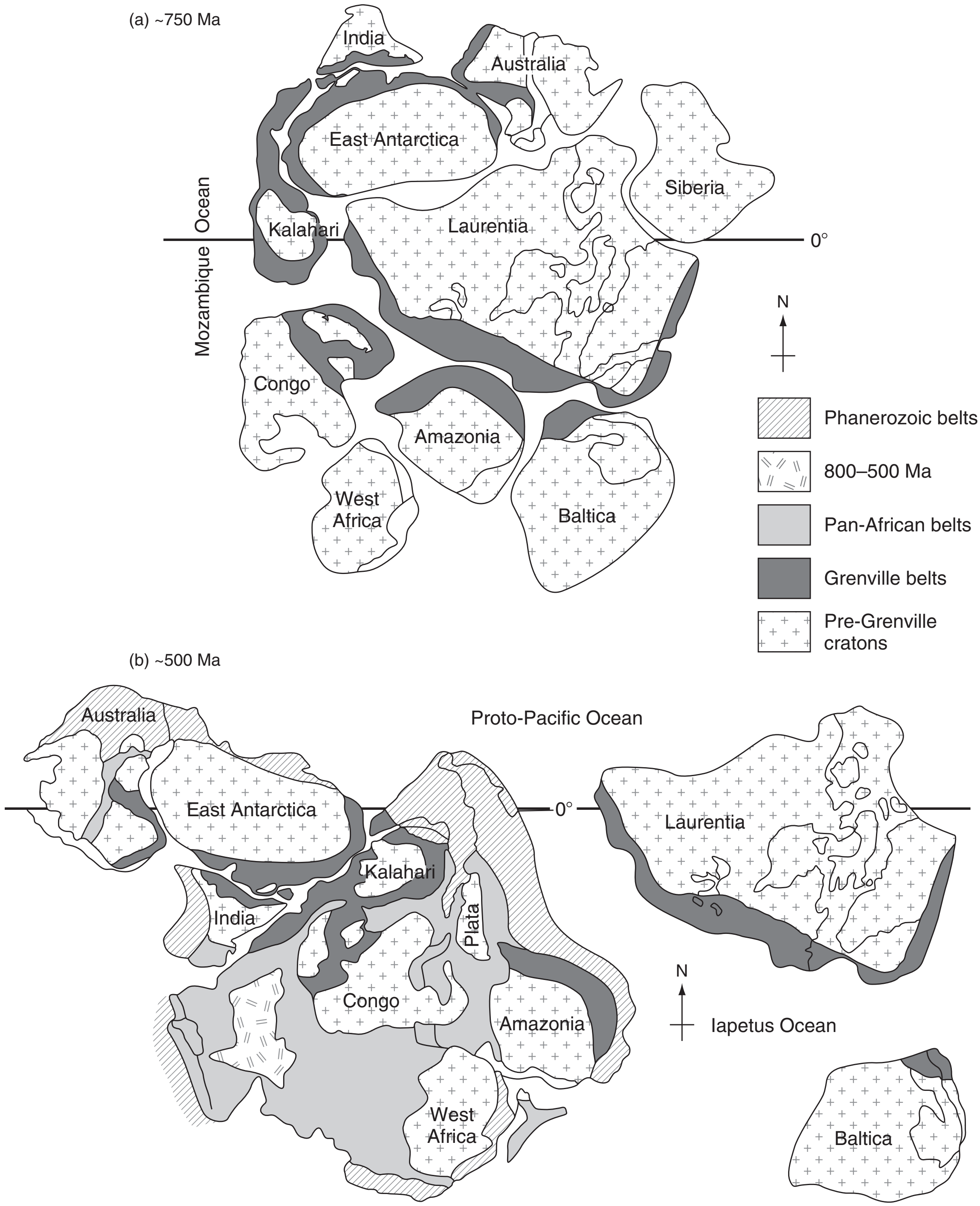


Fig. 11.19 *(a) Reconstruction of the Late Proterozoic supercontinent Rodinia. (b) Late Cambrian paleogeography after the break-up of Rodinia and the formation of Gondwana (after Hoffman, 1991, with permission from* Science ***252****, 1409–12 with permission from the AAAS).*

American Grenville Province continues directly into East Antarctica, and similar belts of this age can be traced over most of the Gondwana fragments. The age of the oldest sedimentary rock associated with break-up, and the provinciality of certain animal groups across the split, suggest that the supercontinent fragmented at about 750 Ma (Storey, 1993). During fragmentation the blocks now making up East Gondwana (East Antarctica, Australia, and India) moved anticlockwise, opening the proto-Pacific Ocean (Panthalassa), and collided with the blocks of West Gondwana (Congo, West Africa, and Amazonia). The intervening Mozambique Ocean closed by the pincer-like movements of these blocks and Gondwana was created when they collided to form the Mozambique belt of East Africa and Madagascar. Gondwana then rotated clockwise away from Laurentia about 200 Ma later. Southern Africa was located at the pivot of these movements and Baltica moved independently away from Laurentia, opening the Iapetus Ocean, which subsequently closed during the assembly of Pangea (Section 11.5.5). Figure 11.19b shows a postulated configuration at 500 Ma.

The first paleomagnetic test of the SWEAT hypothesis was carried out by Powell *et al.* (1993) who showed that paleomagnetic poles at 1055 Ma and at 725 Ma for Laurentia and East Gondwana are in agreement when repositioned according to the Rodinia reconstruction, thereby lending support to the hypothesis. Between 725 Ma and the Cambrian the APWPs diverge, suggesting that East Gondwana broke away from Laurentia after 725 Ma. The only fragment of Rodinia for which a detailed Apparent Polar Wanderer (APW) path can be defined for the period 1100–725 Ma is Laurentia (McElhinny & McFadden, 2000). This, therefore, has been used as a reference path against which repositioned paleomagnetic poles from other Rodinian fragments can be compared. However, many of the tests were hindered by a lack of high quality geochronology. As new data were collected, the existence of a Late Proterozoic supercontinent gained acceptance, although numerous modifications have been proposed (Dalziel *et al.*, 2000b; Karlstrom *et al.*, 2001; Meert & Torsvik, 2003). There is now considerable geologic and paleomagnetic evidence that, except for Amazonia, the cratons of South America and Africa were never part of Rodinia, although they probably were close to it (Kröner & Cordani, 2003). Newer models also indicate the piecemeal assembly of Rodinia beginning with Grenville-age collisions in eastern Canada and Australia at 1.3–1.2 Ga, followed by an Amazonia–Laurentia collision at 1.2 Ga (Tohver *et al.*, 2002), the majority of assembly between 1.1 and 1.0 Ga, and minor collisions between 1.0 and 0.9 Ga (Li *et al.*, 2008). Most current models of Rodinia also show a fit between the cratons at 750 Ma that differs substantially from the older hypotheses (Wingate *et al.*, 2002). Torsvik (2003) published a model (Fig. 11.20) that summarizes some of these changes. The position of the continents suggests that the break-up of Rodinia had begun by 850 or 800 Ma with the opening of the proto-Pacific ocean between western Laurentia and Australia-East Antarctica. The emplacement of mafic dike swarms in western Laurentia at 780 Ma may reflect this fragmentation (Harlan *et al.*, 2003). The position of Australia-East Antarctica also suggests that India was not connected to East Antarctica until after ~550 Ma. This model emphasizes that the internal geometry of Rodinia probably changed repeatedly during the few hundred million years it existed.

The differences among the new and old models of Rodinia illustrate the controversial and fluid nature of Precambrian reconstructions. Numerous uncertainties in the relative positions of the continents exist, with the paleolatitudes of only a few cratons being known for any given time. It also must be remembered that paleomagnetic methods give no control on paleolongitude (Section 3.6), so that linear intercratonic regions whose strike is directed towards the Eulerian pole used to bring the cratons into juxtaposition are not constrained to have had any particular width. For these reasons, most reconstructions rely on combinations of many different data sets, including geological correlations based on orogenic histories, sedimentary provenance, the ages of rifting and continental margin formation, and the record of mantle plume events (Li *et al.*, 2008).

Another controversial aspect of the Rodinia supercontinent concerns the effect of its dispersal on past climates. Some studies suggest that as Rodinia fragmented the planet entered an icehouse or *snowball Earth* state in which it was intermittently completely covered by ice (Evans, 2000; Hoffman & Schrag, 2002). The geologic evidence for this intermittent but widespread glaciation includes glacial deposits of Late Proterozoic age that either contain carbonate debris or are directly overlain by carbonate rocks indicative of warm waters. In addition, paleomagnetic data suggest that during at least two Late Proterozoic glacial episodes ice sheets reached the equator. One possible explanation of these observations is that periods of global glaciation during the Late Proterozoic were controlled by anomalously low atmospheric CO_2 concentrations (Hyde *et al.*, 2000;

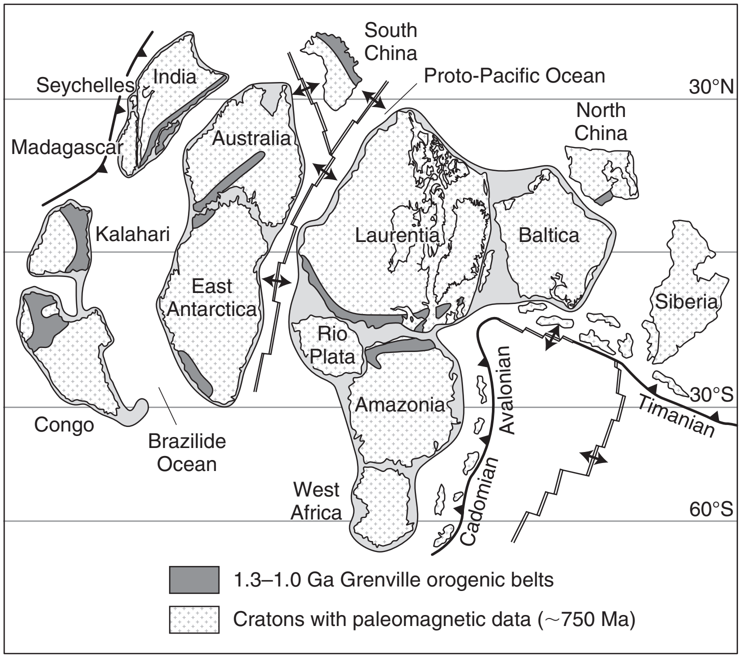


Fig. 11.20 *Reconstruction of Rodinia at ~750 Ma (after Torsvik, 2003, with permission from* Science ***300****, 1379–81, with permission from the AAAS).*

Donnadieu *et al.*, 2004). During break-up, the changing paleogeography of the continents may have led to an increase in runoff, and hence consumption of CO_2, through continental weathering that decreased atmospheric CO_2 concentrations (Section 13.1.3). The extreme glacial conditions may have ended when volcanic outgassing of CO_2 produced a sufficiently large greenhouse effect to melt the ice. The resulting "hothouse" would have enhanced precipitation and weathering, giving rise to the deposition of carbonates on top of the glacial deposits during sea-level (Hoffman *et al.*, 1998). Alternatively, these transitions may have resulted mainly from the changing configuration of continental fragments and its effect on oceanic circulation (Sections 13.1.2, 13.1.3). Whichever view is correct, these interpretations suggest that the break-up of Rodinia triggered large changes in global climate. However, the origin, extent, and termination of the Late Proterozoic glaciations, and their possible relationship to the supercontinental break-up, remains an unresolved and highly contentious issue (Kennedy *et al.*, 2001; Poulsen *et al.*, 2001).

11.5.4 Earlier supercontinents

The origin of the first supercontinent and when it may have formed are highly speculative. Bleeker (2003) observed that there are about 35 Archean cratons today (Plate 11.1a between pp. 244 and 245) and that most display rifted margins, indicating that they fragmented from larger landmasses. Several possible scenarios have been envisioned for the global distribution of the cratons during the transition from Late Archean to Early Proterozoic times (Fig. 11.21). These possibilities include a single supercontinent, called *Kenorland* by Williams *et al.* (1991) after an orogenic event in the Canadian Superior Province, and the presence of either a few or many independent aggregations called *supercratons*. Bleeker (2003) concluded that the degree of geologic similarity among the exposed cratons favors the presence of several transient, more or less independent supercratons rather than a single supercontinent or many small dispersed landmasses. He defined a

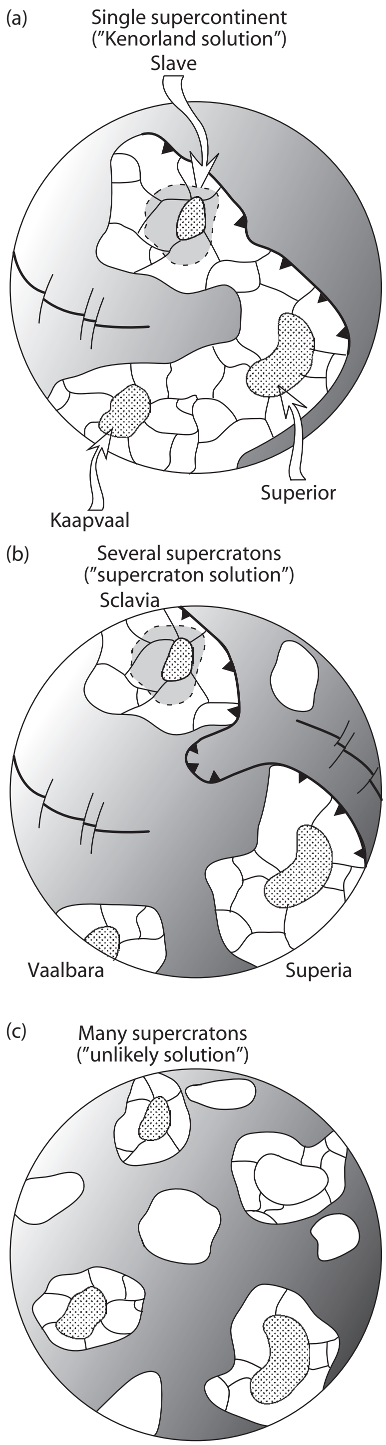


Fig. 11.21 *Cartoons representing possible craton configurations during Late Archean–Early Proterozoic times. Three well-known cratons (Slave, Superior and Kaapvaal) are shown shaded in (a). These cratons may have been spawned by the larger supercratons shown in (b) (after Bleeker, 2003, with permission from Elsevier).*

minimum of three supercratons, *Sclavia*, *Superia*, and *Vaalbara*, that display distinct amalgamation and break-up histories (Fig. 11.21b). The *Sclavia* supercraton appears to have stabilized by 2.6 Ga. Confirmation of these tentative groupings awaits the collection of detailed chronostratigraphic profiles for each of the 35 Archean cratons.

Diachronous break-up of the supercratons defined by Bleeker (2003) occurred during the period 2.5–2.0 Ga, spawning the 35 or more independently drifting cratons. Paleomagnetic evidence supports the conclusion that significant differences in the paleolatitudes existed between at least several of these fragments during the Early Proterozoic (Cawood *et al.*, 2006). Following the break-up the cratons then appear to have amalgamated into various supercontinents. Hoffman (1997) postulated a Middle Proterozoic supercontinent called *Nuna*, which Bleeker (2003) considered to represent the first true supercontinent. Zhao *et al.* (2002) also recognized that most continents contain evidence for 2.1–1.8 Ga orogenic events (Section 11.4.3) (Fig. 11.12). They postulated that these orogens record the collisional assembly of an Early–Middle Proterozoic supercontinent called *Columbia* (Fig. 11.22). These studies, while still speculative, suggest that at least one supercontinent formed prior to the final assembly of Rodinia and after the Archean cratons began to stabilize during the Late Archean.

11.5.5 Gondwana–Pangea assembly and dispersal

The assembly of Gondwana began immediately following the break-up of Rodinia in Late Proterozoic times. According to the SWEAT hypothesis (Section 11.5.3) West Gondwana formed when many small ocean basins that surrounded the African and South American cratons closed during the opening of the proto-Pacific Ocean, creating the Pan-African orogens (Fig. 11.19b). Subsequent closure of the Mozambique Ocean resulted in the collision and amalgamation of West Gondwana with the blocks of East Gondwana. This amalgamation may have created a short-lived Early Cambrian supercontinent called *Pannotia*. The existence of this supercontinent is dependent on the time of rifting between Laurentia and Gondwana (Cawood *et al.*, 2001). Models of Pannotia (Fig. 11.23a) are based mostly on geologic evidence that Laurentia and Gondwana were attached

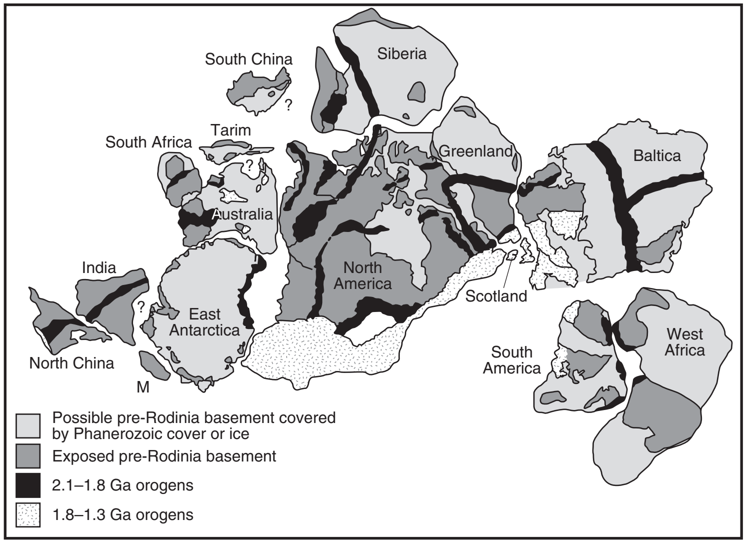


Fig. 11.22 *Reconstruction of the postulated Early–Middle Proterozoic supercontinent Columbia (after Zhao* et al.*, 2002, with permission from Elsevier). M, Madagascar. Early Proterozoic orogens (2.1–1.8 Ga) are identified in Fig. 11.12.*

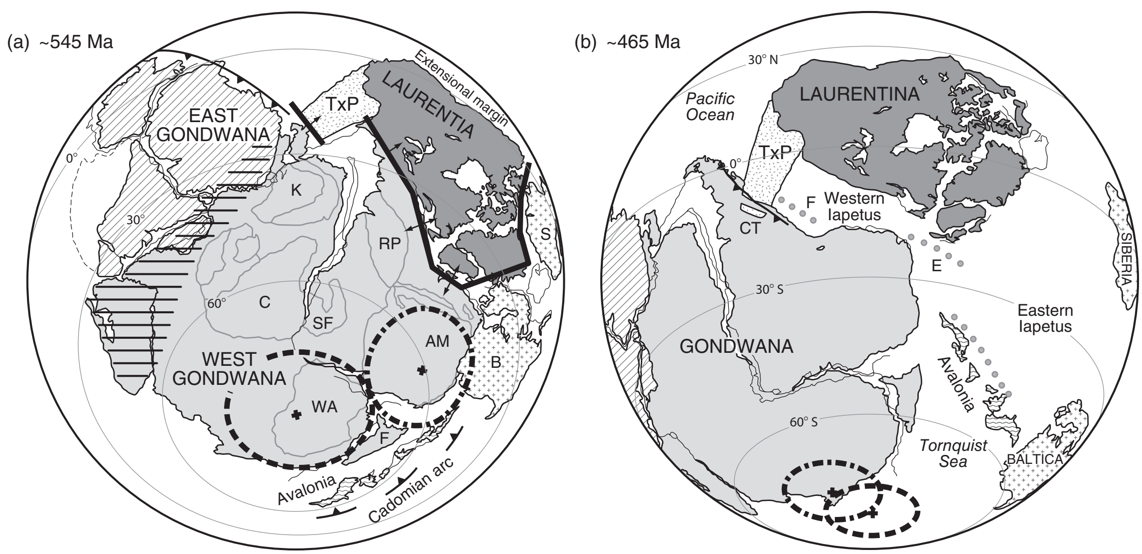


Fig. 11.23 *Postulated reconstructions of (a) Pannotia at ~545 Ma and (b) the rifting of Laurentia and Gondwana at ~465 Ma emphasizing the paleogeography of Laurentia relative to Gondwana (after Dalziel, 1997, with permission from the Geological Society of America). Crosses with 95% confidence circles shown in (a) with a dashed line and a dashed-dotted line indicate paleomagnetic poles for Laurentia and Gondwana, respectively. Horizontal lines in (a) denote limit of the Mozambique orogenic belt; thick solid black line marks the location of a Laurentia–Gondwana rift. In (b) paleomagnetic poles are for Laurentia + Baltica + Siberia + Avalonia (cross with dashed confidence circle) and Gondwana (cross with dashed-dotted confidence circle). Cuyania (CT) has accreted onto the Gondwana margin. Abbreviations: C, Congo; K, Kalahari; WA, West Africa; AM, Amazonia; RP, Río de la Plata; SF, São Francisco; S, Siberia; B, Baltica; TxP, the hypothetical Texas Plateau; F, Famatina arc; E, Exploits arc.*

or in close proximity at the end of the Late Proterozoic (Dalziel, 1997). However, the paleomagnetic poles for these two landmasses do not overlap, suggesting that an alternative configuration where Laurentia is separated from Gondwana at this time also is possible (Meert & Torsvik, 2003).

Most models suggest that the break-up of Pannotia began with the latest Proterozoic or Early Cambrian opening of the Iapetus Ocean as Laurentia rifted away from South America and Baltica (Figs 11.19b, 11.23b). Subduction zones subsequently formed along the Gondwana and Laurentia margins of Iapetus, creating a series of volcanic arcs, extensional backarc basins, and rifted continental fragments. As the ocean closed this complex assemblage of terranes accreted onto the margins of both Laurentia and Gondwana. The provenance of these terranes provides a degree of control on the relative longitudes and paleogeography of these two continents prior to the Permo-Carboniferous assembly of Pangea (Dalziel, 1997).

The Early Paleozoic accreted terranes of Laurentia and Gondwana are classified into groups according to whether they are native or exotic to their adjacent cratons (Keppie & Ramos, 1999; Cawood, 2005). Those native to Laurentia include the Notre Dame–Shelburne Falls (Taconic) and Lough–Nafooey volcanic arcs (Figs 10.34, 11.24a), which formed near and accreted onto Laurentia during Early–Middle Ordovician times. These collisions were part of the Taconic Orogeny in the Appalachians (Karabinos *et al.*, 1998), the Grampian Orogeny in the British Isles, and the Finnmarkian Orogeny in Scandinavia. During the same period, the Famatina arc terrane (Fig. 11.23b), of Gondwana affinity, formed near and accreted onto the western margin of South America (Conti *et al.*, 1996).

Terranes exotic to Laurentia include Avalonia, Meguma, Carolina, and Cadomia (Fig. 11.24a). These

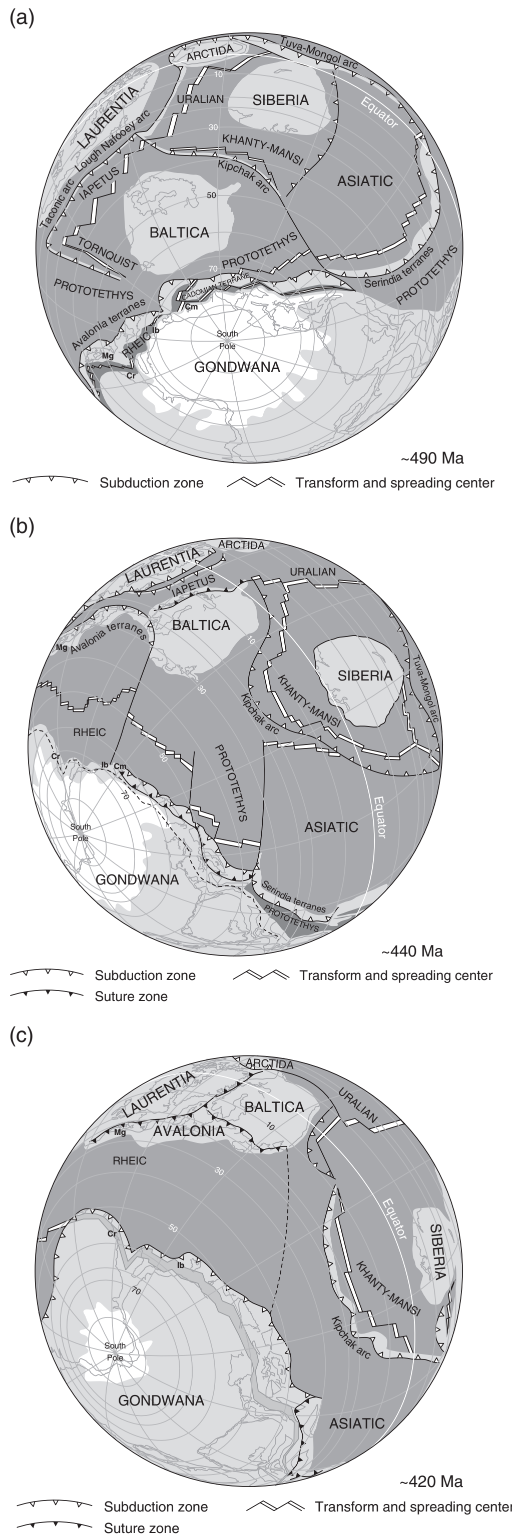


Fig. 11.24 *Postulated Paleozoic plate reconstructions for (a) 490 Ma, (b) 440 Ma, and (c) 420 Ma emphasizing the paleogeography of terranes derived from northern Gondwana and the opening of the Rheic Ocean (images provided by G.M. Stampfli and modified from von Raumer et al., 2003, and Stampfli & Borel, 2002, with permission from Elsevier). Interpretations incorporate the dynamics of hypothesized convergent, divergent and transform plate boundaries. Labeled terranes in (a) are: Mg, Meguma; Cm, Cadomia; Ib, Iberia; Cr, Carolina.*

and other terranes rifted from northwestern Gondwana in the Early Ordovician and later accreted onto the Laurentian margin, forming part of the Silurian-Devonian Acadian and Salinic orogens in the northern Appalachians and the Caledonides of Baltica and Greenland (Figs 11.24c, 11.25). Cuyania, an exotic terrane located in present day Argentina (Fig. 11.23b), rifted from southern Laurentia during Early Cambrian times and later accreted onto the Gondwana margin (Dalziel, 1997). These tectonic exchanges suggest that at least two different plate regimes existed in eastern and western Iapetus during the Paleozoic with subduction zones forming along parts of both Gondwana and Laurentia (Fig. 11.24a). Although the geometry of the plate boundaries is highly speculative, the interpretation of distinctive plate regimes explains the piecemeal growth of both continents by terrane accretion prior to the assembly of Pangea.

The rifting of the Avalonia terranes from Gondwana in the Late Cambrian and Early Ordovician led to the opening of the Rheic Ocean between the Gondwana mainland and the offshore crustal fragments (Fig. 11.24a,b). After the closure of Iapetus and the accretion of Avalonia, the Rheic Ocean continued to exist between Laurentia and Gondwana, although its width is uncertain (Fig. 11.24c). During these times a new series of arc terranes rifted from the Gondwana margin, resulting in the opening of the Paleotethys Ocean (Fig. 11.26a). The opening of Paleotethys and the closure of the Rheic Ocean eventually resulted in the accretion of these Gondwana-derived terranes onto Laurentia followed by a continent–continent collision between Laurussia and Gondwana (Fig. 11.26b). This latter collision produced the Permo-Carboniferous Alleghenian and Variscan orog-

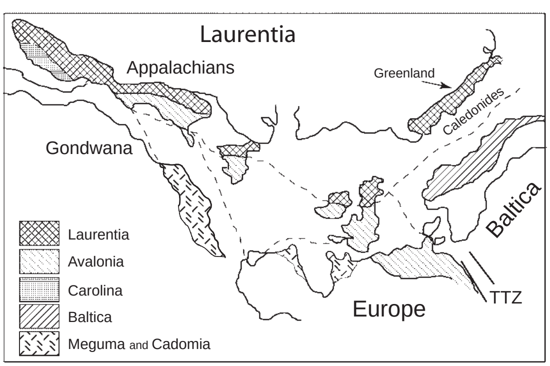


Fig. 11.25 *Late Paleozoic reconstruction showing the Silurian–Devonian Appalachian (Acadian and Salinic)–Caledonian orogens (after Keller & Hatcher, 1999, with permission from Elsevier). TTZ is the Teisseyre-Tornquist zone, representing a major crustal boundary between Baltica and southern Europe.*

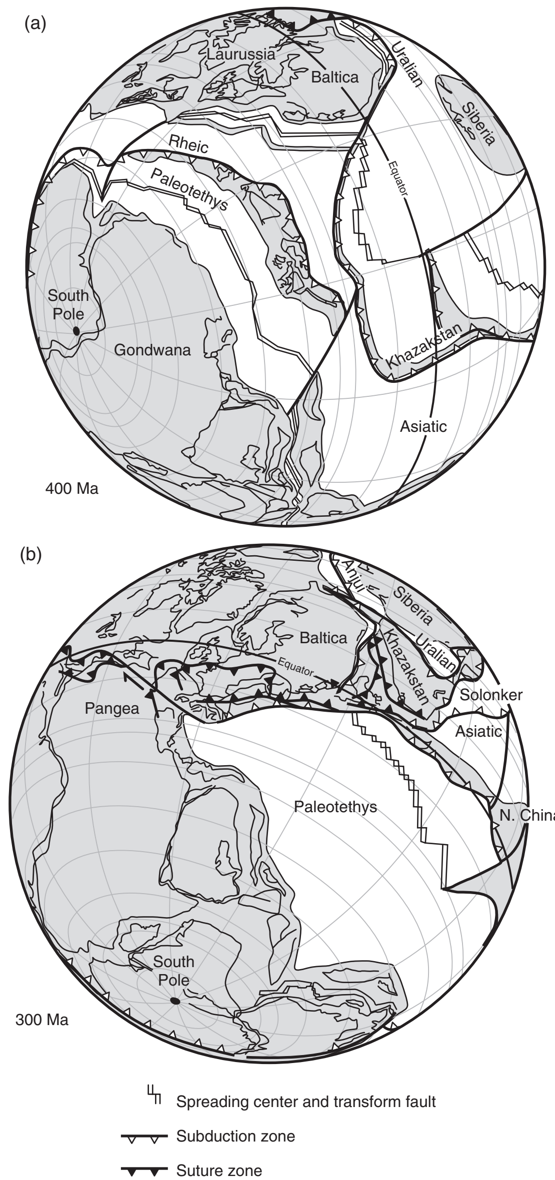


Fig. 11.26 *Postulated Paleozoic plate reconstructions for (a) 400 Ma and (b) 300 Ma (images provided by G.M. Stampfli and modified from Stampfli & Borel, 2002, with permission from Elsevier). In (a) The Rheic Ocean closes as Paleotethys opens. In (b) Gondwana has collided with Laurussia creating the European Variscides and Alleghenian Orogen.*

Fig. 11.27 *Reconstruction of Pangea at 250 Ma (after Torsvik, 2003, with permission from Science* **300**, *1379–81, with permission from the AAAS). Major cratons are shown.*

enies in North America, Africa, and southwest Europe. Collisions in Asia, including the suturing of Baltica and Siberia to form the Ural Orogen at ~280 Ma, resulted in the final assembly of Pangea. The supercontinent at the height of its extent at ~250 Ma is shown in Fig. 11.27.

Like its assembly, the fragmentation of Pangea was heterogeneous. Break-up began in the mid-Jurassic with the rifting of Lhasa and West Burma from Gondwana and the opening of the central Atlantic shortly after 180 Ma (Lawver *et al*., 2003). Magnetic anomalies indicate that by 135 Ma the southern Atlantic had started to open. Rifting between North America and Europe began during the interval 140–120 Ma. Africa and Antarctica began to separate by 150 Ma. Australia also began to rift from Antarctic by 95 Ma with India separating from Antarctica at about the same time. These data indicate that the majority of Pangea break-up occurred during the interval 150–95 Ma. Small fragments of continental crust such as Baja California and Arabia continue to be rifted from the continental remnants of Pangea. As with the older supercontinents, the break-up of Pangea was accompanied by the closure of oceans, such as Paleotethys and Neotethys (Fig. 11.27), and by collisions, including those that occur presently in southern Asia (Fig. 10.13), southern Europe, and Indonesia (Fig. 10.28).