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A lithospheric perspective on structure and evolution of Precambrian cratons

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5.1 Introduction

The purpose of this chapter is to provide a summary of geophysical data on the structure of the stable continental lithosphere and its evolution since the Archean. Here, the term lithosphere is used to define the outer layer of the Earth which includes the crust and uppermost mantle, forms the roots of the continents, and moves together with continental plates. Depending on geophysical techniques (and physical properties measured), the lithosphere has different practical definitions. Most of them (i.e., seismic, electrical) are on the basis of a sharp change in temperature-dependent physical properties at the transition from conductive to convecting mantle.

Here, conventionally, the term craton is used to refer to a stable continental terrane which has not undergone any significant tectonic events since Precambrian. However, as the Proterozoic–Paleozoic boundary is defined by global paleontological, but not by geological changes, some Paleozoic structures which remained tectonically stable for several hundred million years can also be considered as cratons (e.g., the West Siberian Basin).

Two approaches are used when discussing the ages of continental terranes, that is, geologic and tectono-thermal ages. Isotopic studies provide the geologic ages of the rocks forming the crustal basement. Cratons with Precambrian basement occupy approximately 70% of the continents; their tectonic boundaries are well established by geological studies (Fig. 5.1A). Furthermore, the age of the oldest continental subcrustal lithosphere is >3.0 Ga

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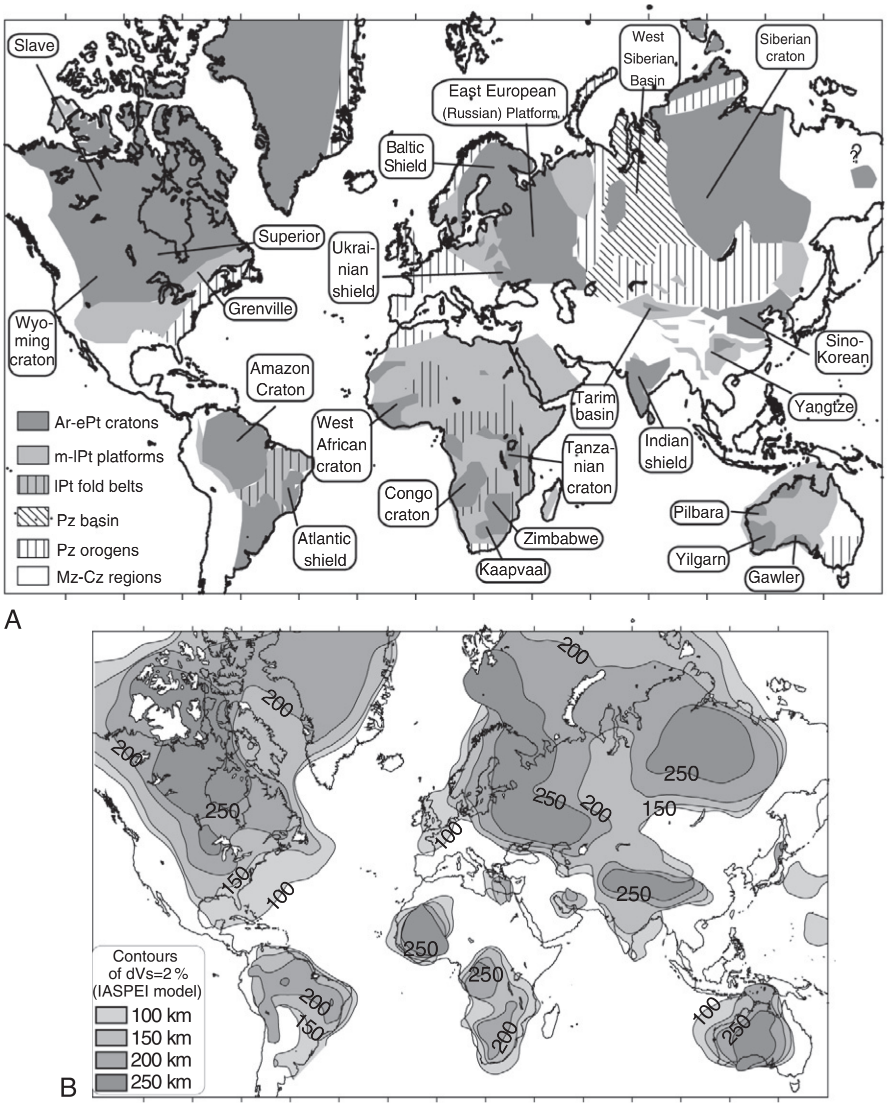

Figure 5.1 (A) Tectonic provinces of the world based on geological ages and tectonics. For some structures the tectono-thermal age (i.e., the age of the last major tectonic event) is much younger than the geological (i.e., isotopic) age of the crust. Most basins with Paleozoic tectono-thermal ages were formed on Precambrian terranes and thus are parts of the cratons. The Antarctic craton (for which the amount of data is very limited) and the Pacific region (which does not include any cratonic structures) are not shown in the map and are excluded from discussion. (B) Thickness of continental lithosphere as defined from Rayleigh wave seismic tomography (Shapiro and Ritzwoller, 2002) as the upper mantle layer with a positive 2% velocity anomaly compared to the global continental velocity model IASPEI (Kennett and Engdahl, 1991). As surface waves lose resolution at depths deeper than ca. 250 km, this is the deepest isoline shown. Some cratonic regions do not show distinct positive velocity anomalies in the upper mantle either because they are not resolved (because of their small size or poor ray coverage, e.g., Indian and Tanzanian cratons) or because they have lost deep lithospheric roots during later tectonic events (the Sino-Korean craton). Some non-cratonic regions, associated with modern subduction zones (e.g., Hellenic arc, Tibet, Andes), show strong linear zones of positive velocity anomalies in the upper mantle, caused by cold temperatures in subducting slabs.

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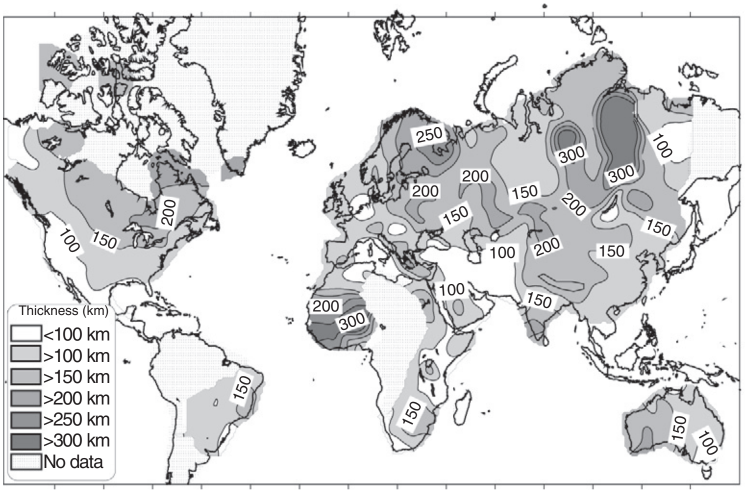

Figure 5.1 Cont'd (C) Thickness of continental lithosphere as determined from surface heat flow data and defined as the depth to the 1300 °C adiabat *(modified after Artemieva and Mooney, 2001).* Lateral smearing of the model is ca. 800 km. Most of the cratons are well resolved as regions with low temperatures and, hence, large lithospheric thickness. No constraints of lithospheric geotherms are possible for parts of South America, Africa, Canada, and Greenland (hatched), as surface heat flow measurements there are absent or very scarce.

down to a depth of >200 km as indicated by isotopic studies of mantle peridotites (e.g., Richardson et al., 1993). Data on the geologic ages of terranes are important for understanding the properties of cratonic lithosphere (e.g., its density, seismic velocities, composition), as similar isotopic ages for its crustal and mantle parts suggest that the entire lithospheric column was formed at the same time and since then evolved as a whole (Pearson, 1999).

However, it is important to note that cratonic margins are not vertical boundaries, implying that crust of one age can be underlain by mantle of a different age. For example, LITHOPROBE studies indicate that the Archean crust of the Slave craton is underlain by early Proterozoic mantle (Bostock, 1999), while the middle Proterozoic crust of the Grenville province is underlain by the Archean mantle of the Superior province (Ludden and Hynes, 2000). Similarly, the lower crust and the uppermost mantle of the Proterozoic Baltic shield extend southwards beneath the Caledonian crust of northern Europe over a distance of ca. 200–300 km (Thybo, 1990), while the Archean-early Proterozoic mantle of the East European craton underlies the Paleozoic Uralides orogen (Poupinet et al., 1997). Furthermore, later tectonic activity resulted in a formation of several large basins on the Archean-early Proterozoic basement, with tectonothermal ages (i.e., the ages of the last major tectonic event) ranging from late Proterozoic (e.g., Moscow basin) through Paleozoic (e.g., Michigan basin) to Meso-Cenozoic (e.g., Chad basin).

5.2 Lateral and depth extent of the cratons

Thick cold lithosphere of stable continental regions is well mapped by geophysical studies. Regions with exposed Precambrian crust are characterised by low surface heat flow (typically 40–50 mW/m^2; Nyblade and Pollack, 1993) and high seismic velocities in the lithospheric mantle (2–8% higher than in global models; Rohm et al., 2000). On the basis of the correlation between surface geology, heat flow and physical properties of the mantle, geophysical methods are used to outline both the cratonic boundaries at mantle depths and the depth extent of the cratonic lithosphere.

Further, the usual conventions of defining the base of the lithosphere are adopted: (1) for seismic lithosphere, it is the level in the Earth where seismic velocities exceed the values in a global reference model by 1–2% (Fig. 5.1B); (2) for thermal lithosphere, it is the depth where a geotherm intersects with a mantle adiabat (Fig. 5.1C; however, in the real Earth, a transition from conductive to convective heat transfer in the upper mantle occurs in a layer up to 40–50 km thick). For stable continents, mantle temperatures are usually constrained by surface heat flow measurements. As seismic velocities are sensitive to temperatures, both definitions give correlated estimates of the lithospheric thickness; however, in some regions, the difference between them can reach several tens of kilometres. The discrepancies between seismic tomography and thermal models result from modelling uncertainties (diverse data coverage, and different lateral and vertical resolution of the models) and from the sensitivity of seismic velocities to both temperature and compositional anomalies.

Figure 5.2 provides a summary of lithospheric thickness on the continents as determined from regional and global seismic tomography models, thermal model and xenolith data. Some global seismic tomography models give the largest (350–400 km) values for the Precambrian cratons. Regional seismic tomography models, which typically have higher resolution, show a positive seismic velocity anomaly extending down to 200–350 km depth. Thermal models also show a strong variability in lithospheric thickness in Precambrian cratons (from 140 to 350 km; Artemieva and Mooney, 2001). Petrologic studies of mantle-derived xenoliths provide independent constraints on mantle geotherms. However, xenolith P–T arrays are likely to indicate mantle temperatures at the time of magmatism and thus often give estimates of lithospheric thickness significantly different from geophysical methods, suggesting a nearly uniform thickness of the cratonic lithosphere (200–250 km). In some cratons reworked in the Meso-Cenozoic (e.g., Sino-Korean, Tanzanian), much smaller values of lithospheric thickness (ca. 140 km) are constrained by xenolith data, in agreement with geophysical estimates (Fig. 5.2). A typical cross-section of stable continental lithosphere is presented in Fig. 5.3.

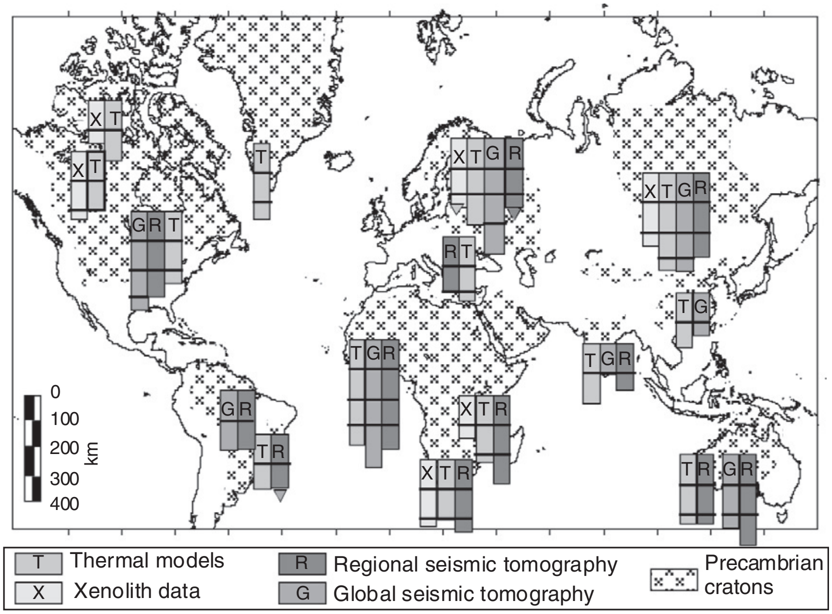

Figure 5.2 Thickness of the cratonic lithosphere (vertical bars) as determined by four different methods *(updated after Artemieva and Mooney, 2002, for new tomography data).* Thermal thickness is defined as the depth to the 1300 °C adiabat. Xenolith constraints are on the basis of P–T data. Lithospheric thickness from global seismic tomography is defined by +0.5% velocity anomaly; from regional seismic tomography by +1.0% velocity anomaly. Gray shading shows the outlines of the Precambrian cratons on the basis of geological data *(after Goodwin, 1996).* Triangles refer to models in which the base of the lithosphere has not been identified down to the indicated depth.

5.3 Correlation between lithospheric thickness and geological age

The Archean lithosphere is unique in that it has two typical thicknesses, 200–220 km and >300 km (Artemieva and Mooney, 2001; Fig. 5.4). Numerical simulations suggest that during its interaction with mantle convection, the depleted cratonic lithosphere tends to two equilibrium thicknesses, 220 and 350 km (Doin et al., 1997), and thus support the possibility of the existence of Archean cratons with two typical lithospheric thicknesses. Thick lithosphere (>250 km) is relatively rare and has been found by different geophysical methods only in the Northern Hemisphere (Fig. 5.2): the Siberian craton, West Africa, the Baltic shield, and the Canadian shield. Cratons of the Southern Hemisphere (South Africa, South America, India and Australia) have an almost uniform thickness of ca. 200 km. Such a geographical distribution of lithospheric thickness may reflect either the geometry of paleocontinents or selective reworking of Archean lithosphere (located at present in the Southern Hemisphere) by mantle superplumes. The first hypothesis is supported by an example of a paleoreconstruction of lithospheric thickness in an Archean supercontinent (Fig. 5.5), which suggests that all cratons with the present-day thick lithosphere could form the nuclei of a supercontinent, while cratons with thin lithosphere were dispersed to the peripheral parts. The second hypothesis is supported by findings of abundant diamonds of superdeep origin (derived from depth of at least 670 km;

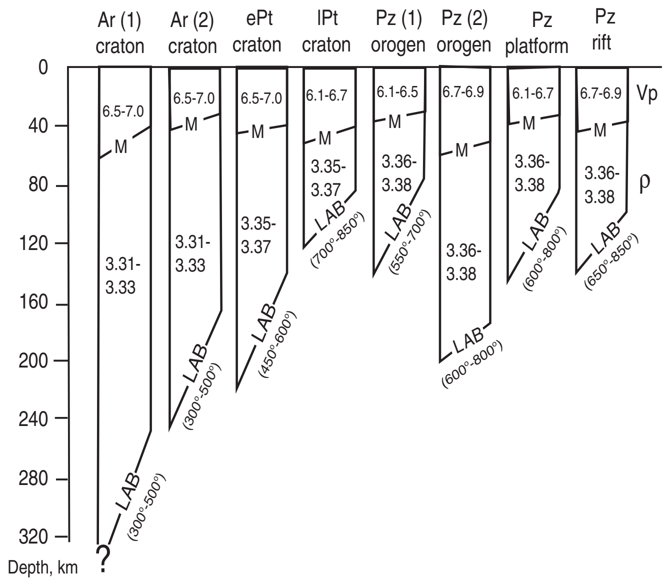

Figure 5.3 Typical cross-sections of stable continental lithosphere, showing ranges of crustal and lithospheric thickness, Moho temperatures (at the base of each column), average crustal velocities, and bulk densities of subcrustal lithosphere at standard P–T conditions *(from Griffin et al., 1998a,b).* Crustal properties derived from unpublished global crustal data base *(courtesy of W. Mooney, Chulick and Mooney, 2002).* Lithospheric thick-ness is on the basis of thermal and seismic tomography models. Ar (1) – Archean cratons with thick lithosphere (Siberian, Baltic Shield, West Africa, Canadian Shield); Ar (2) – Archean cratons with a relatively thin lithosphere (South Africa, South America, India, Australia, probably Antarctica and Congo). Paleozoic orogens are subdivided into two groups: Pz (1) includes the European Caledo-nides and Variscides; Pz (2) includes the Uralides and, probably, the northern Appalachians.

Davies et al., 1999) in kimberlites from the Slave, Gawler and Kaapvaal cratons (Gaul et al., 2000). There is, however, no evidence of any interaction between the present-day thickest cratonic lithosphere (e.g., the Siberian craton or the Superior province) with a lower mantle plume, as such superdeep diamonds were not found there (Gaul et al., 2000).

The question of how the Archean lithosphere has been formed is one of the outstanding questions in Earth sciences. A chemically distinct composition of the Archean lithosphere implies that it was formed under unique conditions, which did not exist in the post-Archean time (e.g., Boyd, 1989); the depleted composition has assisted the survival of cratonic roots through geological time. The proposed mechanisms for the formation of an early continental lithosphere include basal plume accretion of continental nuclei and crustal extraction from a primitive lower mantle source (Fig. 5.6). Episodic age distributions of juvenile crust, granitoids, greenstone belts (e.g., Condie, 1998; McLennan and Taylor, 1985), large igneous provinces, and giant dyke swarms (Yale and Carpenter, 1998) imply that, indeed, deep mantle processes played an important role in the formation and evolution of the continental lithosphere. Studies of the last decades (especially seismic reflection studies in Canada and Fennoscandia) suggest that plate tectonics operated already in the early Archean (Aulbach et al., 2001; Calvert et al., 1995; de Wit, 1998). These results provide support for the mechanisms of early lithosphere growth by stacking of oceanic terranes at pre-existing continental margins, underplating of buoyantly subducted slabs, or melting within the wedge of buoyantly subducting slabs (e.g., Abbott and Mooney, 1995; Fig. 5.6).

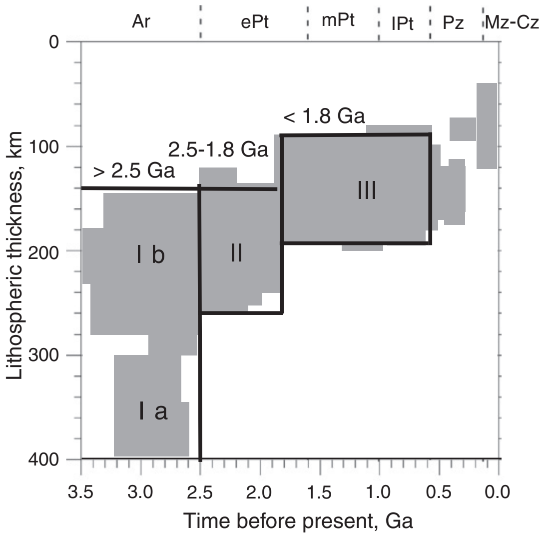

Figure 5.4 Secular changes in lithospheric thermal thickness (gray area) versus geologic ages *(from Artemieva and Mooney, 2001)*. Three stages of formation of the continental lithosphere, numbered I through III, are distinguished. The Archean is subdivided into two parts, corresponding to the regions of thick lithosphere (>300 km; Ia) and thinner lithosphere (~200–250 km; Ib). Key: Ar, Archean; ePt, mPt, lPt, early, middle and late Proterozoic, respectively; Pz, Paleozoic; Mz-Cz, Meso-Cenozoic.

Despite a large scatter in lithospheric thickness values (~100 km) for terranes of all ages, different geophysical methods reveal a clear global trend in a progressive thinning of the continental lithosphere with age (Fig. 5.4). By sharp changes in lithospheric thickness, three stages in the evolution of the continental lithosphere are proposed: >2.5, 2.5–1.8 and <1.8 Ga. These three stages are well correlated with other global phenomena: ages of greenstone belts and juvenile continental crust (with global peaks at 2.6–2.7, 1.7–1.9 and 1.0–1.3 Ga), global extractions of komatites and TTG (tonalite–trondhjemite–granodiorite) magmas, and a sharp change in the composition of cratonic peridotites formed before and after 2.5 Ga (Griffin et al., 1998a,b). All of these global processes are thought to reflect global changes in the pattern of mantle convection which dramatically changed at ca. 1.8–1.9 Ga because of secular cooling of the Earth (Condie, 1997). This sharp change in deep mantle dynamics is reflected in secular variations of lithospheric thickness (Fig. 5.4) and composition (Griffin et al., 1998a,b). Recent high-resolution whole-mantle seismic tomography models have revealed that, although most of subducting slabs stop near the 660 km discontinuity (Fukao et al., 2001), some may penetrate deep into the lower mantle (Van der Hilst et al., 1997). This result provides support for the hypothesis of episodic catastrophic overturns in the mantle (e.g., Condie, 1998): sinking of the subducting slabs into the lower mantle could initiate large mantle plumes (Davies, 1999) and lead to the episodic growth of Precambrian supercontinents (with the ages of 2.5–2.7, 1.7–2.1 and 1.0–1.3 Ga).

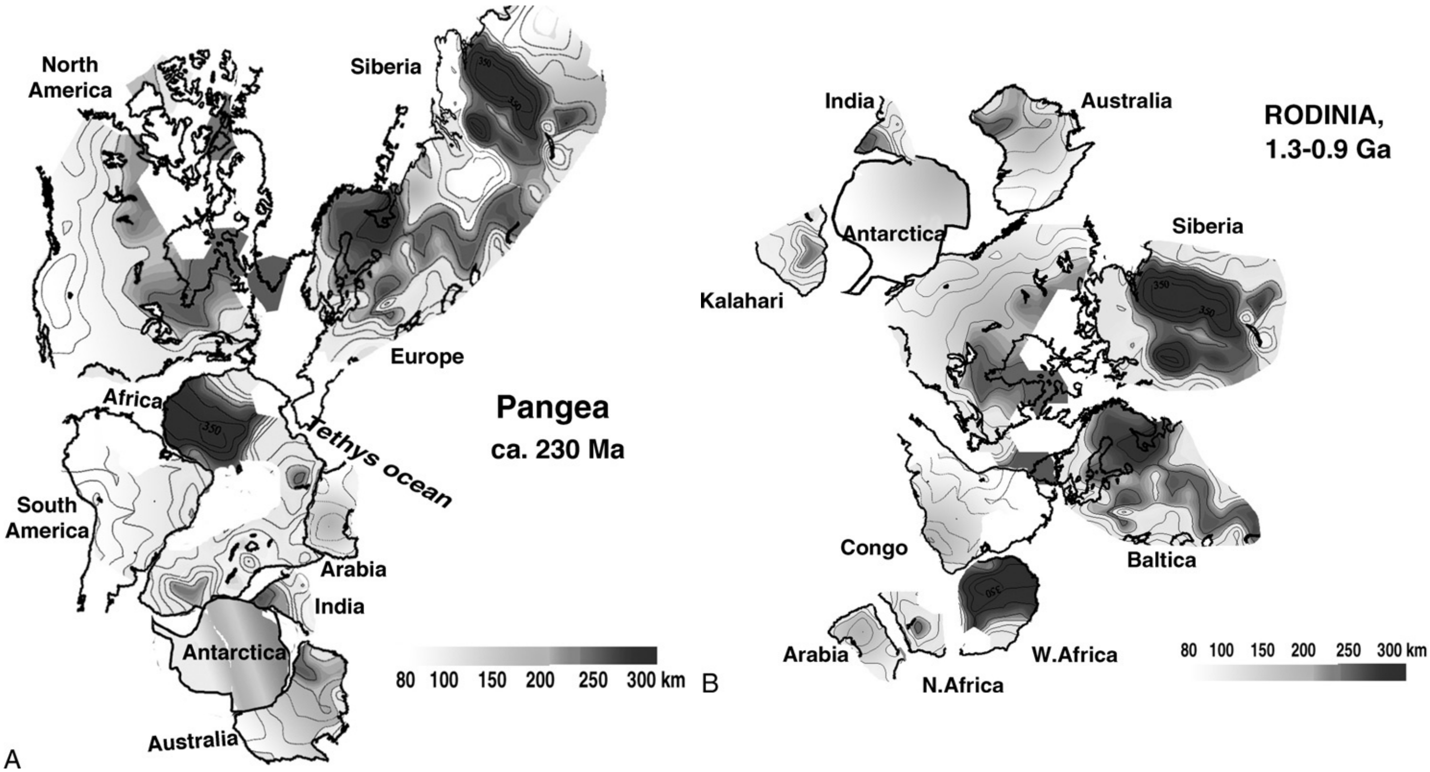

Figure 5.5 An example of a paleoreconstruction of lithospheric thickness in an Archean supercontinent (A) and in Pangea (B). Data on lithospheric thermal thickness (Artemieva and Mooney, 2001) are plotted on top of a paleoreconstruction of Rodinia (Dalziel et al., 2000) and Pangea (Torsvik and Cocks, 2004). The first supercontinent is believed to have existed as early as ca. 2.7–2.6 Ga, though no paleoreconstructions exist. Rodinia is the oldest supercontinent for which a paleoreconstruction of lithospheric thickness is possible. Thick cratonic keels could have impeded plate motion and thus it could be that Rodinia has inherited the core part of an earlier supercontinent.

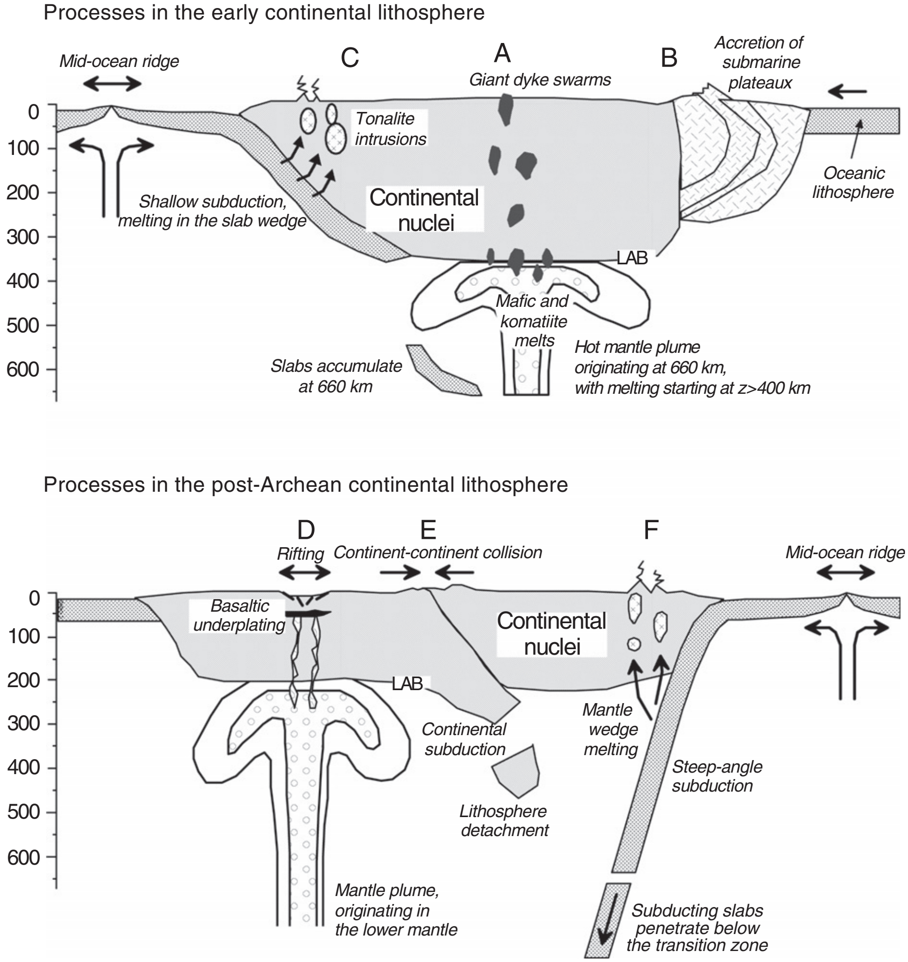

Figure 5.6 Major processes of lithosphere formation and modification in Archean and post-Archean time *(after Artemieva et al., 2002)*. Vertical and horizontal dimensions are not to scale. LAB, lithosphere-asthenosphere boundary. Processes in the early continental lithosphere include (but are not limited to): (A) assembly of continental nuclei above hot mantle plumes, originating from a depth of 660 km and generation of first giant dyke swarms; (B) growth of an Archean craton by the accretion of colliding submarine plateaux; (C) growth of Archean crust by melting within the slab wedge during buoyant subduction. Processes in the post-Archean continental lithosphere (shown are only the processes which differ from the processes in the Archean lithosphere); (D) rifting and basalt underplating caused by lower mantle plumes; (E) collisional orogens, continental subduction and lithosphere detachment; (F) crustal growth by melting in the mantle wedge during steep subduction. Some subducted slabs sink into the lower mantle.

5.4 Lithosphere modification by mantle convection and plumes

The Archean ages (3.0–3.25 Ga) of the oldest known continental rifting events (the Kaapvaal craton and the Keweenawan rift in the north-central United States) and the oldest giant mafic dyke swarms (southwestern Greenland) indicate that interaction between the asthenospheric mantle and the cratonic lithosphere has played an important role in the evolution of the continents since their very formation. Large-scale and vigorous secondary mantle convection and mantle plumes can modify the structure of cratonic lithosphere by thermo-mechanical interaction with its lower parts. An increase in mantle heat flow can lead to lithosphere thermal erosion, while gravitational (density) instabilities in the lithosphere (e.g., caused by magmatism or phase changes) can result in its mechanical delamination. Heating of the lower lithosphere by stirring or friction due to the horizontal movement of a craton over the underlying mantle also results in thermo-mechanical erosion of cratonic lithosphere. Furthermore, infiltration of basaltic magmas into the depleted lithosphere of cratonic roots can lead to compositional modification of cratonic lithosphere.

Figure 5.7 illustrates an evolution of an Archean supercontinent. Numerical simulations of mantle convection and tectonic reconstructions for Gondwanaland (Dalziel et al., 2000) indicate that mantle plumes may play an active role in supercontinental fragmentation. Alternatively, if continents act as thermal insulators above the convecting mantle, upwelling hot mantle can develop deviatoric stresses at the lithospheric base, large enough for continental break-up (Gurnis, 1988). If a supercontinent is split into parts of unequal sizes, further interaction of cratonic lithosphere with mantle convection would be different for the large and small continents. As thick cratonic lithosphere diverts heat coming from the deep mantle away from the craton into the thinner surrounding lithosphere (Ballard and Pollack, 1987), a large craton would be relatively more efficient in diverting the mantle heat and thus would be more resistant to basal thermal erosion. Nevertheless, some thermo-mechanical interaction with convecting mantle at its base will reduce its lithospheric thickness to an equilibrium thickness of $\sim$350 km. On the other hand, vigorous small-scale convection developed at the margins of the large craton would primarily erode it from the sides reducing its lateral size (Doin et al., 1997) and can lead to the extrusion of large igneous provinces, such as is observed at the edges of many cratons (King and Anderson, 1995). If, because of lateral erosion, the size of the large craton becomes less than a critical value, the erosion pattern can change to basal erosion, as is expected for a smaller craton. In this case, the small lateral extent of the cratonic keel will be insufficient to divert heat coming to its base from the mantle. This will result in strong basal thermo-mechanical erosion of the keel and thinning to an equilibrium thickness of $\sim$220 km (Doin et al., 1997).

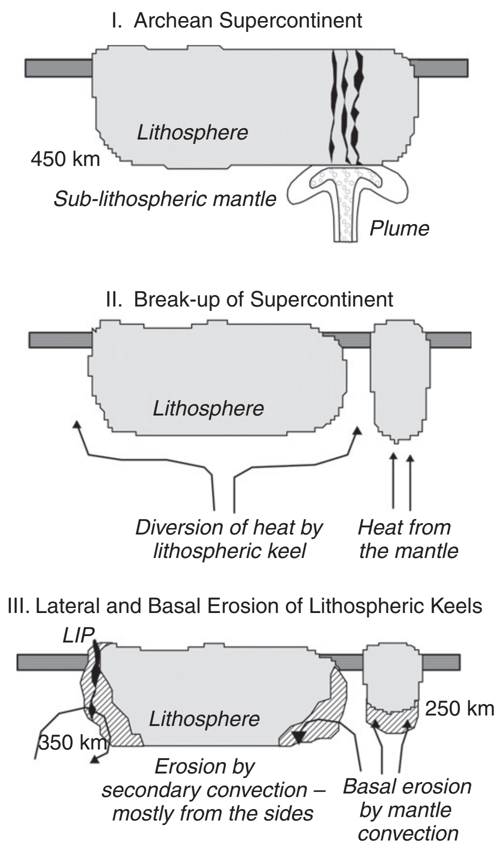

Figure 5.7 Selective erosion of cratonic lithosphere after an early Archean supercontinent with a thick keel is fragmented into two parts with non-equal dimensions *(after Artemieva and Mooney, 2002)*. An interaction of supercontinent fragments with mantle convection is determined by their lateral dimensions. Large craton diverts heat from the mantle and is eroded laterally to the equilibrium thickness of ca. 350 km. Small craton is inefficient in diverting the heat from its base and is eroded from the bottom until it reaches the equilibrium thickness of ca. 220 km (Doin et al., 1997). LIP, Large igneous province.

Strong basal heating of the lithosphere by vigorous mantle convection or by plumes may have other important consequences for the evolution of cratonic lithosphere. Infiltration of basaltic (Fe-enriched) magmas into a depleted (and relatively low-dense) cratonic keel can cause metasomatism, density increase and, as a result, platform subsidence (Fig. 5.8). Such a mechanism can explain the ongoing subsidence of the southern parts of the East European platform that were significantly rifted in the Devonian with an emplacement of large volumes of basaltic magmas into early Proterozoic cratonic lithosphere (Artemieva, 2003). The Tanzanian, Wyoming and the Sino-Korean cratons are other examples of lithospheric keels that were eroded and metasomatised during large-scale lithosphere-mantle interaction, probably associated with mantle plumes (e.g., Eggler et al., 1988; Griffin et al., 1998a,b; Lee and Rudnick, 1999).

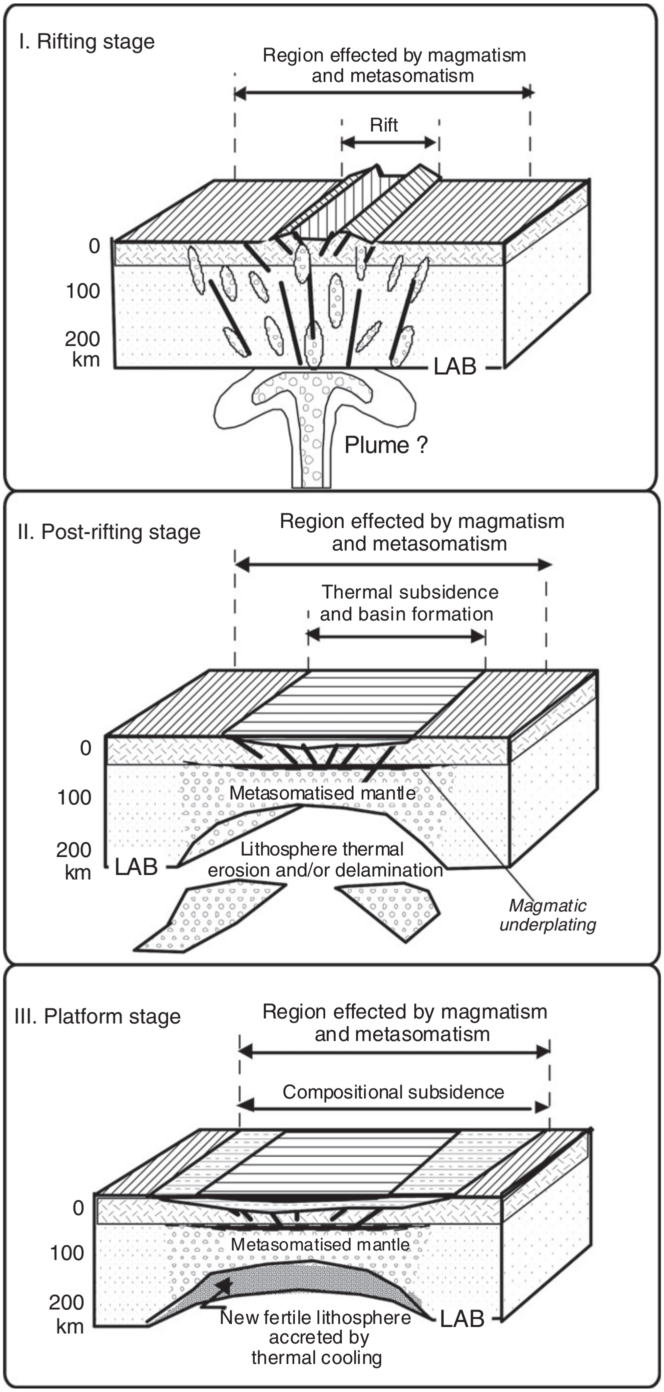

Figure 5.8 Reworking of the cratonic lithosphere of the southern part of the East European Platform *(after Artemieva, 2003).* (I): An interaction of a mantle plume with the cratonic lithosphere causes rifting, intensive magmatism, and metasomatic reworking of the lower lithosphere. (II): Thermal cooling at the post-rifting stage leads to thermal subsidence and basin formation. The lower part of the lithospheric mantle may be removed by thermal erosion and/or by delamination of a dense metasomatised lowermost lithosphere. (III): Further subsidence is a result of fertilisation of cratonic lithosphere during mantle-plume interaction and involves a much larger area than affected by rifting. This process is accompanied by an accretion of a new basal part of the lithosphere, with the fertile composition typical for Phanerozoic regions.

5.5 Correlations between lateral and depth extents of cratonic lithosphere and plate motions

Different styles of interaction of mantle convection with thick and thin lithospheric keels, as revealed in numerical simulations, suggest that vertical and lateral dimensions of cratons should be correlated. Indeed, there is a strong correlation between the lateral size of the Archean cratons (or the total cratonic area) and the lithospheric thickness (Fig. 5.9A and C): large cratons have thick roots; however, this correlation does not hold for the Proterozoic parts which have a more uniform lithospheric thickness (Fig. 5.9B).

Cratonic roots with a large thickness should interact with plate motion. On one hand, a thick lithospheric root could impede a continental plate and keep

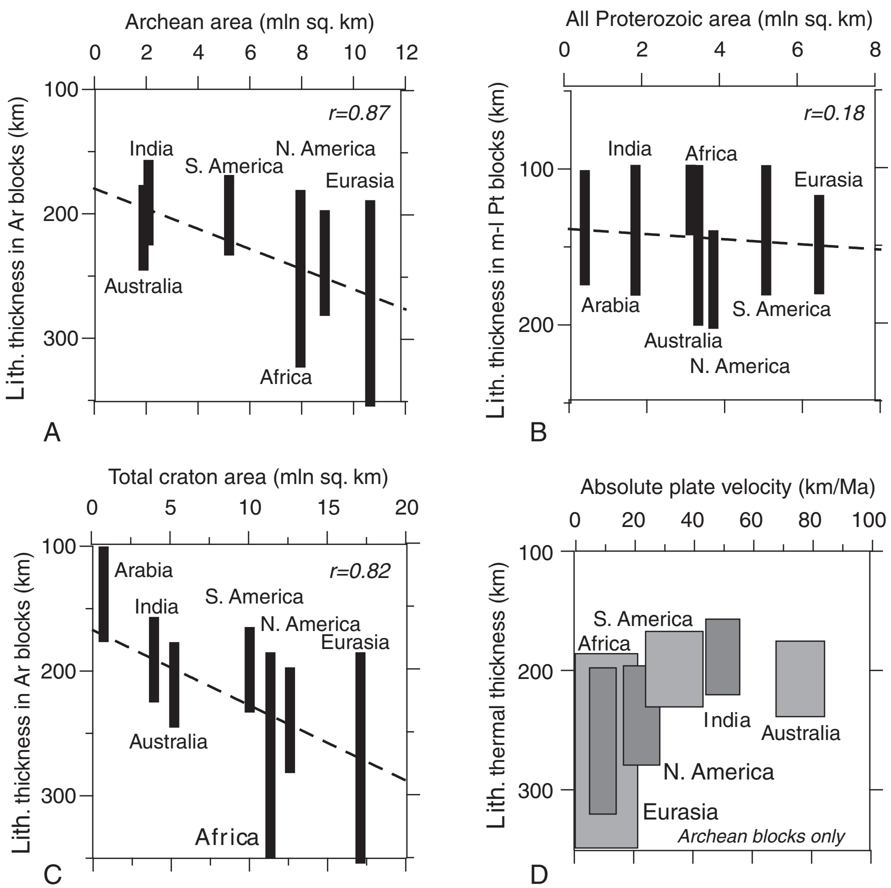

Figure 5.9 Lithosphere thermal thickness in Archean cratons versus area of Archean (A) and middle-late Proterozoic cratons (B); (C) is for the total area of the cratons (*r* is the correlation coefficient; *from Artemieva and Mooney, 2002*). The range of lithospheric thickness for some cratons (e.g., South America and northern Canadian Shield) may be artificially reduced by the absence of heat flow measurements in the regions where the cratons are expected to have their maximum lithospheric thickness from seismic tomography data. Cratons with the lateral size less than 6–8 × 10^6 km^2 have lithosphere thickness ca. 200 km (D). Thickness of cratonic lithosphere versus absolute plate velocity *(from Artemieva and Mooney, 2002)*. Boxes correspond to different cratons; their vertical dimensions – range of lithospheric thickness and horizontal dimensions – the error in plate velocity estimates.

it immobile. On the other, shear forces at the lithospheric base because of a relative movement of a lithospheric plate above the underlying asthenospheric mantle would reduce the depth extent of lithospheric roots. Xenolith data, which suggests a globally nearly uniform thickness of cratonic lithosphere, seem to support the latter hypothesis, that is, that plate motion has eroded all cratonic roots to about the same thickness.

A global analysis of the correlation between thickness of Precambrian continental lithosphere and plate velocity shows that plates having deep Archean keels move slowly (Fig. 5.9D; the correlation coefficient $r = -0.77$ including Australia and $r = -0.97$ without Australia; Artemieva and Mooney, 2002). Australia does not follow the general trend because it is close to a subduction zone and has a high plate velocity (75 mm/yr) because of strong subduction pull. Flattening of the curve at the depth of ca. 200 km (where there is no correlation between lithospheric thickness and plate velocity) suggests that the base of the low-viscosity asthenospheric layer can be reached at this depth. Thus, thin Proterozoic keels located within the asthenospheric low-velocity layer are not affected by plate motion.

An inverse correlation between the two main driving forces of plate motion, subduction pull and ridge push (Forsyth and Uyeda, 1975) and lithospheric thickness in Archean cratons implies that while these two main forces determine plate velocities, thick lithospheric roots of Archean cratons can be eroded by shear forces at the lithospheric base, which are proportional to the plate velocity (Artemieva and Mooney, 2002). This conclusion has important implications for paleoreconstructions as it implies that Archean cratons with present-day thick lithosphere have never been a part of fast moving plates.

The thickness of the early lithosphere is unknown. Was the thickness the same over the globe or was it more like the thickness observed at present? Answering these questions is critical for the debates regarding the formation and evolution of the Archean lithosphere. Extrapolation of the trend shown in Fig. 5.9A permits speculation on the lithospheric thickness of a hypothesised supercontinent during the Archean and to thus make constraint on the mechanisms of early lithosphere formation. Assuming that only one supercontinent existed at some time in the Archean with the size equal to the total surface area of all present Archean cratons, the thickness of cratonic lithosphere during the time of an Archean supercontinent is estimated to be 350–500 km (Fig. 5.10). Similarly, for the cratonic parts which were amalgamated into Gondwana, the lithospheric thickness could be about 280–400 km at ~550–500 Ma. The trend in lithosphere thickness change in the Slave Craton since the time of Gondwanaland until the present agrees with the implied global trend in lithosphere thinning from Archean to the present (Fig. 5.10). It favours a model, in which the Archean cratons had an initial lithospheric thickness of ~450 km, while the present-day variations in their lithospheric thickness result from selective erosion of cratonic lithosphere.

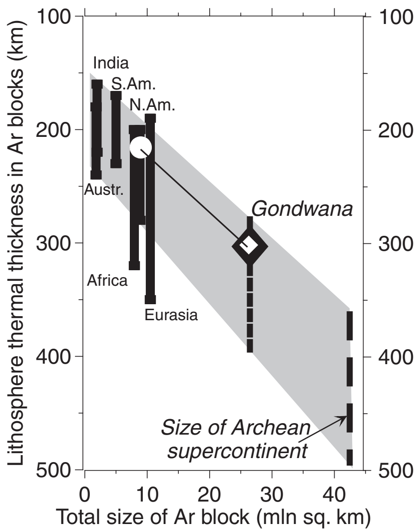

Figure 5.10 Lithosphere thermal thickness in the Archean cratons versus craton area *(from Artemieva and Mooney, 2002)*. The extrapolation of the linear trend in Fig. 5.6a provides an estimate of the thickness of the lithosphere in the early Archean. For some Archean cratons, the range of lithospheric thickness may be artificially reduced by the absence of heat flow measurements in the regions where the cratons are expected (from seismic tomography) to have their maximum lithospheric thickness. White diamond – estimated lithospheric thickness in the Slave Craton at the time of Gondwanaland (Pokhilenko et al., 2001); white dot – typical values of the present-day lithospheric thickness in the Slave province.

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