

On the relations between cratonic lithosphere thickness, plate motions, and basal drag

Irina M. Artemieva^{a,*}, Walter D. Mooney^b

^a*Department of Earth Sciences, Uppsala University, Villavägen 16, S-756 32 Uppsala, Sweden*

^b*USGS MS 977, 345 Middlefield Rd., Menlo Park, CA 94025, USA*

Received 29 June 2001; accepted 15 June 2002

Abstract

An overview of seismic, thermal, and petrological evidence on the structure of Precambrian lithosphere suggests that its local maximum thickness is highly variable (140–350 km), with a bimodal distribution for Archean cratons (200–220 km and 300–350 km). We discuss the origin of such large differences in lithospheric thickness, and propose that the lithospheric base can have large depth variations over short distances. The topography of Bryce Canyon (western USA) is proposed as an inverted analog of the base of the lithosphere.

The horizontal and vertical dimensions of Archean cratons are strongly correlated: larger cratons have thicker lithosphere. Analysis of the bimodal distribution of lithospheric thickness in Archean cratons shows that the “critical” surface area for cratons to have thick (>300 km) keels is $>6-8 \times 10^6$ km². Extrapolation of the linear trend between Archean lithospheric thickness and cratonic area to zero area yields a thickness of 180 km. This implies that the reworking of Archean crust should be accompanied by thinning and reworking of the entire lithospheric column to a thickness of 180 km in accord with thickness estimates for Proterozoic lithosphere. Likewise, extrapolation of the same trend to the size equal to the total area of all Archean cratons implies that the lithospheric thickness of a hypothesized early Archean supercontinent could have been 350–450 km decreasing to 280–400 km for Gondwanaland.

We evaluate the basal drag model as a possible mechanism that may thin the cratonic lithosphere. Inverse correlations are found between lithospheric thickness and (a) fractional subduction length and (b) the effective ridge length. In agreement with theoretical predictions, lithospheric thickness of Archean keels is proportional to the square root of the ratio of the craton length (along the direction of plate motion) to the plate velocity. Large cratons with thick keels and low plate velocities are less eroded by basal drag than small fast-moving cratons.

Basal drag may have varied in magnitude over the past 4 Ga. Higher mantle temperatures in the Archean would have resulted in lower mantle viscosity. This in turn would have reduced basal drag and basal erosion, and promoted the preservation of thick (>300 km) Archean keels, even if plate velocities were high during the Archean.

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Keywords: Continental lithosphere; Plate motions; Archean; Proterozoic; Lithosphere thickness

* Corresponding author. Now at EOST-IPGS, 5 Rue Rene Descartes, F-67084 Strasbourg cedex, France.

E-mail addresses: Irina.Artemieva@eost.u-strasbg.fr, iartem@geofys.uu.se (I.M. Artemieva), mooney@usgs.gov (W.D. Mooney).

1. Introduction

We discuss evidence for and against thick lithospheric keels, and examine mechanisms that may thin the lithosphere. The existence of thick, chemically depleted lithospheric keels beneath continents was first hypothesized on the basis of seismic data that showed high seismic velocities to depths of 300–500

km (Jordan, 1975, 1979, 1988). This tectosphere hypothesis has been discussed and refined for more than two decades.

Global seismic tomography supports thick (~ 400 km) keels beneath Precambrian cratons (defined by a +0.5% shear-wave velocity anomaly), and higher-resolution regional tomographic studies of the cratons show a +1.0% seismic velocity anomaly extending to

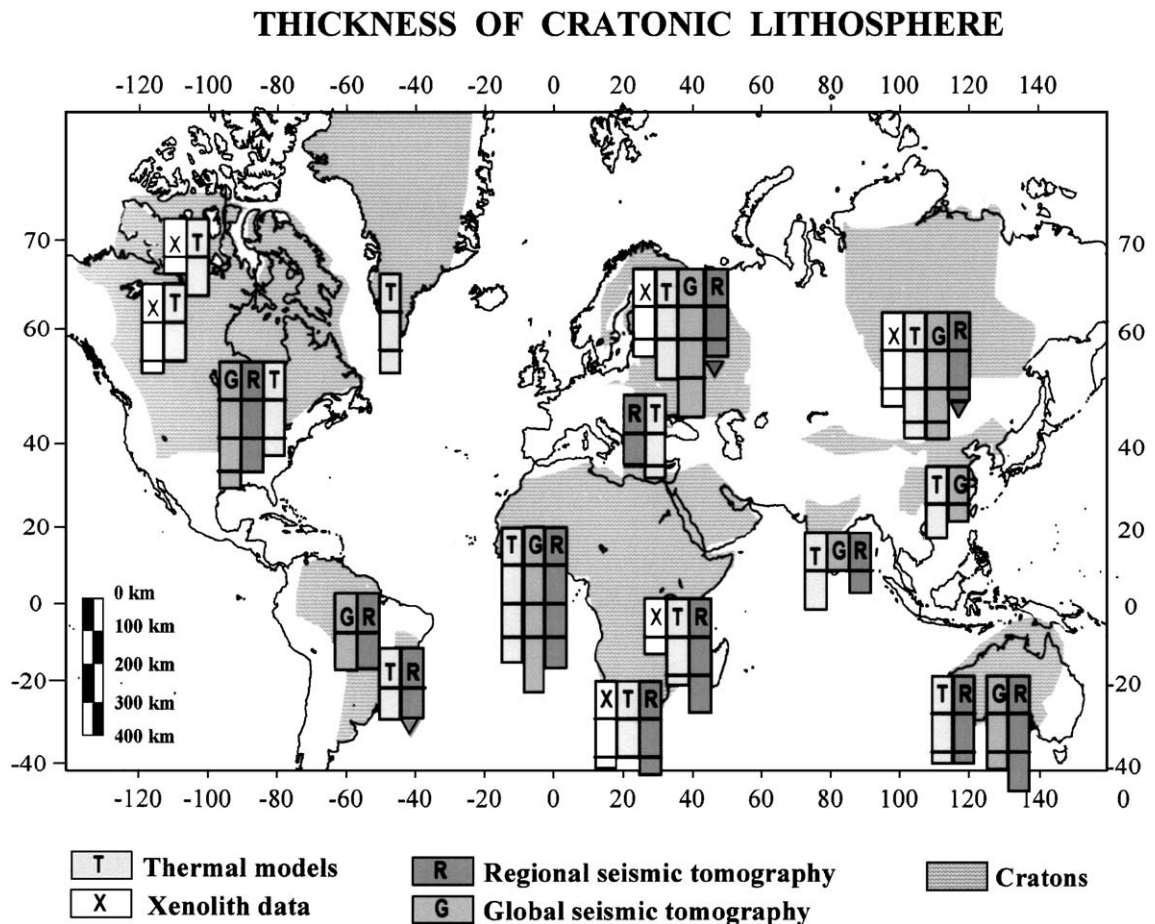


Fig. 1. Thickness of the cratonic lithosphere (vertical bars) as determined by four different methods. Gray shading shows the outlines of the cratons (after Goodwin, 1996). Global seismic tomography (G) usually gives the largest (350–400 km) values; estimates from regional seismic tomography (R) and thermal studies (T) are also large (300–350 km) and show strong variability. Xenolith data (X) suggests a thinner and more nearly uniform thickness of the cratonic lithosphere (200–250 km), and thus locally differs from geophysical estimates. Sources: Thermal thickness estimates—Artemieva and Mooney (2001). Xenolith data—Boyd et al. (1985, 1995); Boyd and Gurney (1986); Chesley et al. (1999); Griffin et al. (1996, 1999); Kopylova et al. (1999); Kukkonen and Peltonen (1999); Nicolaysen et al. (1981); Pokhilenko et al. (1993); Rudnick and Nyblade (1999); Ryan et al. (1996); Schmidberger and Francis (1999). Global seismic tomography (for +0.5% velocity anomaly)—Zhang and Tanimoto (1993); Grand (1994). Regional seismic tomography (for +1.0% velocity anomaly)—Van der Lee and Nolet (1997); Ritsema et al. (1998); Ritsema and van Heijst (2000); Ritzwoller and Levshin (1998); Simons et al. (1999); Vdovin et al. (1999).

250–350-km depth (Fig. 1 and references therein). Thermal estimates of Precambrian continental lithospheric thickness (defined as the depth to the intersection of the geotherm with a 1300 °C mantle adiabat) range from 140 to 350 km (Artemieva and Mooney, 2001). Thickness values greater than 250 km are relatively rare, but are supported by high-quality heat flow data from Archean cratons. In contrast, geotherms based on xenolith studies have been interpreted to suggest a relatively uniform cratonic lithosphere thickness of 200–250 km (Fig. 1 and references in the caption).

Numerical simulations of mantle convection with an overlying, depleted cratonic lithosphere (Doin et al., 1997) imply that there are two equilibrium thickness values for the lithosphere, one ~ 350 km and the other ~ 220 km. These simulations suggest that convective erosion of thick (>300 km) lithosphere primarily occurs from the sides, thereby decreasing the lateral dimension of the lithospheric keel. Conversely, convective erosion of thin (<250 km) lithosphere primarily occurs from below, thereby maintaining the lateral dimension but decreasing lithospheric thickness to the equilibrium value of ~ 220 km. Recent thermal modeling (Artemieva and Mooney, 2001) is consistent with the results of Doin et al. (1997) and shows that, in general, lithospheric thickness increases with the age of the overlying crust (Fig. 2). Archean regions are unique in that they have two typical thicknesses (>300 and 200–220 km).

We try to explain such large variations in cratonic lithosphere thickness and investigate the mechanisms of lithospheric erosion. We test the hypothesis that plate motions influence the size of deep lithospheric keels by examining correlations between the thickness of Precambrian continental lithosphere and plate velocities. As suggested by Chapman and Pollack (1974), who estimated a lithospheric thickness of >400 km beneath West Africa, such a thick keel could impede plate motion and keep the African plate immobile. In contrast to geophysical data, xenolith data, which suggests a globally nearly uniform lithospheric thickness, implies that plate motion does not have any influence on keel thickness, or conversely, that plate motion has eroded all keels to about the same thickness.

In this paper we examine the correlation between the lithospheric thermal thickness and the area of

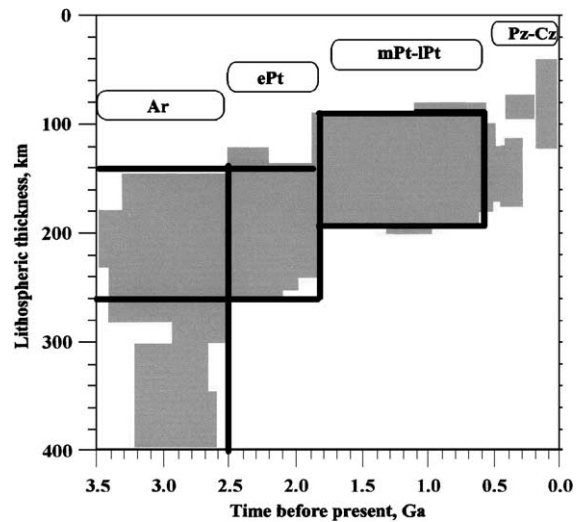


Fig. 2. Lithospheric thermal thickness versus geologic age of the continental lithosphere. The Archean lithosphere has bimodal thickness distribution centered at ~ 350 and ~ 220 km. Gray area shows the lithospheric thickness estimates derived from thermal data (Artemieva and Mooney, 2001). Key: Ar—Archean; ePt, mPt, lPt—early, middle and late Proterozoic, respectively; Pz—Paleozoic; Cz—Cenozoic.

the cratons of different age; estimate lithospheric thickness of a hypothesized early Archean supercontinent; discuss the effect of continental break-up on lithosphere erosion; and evaluate the basal drag model (Forsyth and Uyeda, 1975; Sleep, submitted for publication) using the measured values of lithospheric thickness and plate velocities.

The most comprehensive set of lithospheric thickness estimates for continental regions comes from thermal (heat flow) data, since regional seismic tomography models are not globally available. Our study is based on the recent global calculations of thermal thickness of Precambrian lithosphere (Artemieva and Mooney, 2001). However, as the depth difference between the bottom of the conductive thermal boundary layer (i.e., lithosphere thermal thickness) and the top of the convective mantle (i.e., thickness of seismic lithosphere) is less than 50 km (Jaupart et al., 1998), we expect good agreement between the analysis based on thermal and seismological estimates of lithospheric thickness.

A discrepancy in estimated values of lithospheric thickness follows from the fact that different ap-

proaches employ various definitions of the lithosphere and address its different properties (e.g., mechanical, compositional, rheological) (e.g., Anderson, 1989). We, moreover, propose that the lithospheric base is not a smooth subhorizontal layer as it is usually assumed, but can have a large thickness variations over short distances. The topography of Bryce Canyon in western USA can be thought as an inverted analog of such an eroded structure (Fig. 3). If this is the case, different techniques may sample different parts of the lithospheric base: while xenoliths come from the shallowest parts, seismic studies will be more sensitive to the bottom of the deepest keels below which the mantle is convecting. Heat flow data provides an averaged image of the lithospheric basal geometry and thus thermal constraints give the values intermediate between xenolith and seismic studies. In order to reconcile geophysical estimates of lithospheric thickness with those derived from petrology, we hypothesize that xenolith-based values are not representative of a large-scale lithospheric thickness, but rather refer to a localized thinning of the

lithosphere, associated with low-percentage melting and generation of kimberlite-type magmas. This inference is supported by some petrologic studies (e.g., Thomsen, 1975; Mitchell et al., 1980).

2. Continental lithosphere: relation to plate size and thickness in the Archean

2.1. Lithospheric thermal thickness versus plate size

We examine the correlation between lithospheric thermal thickness and the lateral extent of cratons (Fig. 4). Table 1 summarizes data on their sizes and plate velocities. Following the assumptions of Stoddard and Abbott (1996), we accept here that the Indo-Australian plate is composed of several other plates (e.g., Gordon et al., 1998) and specifically that the Indian and Australian cratons belong to different plates.

We find a positive correlation between craton size and keel thickness for Archean blocks: cratons with a larger Archean area have deeper keels (Fig.

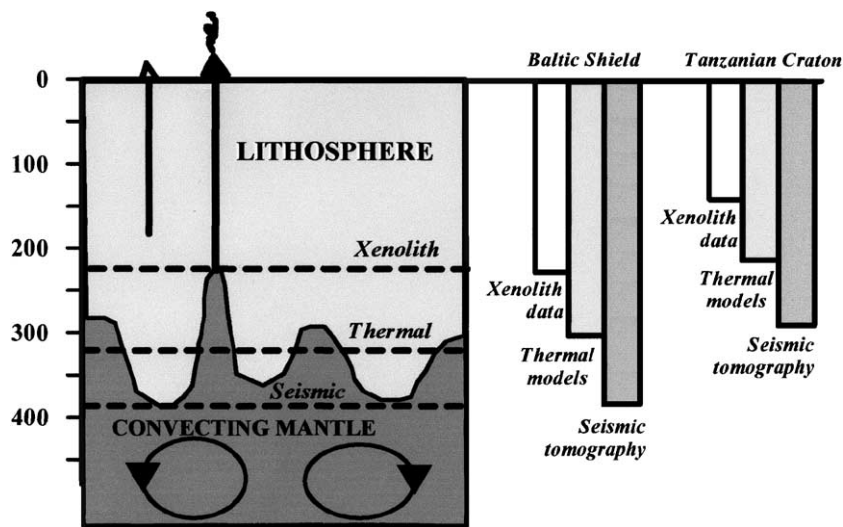


Fig. 3. The “Bryce Canyon” model of the lithospheric structure. The lithospheric base is not a flat sub-horizontal boundary as usually assumed, but an undulating boundary due to a heterogeneous erosion of the lithospheric base by an interaction with mantle plumes and its nonuniform destruction by convection flow due to compositional inhomogeneities. Xenolith data sample the most shallow parts of the lithospheric base (200–250 km), where pressures are low enough to initiate low-percentage melting and produce kimberlite-type magmatism. Seismic tomography samples the top of the convective mantle, i.e. the lowermost parts of the lithospheric base. A diffuse character of the seismic lithospheric base, especially based on surface waves, supports the model. Thermal data provide a smoothed integrated picture of the lithospheric structure and thus give values of the thickness of cratonic keels between xenolith and seismic estimates. The right part of the figure gives an example of the lithospheric thickness estimates for the Baltic Shield and the Tanzanian craton (compare with Fig. 1).

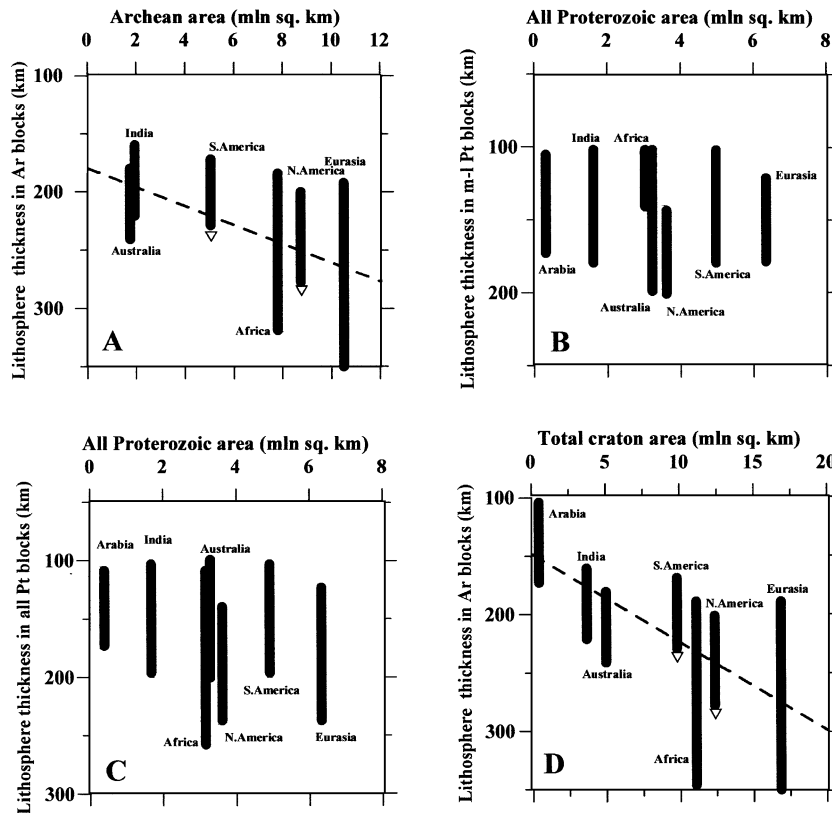


Fig. 4. Lithosphere thermal thickness in Archean cratons versus area of Archean (A), (B) middle–late Proterozoic, (C) early–late Proterozoic and (D) combined cratons. Plots A and D show that Archean cratons with larger size have thicker lithospheric keels. No correlation exists for middle–late Proterozoic cratons (B) because lithospheric thickness is more uniform. The values of the surface area of the cratons are listed in Table 1 and are based on the compilation of Stoddard and Abbott (1996), and the values of the lithosphere thermal thickness are based on the results of Artemieva and Mooney (2001) (Table 2). Note that heat flow data are not available for the Archean Amazon craton (South America), nor for the central portion of the Canadian Shield (Hudson’s Bay region) (Fig. 4 in Artemieva and Mooney, 2001). Thus, the range of lithospheric thickness for South America and North America may be artificially reduced by the absence of data where the cratons are expected to have their maximum lithospheric thickness (shown by arrows).

4A). This means that the horizontal and vertical dimensions of the Archean cratons are correlated (Tables 2 and 3). Noting that Archean lithosphere has two typical keel thicknesses, >300 and 200 – 220 km (Artemieva and Mooney, 2001), the empirical “critical” surface area for Archean cratons to have thick (>300 km) keels is >6 – 8×10^6 km² (Fig. 4A).

The positive correlation between the horizontal and vertical dimensions of cratons still holds strong ($r=0.82$) when the entire cratonic area versus thickness of its deepest, Archean, part is examined (Fig. 4D). However, this correlation does not exist for the middle–late Proterozoic cratons (Fig. 4B): the thick-

ness of the middle and late Proterozoic lithosphere is globally around 120 – 180 km (Fig. 2 and Artemieva and Mooney, 2001), and thus no discernible dependence can be found between its thickness and cratonic size. However, a weak correlation appears when data for the early Proterozoic lithosphere are added into the analysis (Fig. 4C), supporting the results of the thermal modeling (Fig. 2) that suggest that early Proterozoic lithosphere is transitional between Archean and middle to late Proterozoic lithospheres in terms of both its thickness and the correlation between the area and keel depth.

Following Stoddard and Abbott (1996), we also consider lithospheric thickness versus the percentage

Table 1
Sizes of Precambrian cratons and their plate velocities (from Stoddard and Abbott, 1996)

Craton	Archean (Ar) part		Proterozoic (Pt) part		Ar + Pt part		Absolute plate velocity V , km/Ma
	Area, 10^6 km ²	Fraction of plate, %	Area, 10^6 km ²	Fraction of plate, %	Area, 10^6 km ²	Fraction of plate, %	
Africa	7.942	10.2	3.189	4.1	11.131	14.4	8.941 ± 2.980
Australia	1.802	3.8	3.248	6.8	5.05	10.6	75.119 ± 6.919
Eurasia	10.500	15.6	6.283	9.3	16.783	24.9	9.448 ± 9.448
India	1.963	16.3	1.689	14.1	3.652	30.4	48.620 ± 4.420
N. America	8.846	14.8	3.548	5.9	12.394	20.8	20.617 ± 5.891
S. America	5.008	11.8	4.913	11.5	9.921	23.3	33.036 ± 8.259
Arabia	0.0	0.0	0.333	6.7	0.333	6.7	28.724 ± 2.666
Antarctica	5.806	10	0.952	1.6	6.758	11.6	10.311 ± 8.436
Rodinia	~ 42.0	–	–	–	–	–	–
Total	41.867	–	24.155	–	66.022	–	–

of the plate area that is composed of cratonic crust. The lack of a significant correlation between the thickness of Archean lithospheric keels and normalized Archean, Proterozoic and total cratonic areas (Table 3) indicates that only the actual area of the cratons, not the normalized area, plays an important role in the preservation of thick keels.

A high correlation between the area of Archean cratons and the typical thicknesses of their lithospheric keels allows for some speculation regarding keel thickness in the cratons where geophysical data are sparse. For example, heat flow measurements in the Archean crust of South America are available only for the São Francisco craton. These data suggest that lithospheric thickness there is only 170–200 km (Artemieva and Mooney, 2001). However, given the paucity of heat flow data, this result should be viewed with caution. The low values of estimated South American Archean keel thickness are significantly above the best-fitting line in Fig. 4A and suggest that

lithospheric thickness in the Archean Amazon craton may be larger (~ 200–250 km). Similarly, for the Archean parts of the Antarctic craton (Table 1), where heat flow measurements are absent, a lithospheric thickness of around 220–240 km can be estimated based on the area of the craton.

When no Archean lithosphere is present (zero area of Archean craton in Fig. 4A), extrapolated lithospheric thickness tends to be about 180 km, in agreement with the observation that post-Archean lithosphere is nearly always thinner than this value. Many Proterozoic regions are composed of reworked Archean crust (Goodwin, 1996). If such Proterozoic regions were only slightly reworked Archean lithosphere, then the original Archean lithospheric thickness of >200 km would have been preserved. Since this is not the case, we conclude that the reworking (i.e., strong deformation and intrusion of melts and volatiles) of Archean lithosphere also involves lithospheric erosion and possibly chemical modification

Table 2
Lithosphere thermal thickness in Precambrian blocks and the length of Ar blocks

Craton	Range of lithosphere thermal thickness, km			Length of Ar block L_{Ar} , km	Total (Ar + Pt) length of craton L , km	$\sqrt{\frac{L_{Ar}}{V}}$, Ma ^{1/2}	$\sqrt{\frac{L}{V}}$, Ma ^{1/2}
	Ar	ePt	m–l Pt				
Africa	180–330	200–260	100–140	4500	6400	22.434	26.754
Australia	180–240	140–200	100–200	2000	3200	5.159	6.527
Eurasia	180–350	150–250	120–180	13,000	13,000	37.094	37.094
India	170–220	170–220	100–180	2400	3600	7.026	8.605
N. America	200–280	160–240	140–200	11,100	11,100	23.203	23.203
S. America	170–220	150–200	100–180	3700	3700	10.582	10.582
Arabia	–	–	100–180	–	1400	–	6.981

Table 3
Correlation coefficients between lithospheric thickness, plate velocities, typical (mean) cratonic size and length

Lithospheric thickness	For craton area (km ²)	For craton area (% of plate size)	For craton length	For plate velocity
Archean lithosphere	0.87	0.12	0.68	−0.77 (−0.97 without Australia)
Middle–Late Proterozoic lithosphere	0.18	−0.01	0.46	0.15
All Proterozoic lithosphere	0.54	−0.44	–	−0.64
Cratonic lithosphere (thickness in Ar blocks)	0.81	−0.27	0.87	–

(metasomatism) to a more enriched state (Griffin et al., 1998; O'Reilly et al., 2001). Thus, our results imply that the reworking of Archean crust should be accompanied by thinning and reworking of the entire lithospheric column.

2.2. Estimating lithospheric thermal thickness during the Archean

Why do present-day Archean cratons have two characteristic values of lithospheric thickness, 300–350 and 200–220 km? Did all Archean cratons have the same keel thickness at the time of their formation and were some keels selectively eroded with time, or were Archean cratons originally formed with these two characteristic values? In order to investigate these questions, we estimate the lithospheric thickness of a hypothesized supercontinent during the early Archean. We make three simplifying assumptions: (1) we accept the observed positive correlation between lithospheric thickness and cratonic surface area (Fig. 4A and D) for the Archean; (2) we assume that one supercontinent existed during the early Archean, and (3) that its size was equal to the total surface area of all present Archean cratons (Table 1). Extrapolating the trend shown in Fig. 4A under these assumptions, we estimate a 350–500-km thickness of cratonic lithosphere during the time of an early Archean (4.2–4.0 Ga) supercontinent (Fig. 5). Our speculation

contradicts the suggestion of Davies (1979) that Archean cratons had a constant thickness through geological history because they are chemically depleted, and thus isolated from the convecting mantle. As noted below, a basal drag model is able to explain observed variations in Archean lithospheric thickness.

Similarly, we estimate lithospheric thickness of Gondwana (formed at ~ 550–500 Ma) to be about

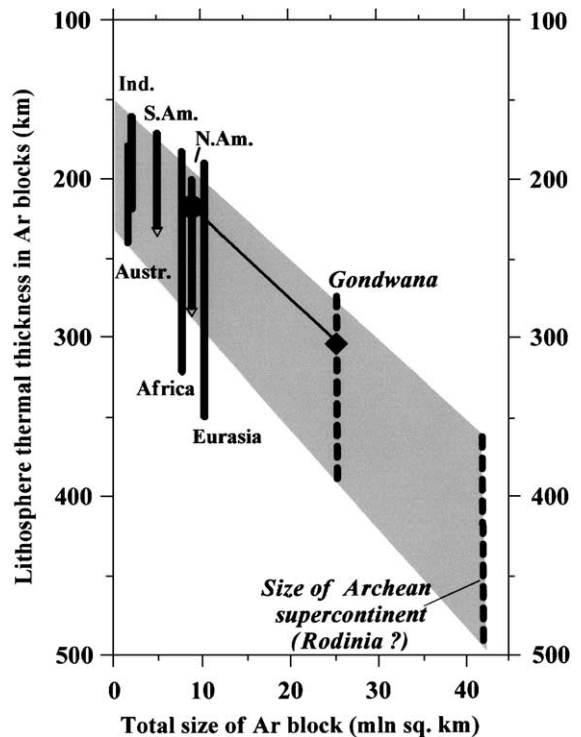


Fig. 5. Lithosphere thermal thickness in the Archean cratons versus craton area. The range of lithospheric thickness for South America and North America may be artificially reduced by the absence of data where the cratons are expected to have their maximum lithospheric thickness (shown by arrows). The extension of the linear trend in Fig. 4A provides an estimate of the thickness of the lithosphere in the early Archean. Assuming that the area of the oldest (e.g., ~ 4.0 Ga) hypothesized Archean supercontinent was equal to the total area of the present Archean cratons, its lithospheric thickness could have been 350–450 km. Similarly, lithospheric thickness at ~ 550–500 Ma (when Gondwanaland was formed) is estimated to be about 280–400 km. Black diamond shows estimated lithospheric thickness in the Slave Craton at the time of Gondwanaland (Pokhilenko et al., 2001; see text for explanations). For a comparison, black dot shows typical estimates of the present-day lithospheric thickness in the Slave province.

280–400 km (Fig. 5). This range of lithospheric thickness well agrees with the results of the recent study of crystalline inclusions in diamonds and pyrope compositions in the Slave Craton, which show that at ~ 523 Ma (i.e., at the time of Gondwanaland assemblage) lithospheric thickness there was about 300 km (Pokhilenko et al., 2001). Noting that xenolith-based estimates of the present-day lithospheric thickness in the Slave Craton vary from 180 to 240 km (Griffin et al., 1999; Kopylova et al., 1999; Rudnick and Nyblade, 1999), the trend in the lithospheric thinning in the craton from the Cambrian to present is similar to the general trend outlined in Fig. 5 for all of the cratonic keels since Archean to present.

A proposed maximum thickness of ~ 450 km in the early Archean is reasonable based on the following arguments. The mantle potential temperature of 1300 °C, usually associated with the lithospheric base, corresponds to experimental estimates for the phase transition temperature at 410 km (Ito and Takahashi, 1989). And thus the present-day 410-km mantle transition puts a natural limit on the base of the thermal lithosphere. Assuming that the early Archean mantle temperature was 150 – 200 °C higher than at present (McGovern and Schubert, 1989; Abbott et al., 1994), the endothermic olivine–spinel phase transition at 410-km depth would have been some 30 km deeper in the Archean than at present, i.e., at a depth of about 440–450 km. Thus, the thermal lithosphere may have extended to this depth at the time of the formation of an early Archean supercontinent. This accords with Ballard and Pollack (1988), who modeled the thermal structure of the South African lithosphere during the Archean, accounting for higher heat production, heat flow and mantle temperatures at that time. These authors suggest that an initial lithospheric thickness beneath the Kaapvaal craton could have been as large as 400–500 km.

The mechanism by which the early Archean lithosphere formed is one of the outstanding problems in the history of the Earth. Alternate hypotheses include melting processes such as crustal extraction from a primitive upper mantle source (e.g., Campbell et al., 1989), and processes related to plate tectonics (e.g., the stacking of oceanic lithospheric plates (Fyson and Helmstaedt, 1989; Abbott and Mooney, 1995)). In some cratons (e.g., the Slave Craton in North America and the Gawler Craton in southern Australia), petro-

logical data suggest a multi-stage formation of the lithospheric keel in the Archean (Gaul et al., 2000). A discussion of all these mechanisms is beyond the scope of the present paper, however, we emphasize that constraints on the initial thickness of the lithosphere in the early Archean—either ~ 250 or ~ 450 km, as proposed here—are critical to this debate.

Based on the results of Fig. 5, we favor a model whereby the present-day variations in the thickness of the Archean cratonic keels result from selective erosion of an Archean lithosphere that had an initial uniform thickness of ~ 450 km. We next examine possible mechanisms for lithospheric erosion.

3. Erosion of the thermal lithosphere

3.1. Mechanisms of lithospheric erosion

If the early Archean cratonic lithosphere was more than 400-km thick, how and why was it eroded? Even if we cannot be certain regarding the initial thickness of the Archean lithosphere, a mechanism is needed to explain the present-day bimodal thickness of ~ 350 and ~ 220 km of the Archean keels. Furthermore, Proterozoic cratons, which have a lithospheric thickness of 100–180 km (Fig. 2), may also have been eroded, since secular cooling would tend to thicken the lithosphere.

Various possible mechanisms to erode the lithosphere have been proposed. They include: (1) thermo-mechanical erosion by large-scale and vigorous secondary mantle convection (e.g., Fleitout and Yuen, 1984; Yuen and Fleitout, 1985; Olson et al., 1988; Doin et al., 1997); (2) thermo-mechanical erosion by mantle plumes (e.g., Davies, 1994; Ribe and Christensen, 1994; Sleep, 1994), which includes thermal thinning by conduction due to an increase of mantle heat flow at the lithospheric base (e.g., Crough and Thompson, 1976; Spohn and Schubert, 1982, 1983) and mechanical thinning by delamination due to gravitational instabilities in the lower lithosphere (e.g., Bird, 1979; Mareschal, 1983); and (3) thermo-mechanical erosion by basal drag, whereby the lower parts of the lithosphere are heated by stirring or friction due to the horizontal movement of the craton over the underlying mantle (e.g., Schubert and Turcotte, 1972; Hager and O'Connell, 1981; Sleep, submitted for publication).

Lithospheric erosion by mantle plumes can result from magmatic intrusions, ponding, and/or underplating of plume material beneath the lithosphere if these processes are accompanied by conductive thinning of the lithosphere and/or lithospheric delamination due to gravitational instability. The dominance of one mechanism over others is determined by mantle viscosity (Neugebauer, 1983; Sleep, 1994). During the Archean, higher heat flow from the core–mantle boundary (where most large plumes may originate) and/or a larger cratonic size, may have led to mantle plumes impinging on the base of the lithosphere more frequently, leading to a pronounced basal erosion and a rapid break-up of supercontinents (e.g., Courtillot et al., 1999; Dalziel et al., 2000).

Furthermore, the presence of abundant diamonds of the superdeep origin (derived from depth of at least 670 km (Davies et al., 1999)) in the kimberlites from the Slave, the Gawler, and the Kaapvaal cratons (Gaul et al., 2000) suggests a modification of the lithosphere of the cratons with the present-day thin (~ 200 – 220 km) keels by lower mantle plumes. An interaction of lower mantle plumes with lithosphere of these cratons could have resulted in the removal (delamination) of an essential (up to ~ 200 km) lower part of their keels, followed (as suggested by their layered lithospheric structure; Gaul et al., 2000) by an addition of a 70–80-km-thick layer of lithospheric material of the lower mantle paragenesis. However, the superdeep diamonds were not found up to date in the thickest (~ 350 km) parts of the Siberian cratonic lithosphere (the Daldyn–Alakit region) (Gaul et al., 2000). An absence of evidence for an interaction of the present-day thickest cratonic lithosphere with a lower mantle plume supports our hypothesis that the initial lithospheric thickness of all of the Archean cratons was at least 350 km.

Beginning with an Archean supercontinent with a 450-km-thick keel, we hypothesize the following model for a supercontinent break-up and lithospheric erosion (Fig. 6). The supercontinent is split into two continents of unequal size, for example due to interaction with a mantle plume. However, we note that a study of Gondwanaland break-up showed that plumes are not the ultimate driving force for a break-up of a supercontinent and plate boundary forces associated with changes in a subduction dipping may initiate continental break-up (Storey, 1995). Alternatively,

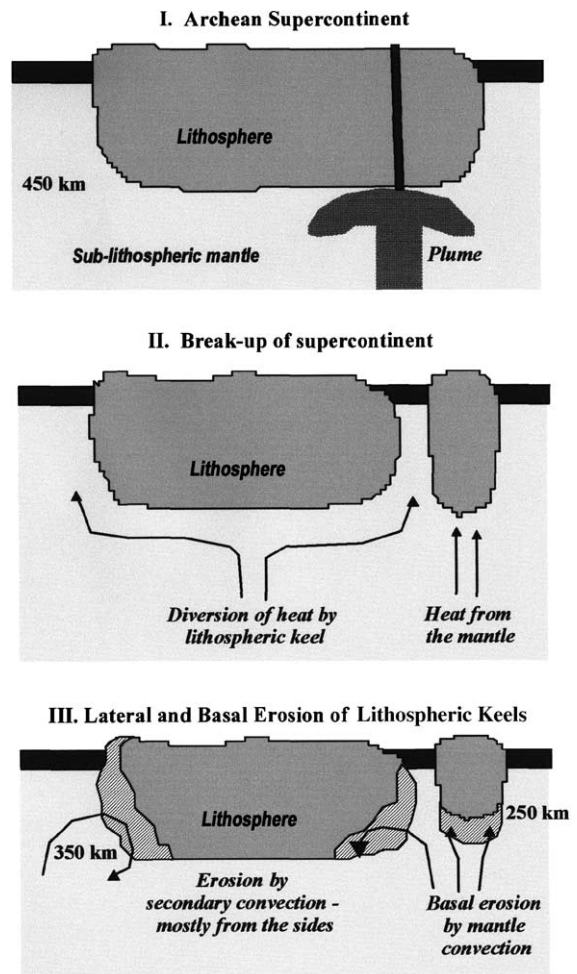


Fig. 6. Cartoon depicting the process of erosion of cratonic lithosphere. An early Archean supercontinent with a keel down to the mantle transition zone (400–450 km) is split by a mantle plume into two parts with nonequal dimensions. The fate of these supercontinent fragments is determined by their lateral dimensions. The larger craton diverts the mantle heat from its base (Ballard and Pollack, 1987) and is mostly affected by secondary convection on its margins (Doin et al., 1997). This process promotes the preservation of a thick keel (~ 350 km) in the interior of the craton, but leads to erosion of cratonic margins and thus a gradual decrease of the cratonic size. When the area of the larger craton is reduced to a critical value of about $6\text{--}8 \times 10^6 \text{ km}^2$ (Fig. 4A), it will start to evolve as a smaller craton. The smaller craton is not effective in diverting the heat because of its smaller lateral dimension and is more subject to erosion from below by mantle convection until an equilibrium lithospheric thickness of ~ 220 km is reached.

continents can be rifted and dispersed by mantle convection if their size is much greater than the depth of the convecting mantle and the mantle beneath a continental interior is not efficiently cooled by subduction (Hoffman, 1989). In this case, continents become effective thermal insulators and upwelling hot mantle can develop deviatoric stresses at the lithospheric base, large enough for a continental break-up (Gurnis, 1988; Lowman and Jarvis, 1999).

The split of the continent into two (Fig. 6) has important consequences. Based on a numerical analysis of the thermal regime of the Archean Kaapvaal craton and the surrounding Proterozoic mobile belts, Ballard and Pollack (1987) proposed that thick cratonic lithosphere diverts heat coming from the deep mantle away from the craton into the thinner surrounding lithosphere. They found that a 200–400-km-thick lithospheric root can divert enough heat to account for 50–100% of the observed contrast in surface heat flow between Archean cratons and adjacent Proterozoic terrains.

Further analysis of mantle convection models (Lenardic and Moresi, 2001) showed that the amount of heat deflected by the keel depends on the style of convection. Heat is diverted more efficiently when convection is layered. Fifty percent of an observed $\sim 20\text{--}25\text{ mW/m}^2$ increase in surface heat flow from Archean cratons to adjacent Proterozoic terrains (e.g., Ballard and Pollack, 1987) can be explained by a doubling in thickness of cratonic and non-cratonic lithosphere in models with upper-mantle ($\sim 1000\text{-km}$ deep) convection (Nyblade and Pollack, 1993). In contrast, this model requires that lithospheric thickness differs by a factor of 6 between cratons and adjacent regions in the case of whole-mantle ($\sim 2900\text{-km}$ deep) convection (Lenardic, 1997). Despite a continuing debate on the style of mantle convection (e.g., Widmer-Schnidrig, 1999), recent seismic tomography results (Van der Hilst et al., 1997; Kellogg et al., 1999) suggest that whole-mantle convection is a more accurate depiction of the present-day convection pattern.

During the Archean, higher mantle temperatures and thus higher Rayleigh numbers would have favored layered mantle convection (e.g., Condie, 1997), thereby enhancing the effect of the 660-km discontinuity. Cooling of the Earth may have resulted in a transition from layered to whole-mantle convec-

tion, most likely in early Proterozoic (2.5–1.3 Ga) (Condie, 1997). Besides, in the Archean (4.0–2.5 Ga) the cratons were not surrounded by Proterozoic mobile belts, but by oceanic lithosphere, and thus the contrast in the lithospheric thickness between the keels and the adjacent regions was much larger than later in the Earth's history. Thus, we conclude that in the Archean and probably even in the early Proterozoic, cratons were more efficient in deflecting mantle heat from their lithospheric base into the surrounding mantle.

This argument implies that the large craton in our model (Fig. 6) would be relatively more efficient, compared to the small craton, in diverting the mantle heat coming to its base and thus would be more resistant to basal thermal erosion. The efficiency in heat deflection by the keels in the Archean is supported by diamond thermobarometry data, which suggests that ancient and modern geotherms in the Archean cratons are very similar despite higher mantle temperatures and two to three times higher heat production in the Archean than at present. This problem was examined in detail by Ballard and Pollack (1988).

A positive correlation between craton size and keel thickness found for Archean blocks (Section 2.1) and the results of numerical simulations by Doin et al. (1997) suggest that the erosion of the depleted lithospheric keel of a large continent by mantle convection would primarily occur from the sides, not from below, until an equilibrium thickness of $\sim 350\text{ km}$ is reached (Fig. 6). However, if the lateral size of the large craton is less than a critical value of about $6\text{--}8 \times 10^6\text{ km}^2$ (as Fig. 4A suggests), the erosion pattern will change from lateral erosion to basal erosion, as is expected for a smaller craton (Doin et al., 1997). We note that vigorous small-scale convection at the edge of a continental keel can have another consequence, namely the extrusion of large igneous provinces, such as is observed in Norilsk, Russia, and the Deccan traps, India (King and Anderson, 1995), and intraplate hot spot volcanism as in Africa and South America (King and Ritsema, 2000). We do not include these processes in our cartoon (Fig. 6).

Due to strong basal erosion, the fate of the smaller craton will differ from that of the larger craton (Fig. 6). The small lateral extent of the keel will be insufficient to divert enough heat from the mantle

and, as a result, large values of mantle heat flow at the base of a smaller craton will lead to its rapid thinning to an equilibrium thickness of ~ 220 km (Doin et al., 1997). This is consistent with the conclusions of Ballard and Pollack (1987) that lithospheric thinning by thermal erosion will have a larger effect on heat diversion mechanism for a relatively thin cratonic keel than for a thick root.

3.2. Lithospheric erosion by basal drag

Forsyth and Uyeda (1975) examined the driving forces in plate tectonics (mantle drag, slab pull, ridge push, and transform fault resistance) and analyzed the correlations between absolute plate velocities and plate geometries (plate and continental area; ridge, trench, and transform fault length). They found that plate area and absolute plate velocities are correlated and concluded that: (1) mantle drag has the major effect on the velocity of continental plates not attached to subducting slabs and is eight times stronger under the continents than under the oceans; (2) the velocity of plates attached to subducting slabs is primarily controlled by slab pull forces and is independent of the plate surface geometry. Hence, plates connected to subducting slabs are all moving fast (at 60–90 mm/year) relative to the fixed hot spot frame. This category includes all oceanic plates and the Indian–Australian continental plate.

The analysis of Forsyth and Uyeda (1975) was extended by Stoddard and Abbott (1996), who examined correlations between the Archean and Proterozoic area of cratons and plate velocities (Table 4). They demonstrated that plates with larger surface area

of Archean crust move slower, which they attributed to basal drag by the deep (200–400 km) lithospheric keel (e.g., Zhang and Tanimoto, 1993). However, contrary to intuition, they found that plates with a high percentage of Proterozoic crustal area have higher-than-average plate velocities. Noting that the lithospheric keels are generally thinner (120–180 km) beneath Proterozoic cratons than beneath Archean cratons (200–400 km), Stoddard and Abbott (1996) suggested that the fast-moving Proterozoic keels are entirely within the low-viscosity asthenosphere, whereas the slow-moving Archean keels encounter basal drag within the higher-viscosity upper mantle.

We next extend the analysis of Stoddard and Abbott (1996) to examine how the thickness of lithospheric keels correlates with plate motions. If mantle drag is proportional to the plate area and the plate velocity relative to the underlying mantle (Forsyth and Uyeda, 1975), it should have important implications to the mechanism of thermo-mechanical erosion of the continental lithospheric plates, and one would expect a clear correlation between the thickness of cratons and plate velocity.

Depending on the relative velocity between the drifting plate and the convective mantle, mantle drag can act either as driving or resistive forces. Following the model of Forsyth and Uyeda (1975) and the results of Hager and O’Connell (1981), we assume that basal drag resists plate motion and disregard the active role of mantle flow assisting plate motion because theoretical models of mantle convection suggest that strong coupling of plate motion to active flow in asthenosphere is unlikely, and the rate of mantle flow cannot be directly measured. The fact that up to 90% of plate-driving forces result from subduction pull (Lithgow-Bertelloni and Richards, 1998) also suggests that plate motion can perhaps be coupled to mantle convection by as little as 10%, as opposed to the early ideas of Morgan (1971) and McKenzie (1972). A detailed analysis of the relations between mantle convection and plate tectonics is presented by Bercovici et al. (2000).

The basal drag model (Sleep, submitted for publication) suggests that, to the first order, the drag of the underlying mantle due to the horizontal movement of a lithospheric plate produces simple shear at the base of the lithosphere. This shearing leads to the removal of the warm, soft basal part of the cratonic keel and its

Table 4
Correlation coefficients between plate velocities and craton size calculated for the data of Stoddard and Abbott (1996)

Craton size	Correlation coefficient r for absolute plate velocity	
	Including Australia	Without Australia
Archean area (km ²)	–0.85	–0.82
Proterozoic area (km ²)	–0.12	–0.27
Cratonic area (km ²)	–0.54	–0.59
Archean area (% of plate size)	–0.48	0.49
Proterozoic area (% of plate size)	0.42	0.79
Cratonic area (% of plate size)	–0.03	0.45

replacement with sub-lithospheric mantle from adjacent regions. As initially developed, the basal drag model sought to explain the apparent uniformity of the cratonic lithosphere thickness (Sleep, submitted for publication) as suggested by xenolith data. We here adopt it to explain the large variation in thickness of Archean lithosphere.

For the case of stirring-dominated convection with vertical conduction and horizontal and vertical advection of heat, the model suggests (Sleep, submitted for publication) that the lithospheric thickness z_0 can be expressed as:

$$z_0 = \frac{T_0}{2T_A} \sqrt{\kappa t}, \quad (1)$$

where t is the time required for the craton to move a distance equal to its length relative to the underlying mantle, κ is thermal diffusivity; T_0 is the temperature at the base of the thermal boundary layer and the scaling temperature T_A defines the change in viscosity with temperature:

$$\eta = \eta_0 \exp \frac{T_0 z}{2T_A z_0}. \quad (2)$$

The velocity at the base of the thermal boundary layer is close to the plate velocity V when there is a strong depth-dependence of viscosity and the time t is close to the time needed for the plate to move the length L_{Ar} of the craton (i.e., the length in the direction of the plate movement) (Sleep, submitted for publication). Then Eq. (1) takes the form:

$$z_0 \propto C \sqrt{\frac{L_{Ar}}{V}}, \quad (3)$$

where the constant $C = 2.66 \text{ km/Ma}^{1/2}$ for the values of the parameters typical for the mantle ($\kappa = 0.8 \times 10^{-6} \text{ m}^2/\text{s}$, $T_0 = 1600 \text{ K}$, and $T_A = 43 \text{ K}$) implying viscosity changes by an order of magnitude over 100 K (Sleep, submitted for publication).

To test the validity of the basal drag model we compare the model predictions (Eq. (3)) with the present-day plate velocities, estimates of the present-day lithospheric thermal thickness (Artemieva and Mooney, 2001), and the lateral extent of cratons. Thereby, we implicitly assume that the present-day lithospheric thickness is the result of two processes:

primary formation and recent erosion. Craton length (Table 2) is measured along the path of plate motion as based on Fig. 1 of Stoddard and Abbott (1996) and the geological maps of Goodwin (1996). As discussed above, we use plate velocities relative to the deep mantle, i.e., the absolute plate velocities in the hotspot reference frame (Gripp and Gordon, 1990).

3.3. Correlation between lithospheric thermal thickness, craton length, and plate motion

Eq. (3) suggests that lithospheric thickness should correlate with the length of cratons in the direction of plate motion. For Archean cratons we find that a good correlation ($r = 0.68$; Table 3) exists between these two parameters (Fig. 7A): keel thickness in general increases with length of the Archean part of cratons. Similarly to Fig. 4A, when the Archean lithosphere is

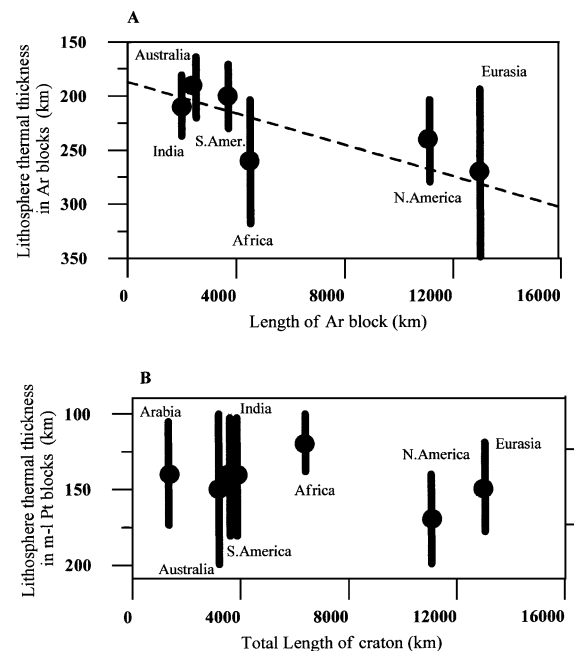


Fig. 7. Thickness of cratonic lithosphere versus craton length. Solid circles correspond to the mean values of the lithospheric thickness (Table 2). (A) Data for the Archean cratons supports the “basal drag” model, which predicts the lithospheric thickness to be proportional to the square root of the length of the craton. Large Archean cratons have thick lithospheric keels. (B) No correlation exists between the length of the Proterozoic cratons and the keel thickness since middle–late Proterozoic lithosphere has a uniform thickness.

absent (zero length of the Archean block), lithospheric thickness tends to be ~ 180 km, which is the value typical for post-Archean lithosphere. Thickness of Proterozoic lithosphere does not show any correlation with cratonic length (Fig. 7B).

According to the basal drag model (Eq. (3)), plate velocities should be related to lithospheric thickness. We examine this correlation for the present-day plate velocities and the present-day lithospheric thickness as only few constraints on the paleo-velocities exist (e.g., Irving, 1977; Lithgow-Bertelloni and Richards, 1998), and all of them are for Paleozoic–Cenozoic time. An early Archean lithosphere with a thickness of ~ 450 km would have been underlain by an upper mantle with a viscosity that was 1–2 orders of magnitude lower than the present-day mantle due to the high mantle temperature (Davies, 1999). Therefore, the plate velocity of the hypothesized early Archean supercontinent may have been significantly higher than velocities of present-day plates with thick (~ 350 km) keels, implying that lithospheric erosion by basal drag could have been different in Archean than in present. However, indirect constraints on paleo-velocities suggest that in Archean they were even slower than at present.

The analysis of ancient subduction (Hargraves, 1986) implies a greater ridge length (10–15 times of the present length) in Archean if heat loss at that time was three times greater than at present. This result suggests that the Archean Earth, prior to the aggregation of the Archean supercontinent and/or after its dispersal, could have been covered by many small plates moving with slow velocities. The conclusions of Hargraves (1986) on a greater ridge length and slower plate velocities in Archean are supported by the results of Abbott and Menke (1990), who estimated that the length of plate boundaries at 2.4 Ga was ~ 2.2 times greater than at present, and thus required plate velocities $\sim 16\%$ slower than at present to fit data on the higher mantle heat production in the Precambrian.

An indirect constraint on paleo-velocities follows from the results of England and Houseman (1984), who have modeled two-dimensional mantle convection with a constant-velocity upper boundary condition (simulating the presence of lithospheric plates) to explain the genesis of kimberlites, which have their source at a depth of 150–220 km. Such formulation of

the problem can appear dubious as one can question if plumes originating from the lower mantle “know” what is above them in the upper mantle. However, mantle convection models show that the incorporation of continents modifies the styles of mantle convection (e.g., Trompert and Hansen, 1998; Tackley, 2000),

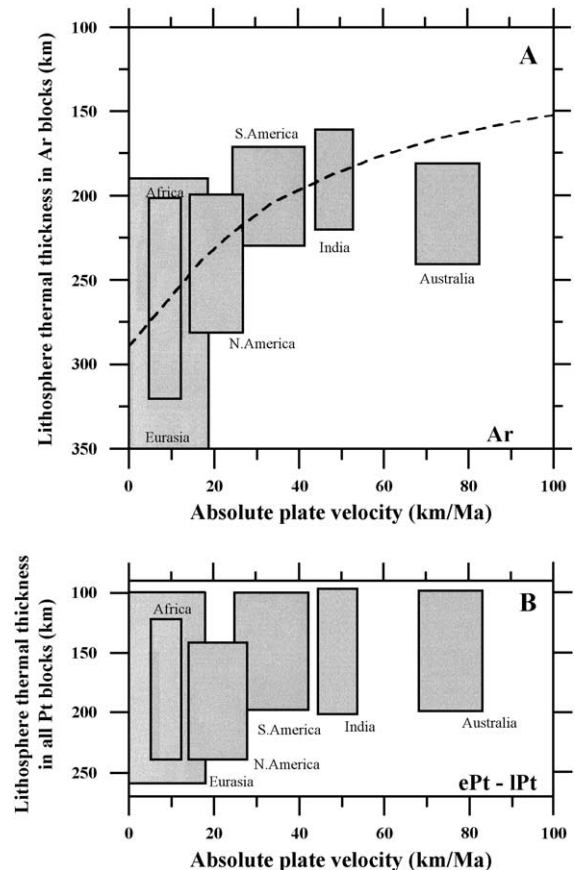


Fig. 8. Thickness of cratonic lithosphere versus absolute plate velocity. Boxes correspond to different cratons. The vertical dimensions of the boxes show range of lithospheric thickness and the horizontal dimension shows the error in plate velocity estimates. (A) The data for the Archean cratons supports the “basal drag” model, which predicts the lithospheric thickness to be inversely proportional to the square root of the plate velocity. The keels of the fast moving plates are smaller due to erosion by basal drag. Australia, which is located close to the subduction zone and thus has a high plate velocity, is the only plate to plot well off the curve. (B) Thickness of early–late Proterozoic lithosphere vs. plate velocity. The correlation between the length of the Proterozoic cratons and the keels is much weaker than for the Archean cratons. The plate velocities are for the hotspot reference frame (Gripp and Gordon, 1990).

indicating that the geometry of the continental lithospheric plates influences the mantle convection pattern. Davies (1999) suggests that the locations of convection upwellings and downwellings are influenced, or perhaps even controlled, by lithospheric structure rather than by properties of the deeper mantle. Moreover, recent results of Richards et al. (2000) show that changes in plate motions affect the dynamics not only of the upper mantle, but the lower mantle and D''-layer as well. England and Houseman

(1984) found that plates with slow velocities (less than 20 km/Ma) favor the formation of plumes beneath them, which can lead to partial melting of the lowermost lithosphere and generation of kimberlite magmas. Thus, the existence of kimberlites of many different ages in all of the cratons suggests that paleo-velocities of continental plates were often slow since Archean.

Fig. 8A shows a strong inverse correlation between absolute plate velocity and thickness of Archean lithosphere ($r = -0.77$ including Australia and $r = -0.97$ without Australia, Table 3). Plates having deep Archean keels move slow. North America, Africa, and Eurasia, all of which have thick Archean keels (300 km and more), move with a plate velocity less than 21 mm/year (Table 2). Australia is the only exception to the observed trend and has a high plate velocity (75 mm/year) due to a strong subduction pull (Forsyth and Uyeda, 1975; Lithgow-Bertelloni and Richards, 1998). Thus, basal drag plays a minor role in the motion of the Australian plate, still being an important mechanism in the basal erosion of Australian lithosphere. We note that, in order for the Australian plate to meet the best-fit curve in Fig. 8A, the basal 50 km of the Australian lithosphere would need to be eroded. We speculate that such a scenario may be possible, with the high velocity of the Australia plate resulting in a high degree of basal drag (and lithosphere erosion to 170 km).

The curve in Fig. 8A flattens for Archean lithospheric thickness less than ~200 km, indicating a weak correlation between plate velocity and lithospheric thickness. This suggests that the depth of

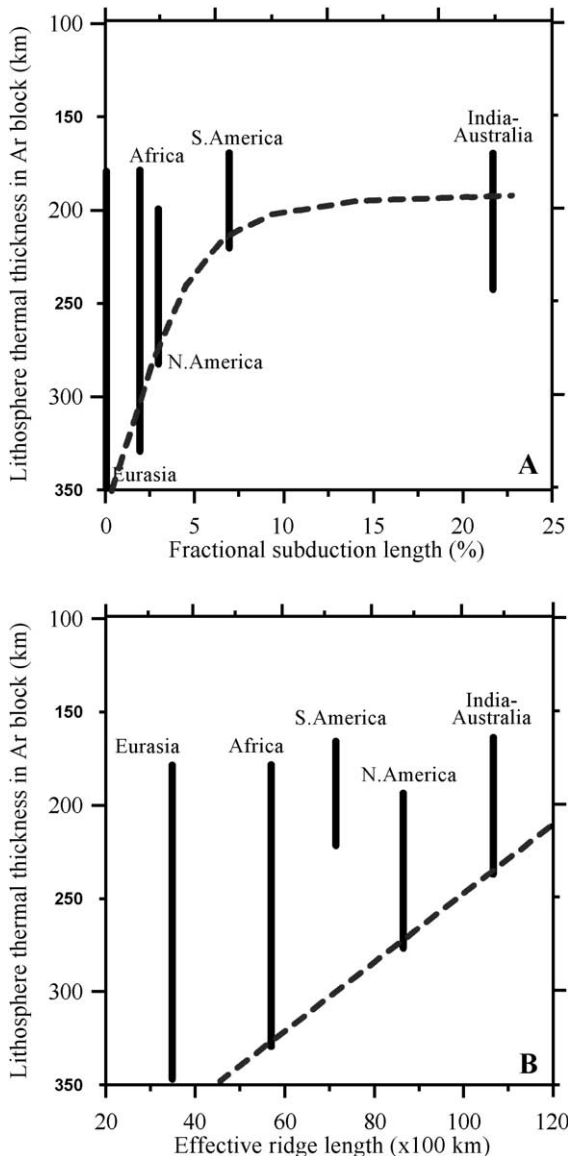


Fig. 9. Thickness of cratonic lithosphere versus “slab pull” and “ridge push”. The correlations between the parameters are especially strong for the maximal values of the lithospheric thickness in Archean blocks (dashed lines). No correlations exist for the middle-late Proterozoic lithosphere as it has a uniform thickness. (A) Lithospheric thermal thickness versus fractional subduction length (i.e. the percent of a plate perimeter taken by subduction zones). (B) Lithospheric thermal thickness versus effective ridge length (the ridge length exerting a net force, i.e. the force which is not cancelled, for example, by a ridge on the opposite side of a plate). The values of subduction length and ridge length are from Table 1 of Forsyth and Uyeda (1975), vertical bars give the range of the lithosphere thermal thickness estimated for the Archean–early Proterozoic blocks of the cratons (Artemieva and Mooney, 2001).

~ 200 km can be the base of the low-viscosity asthenospheric layer. Middle and late Proterozoic lithosphere is in general thinner than 180 km (Artemieva and Mooney, 2001) and, similarly to Figs. 4B and 7B, a plot of plate velocity versus lithospheric thickness for Proterozoic lithosphere shows no correlation (Fig. 8B and Table 3). This accords with Stoddard and Abbott (1996) who found that thick Archean keels encounter basal drag in the sub-asthenospheric mantle, whereas thin Proterozoic

keels encounter minimal drag within the asthenospheric low-velocity layer.

We further expand the analysis of Forsyth and Uyeda (1975) and examine the relative role of subduction pull and ridge push on plate motions. We compare lithospheric thickness with (1) fractional subduction length (i.e., the connectivity of a plate to a slab expressed as percentage of the subduction length versus the total plate boundary) and (2) with the effective ridge length (which is the ridge length exerting a net force, i.e., the force not cancelled, for example, by a ridge on the opposite side of a plate) (Fig. 9A and B, correspondingly). The values of both of the parameters are derived from Table 1 of Forsyth and Uyeda (1975), and thus in this part of our analysis we follow their assumption that India and Australia belong to the same plate.

Since absolute plate velocity is proportional to the amount of subduction pull (Forsyth and Uyeda, 1975) and increases with a decrease of lithospheric thickness (Fig. 8A), we expect an inverse correlation to exist between lithospheric thickness and subduction pull. While such a trend may be present in Fig. 9A, an inadequate distribution of subduction length values prevents definite conclusions, beyond general consistency with respect to Fig. 8A. Likewise, a

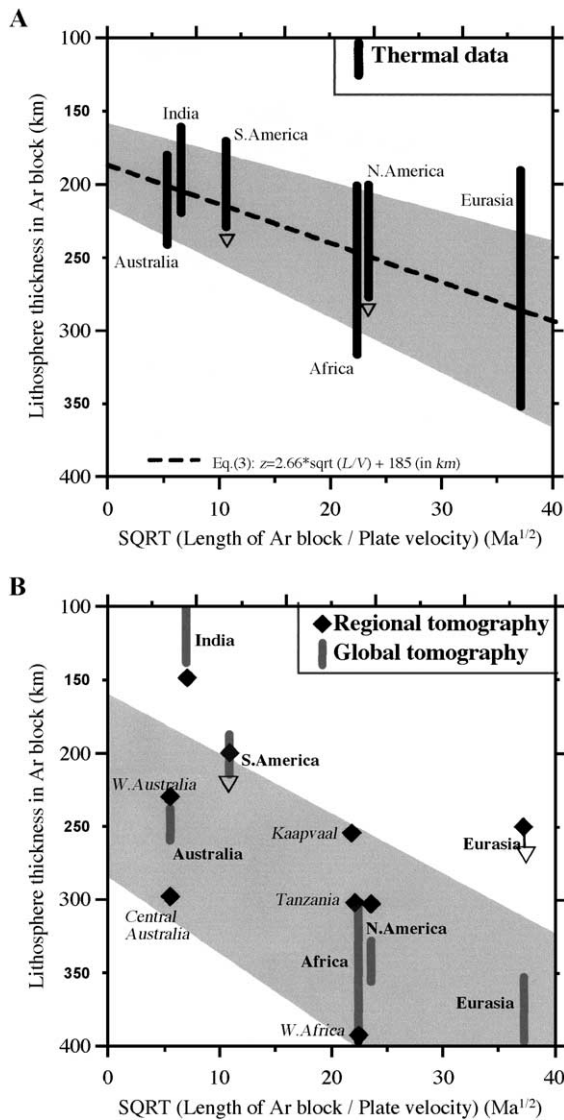


Fig. 10. Lithospheric erosion by basal drag. This model predicts that the thickness of eroded lithosphere is proportional to the square root of the time needed for the plate to move the width of the craton (i.e. the square root of the ratio of the Archean craton length to the plate velocity, see text). Lithospheric thickness of Archean cratons shows a linear correlation with the basal drag relation (Eq. (3)). (A) The relationship based on thermal estimates of lithospheric thickness (Artemieva and Mooney, 2001). The range of lithospheric thickness for South America and North America may be artificially reduced by the absence of data where the cratons are expected to have their maximum lithospheric thickness (shown by arrows). Dashed line gives the best-fit for the mean values of the lithospheric thickness (Table 2). (B) The relationship based on seismic estimates of lithospheric thickness. Diamonds—regional seismic tomography (for +1.0% velocity anomaly); gray lines—global seismic tomography (for +0.5% velocity anomaly). For references see Fig. 1. Regional tomography results published for South America and Eurasia did not reveal the base of the high-velocity lithospheric keel (shown by arrows). In agreement with Jaupart et al. (1998), the trend for seismic lithospheric thickness, which corresponds to the top of the convecting mantle, lies ~ 50 km deeper than for thermal thickness (shaded areas).

comparison of lithospheric thickness with effective ridge push (Fig. 9B) provides evidence for an inverse correlation between these two parameters. The correlation is especially strong not for the mean, but for the maximum values of lithospheric thickness in Archean cratons (dashed lines in Fig. 9).

In conclusion, we find that there is an inverse correlation between lithospheric thickness in Archean cratons and the two main driving forces of plate motion. This implies that while subduction pull and ridge push determine velocities of lithospheric plate, the lithosphere itself is eroded by mantle drag which is proportional to the plate velocity.

3.4. Further support for the basal drag model

In agreement with the predictions of the basal drag model of Sleep (submitted for publication) (Eq. (3)), a plot of lithospheric thermal thickness versus the square root of Archean cratonic length divided by plate velocity shows a positive correlation (Tables 2 and 3) (Fig. 10A). We conclude that the basal drag model provides one viable explanation for the variation in thickness of Archean cratonic roots. Cratons with low plate velocities and large size have better chances to preserve deep keels than small fast-moving cratons, where the basal erosion due to the relative movement between the plate and the mantle will gradually remove the lowermost parts of the lithosphere. The predictions of the basal drag model do not hold for Proterozoic cratons, for which no correlation between the square root of cratonic length divided by plate velocity exists.

The best-fit line in Fig. 10 suggests that removal of the lowermost part of the Archean lithosphere by basal drag results in a limiting value of lithospheric thickness of 185 ± 20 km. This value also corresponds to the maximum thickness of middle–late Proterozoic lithosphere (Fig. 2) and supports geological evidence that Proterozoic lithosphere is often reworked Archean lithosphere (e.g., Goodwin, 1996). The depth of 185 km also corresponds to the top of the diamond stability field in the continental lithosphere (Kennedy and Kennedy, 1976). The absence of diamonds in the Proterozoic lithospheric keels can be explained by the removal of the lowermost part of the reworked Archean lithosphere by basal drag.

For a comparison we analyzed the predictions of the basal drag model for the thickness of seismic lithosphere as defined from global and regional seismic tomography studies (Fig. 10B). Except for the Indian craton, which is small and thus not well resolved in tomography, all other cratons follow the trend similar to the one in Fig. 10A, but shifted to a greater depth because of a different sampling of the lithospheric base by thermal and seismic data (Fig. 3). Thus, data from seismic tomography studies also confirms the mantle drag model of the lithospheric basal erosion.

However, the erosion of the cratonic lithosphere by mantle convection and plumes is to a large degree a spatially random process, and does not depend only on the geometry of the cratons or plate motion. Therefore, the possible effects of small-scale convection and plume–lithosphere interaction are difficult to separate from the effect of basal drag. Nevertheless, the predicted magnitude of the effect of basal drag is clearly evident in Fig. 10, where the observed lithospheric thickness fits the predicted relationship (Eq. (3)). We conclude that stirring at the base of the cratonic lithosphere caused by the relative movement of the lithospheric keel and the underlying mantle has played an important role in thinning of the cratonic lithosphere since the early Archean.

4. Concluding remarks: plate motion, basal erosion, and preservation of cratonic lithosphere

4.1. Lithospheric thickness

We summarize seismic and thermal data that show that the thickness of Archean cratonic lithosphere ranges from 140 to ~ 350 km. As petrological data contradict tomographic and thermal models, we propose that the lithospheric base can have large thickness variations over short distances and different techniques sample different parts of it. The topography of Bryce Canyon (western USA) may be an inverted analog of such an eroded structure.

4.2. Vertical versus lateral dimensions of the cratons

We use estimates of lithosphere thermal thickness (Artemieva and Mooney, 2001) to examine the corre-

lations between cratonic size, plate velocities, and lithospheric thickness. The results suggest that the horizontal and vertical dimensions of the Archean cratons are well correlated: larger Archean cratons have thicker lithosphere. However, this correlation does not hold for middle–late Proterozoic lithosphere. The extrapolation of the linear trend for cratonic size versus lithospheric thickness to the total size of all present-day Archean cratons (hypothesized to have formed an early Archean supercontinent) suggests that the ancient (~ 4.0 Ga) lithosphere may have been ~ 450 -km thick.

4.3. Interaction between plate motion and lithospheric keels

We address the question of how early Archean lithosphere that may have been ~ 450 -km thick was eroded to its present-day bimodal thickness of 200–220 and 300–350 km (Doin et al., 1997; Artemieva and Mooney, 2001). One would also expect that such deep lithospheric keels could influence plate motions (e.g., Chapman and Pollack, 1974). This idea is supported by the results of Stoddard and Abbott (1996) who showed that the Proterozoic part of the keels, located within the low-viscosity asthenosphere at the depth around 120–180 km, has a weak effect on plate movement, while deeper Archean keels resist plate motion.

Conversely, plate motion also strongly influences the thickness of the Archean keels. We propose that the interaction between plate motion and keel thickness is double-sided: for plates not attached to subducting slabs, thick Archean keels can slow plate velocities, while plate motion erodes the keels by basal drag. The first detailed analysis of the driving forces for plate motions (Forsyth and Uyeda, 1975) have shown the importance of mantle drag, a conclusion that is supported by the inverse correlation between the area of Archean cratons and plate velocity (Stoddard and Abbott, 1996).

4.4. Basal drag model: correlation of plate velocities with lithospheric thickness and cratonic area

We evaluate the basal drag model, whereby the lithosphere is eroded due to its relative movement with respect to the underlying mantle (Sleep, submit-

ted for publication), and find that for Archean cratons this model is in excellent agreement with observed data: lithospheric thickness is proportional to the square root of the ratio of the craton length (along the direction of plate motion) to the plate velocity. Large cratons with thick keels and low plate velocities (e.g., Eurasia and North America) are less eroded by basal drag than fast-moving small cratons (e.g., India and Australia). This means that earlier studies (Forsyth and Uyeda, 1975; Stoddard and Abbott, 1996) have addressed only one aspect of a more complicated relationship, whereby both craton size and plate velocity correlate with the lithospheric thickness: large Archean cratons tend to have thick lithospheric keels and very slow plate velocities. We emphasize that these correlations hold only for the Archean cratons, not for middle–late Proterozoic cratons; for early Proterozoic cratons the correlation is very weak.

4.5. Relations between lithospheric thickness, slab pull, and ridge push

We further address the question how lithospheric thickness correlates with two major driving forces of plate motion. The results show an inverse correlation between lithospheric thickness and: (a) fractional subduction length; and (b) the effective ridge length. These results indicate that lithosphere erosion by mantle drag is proportional to the plate velocity.

4.6. Preservation of the lithospheric keels

Our results suggest that the slower the plate moves the weaker is the erosion of the keel. This implies that thick Archean keels can be preserved for a long time (i.e., 3–4 Ga). Cratons with large sizes are also more stable with respect to basal erosion by mantle convection due to an efficient deflection of heat from the deep mantle (Ballard and Pollack, 1987; Lenardic and Moresi, 2001). Their stability is further maintained by the depleted (e.g., Jordan, 1988) and dry (e.g., Pollack, 1986) composition of the Archean lithosphere. However, very thick (~ 400 km) lithospheric keels could have survived until present probably only locally, an observation supported by the fact that Archean

cratonic lithosphere thickness rarely exceeds 300–350 km.

4.7. Reworking of Archean keels in Proterozoic

When erosion due to secondary convection at the margins of a thick (~ 350 km), large craton reduces its lateral dimension to a critical value of $\sim 6\text{--}8 \times 10^6$ km², the keel fails to divert the basal heat efficiently. In this case, the Archean keel is thinned by mantle convection to an equilibrium thickness of ~ 220 km. Since all Proterozoic cratons have lithospheric keels that are less than 200 km thick, we infer that only thinned (~ 220 km) Archean lithosphere is a candidate for the geologically known reworking into Proterozoic lithosphere.

Due to the viscosity–depth structure of the upper mantle, thinning of the Archean lithospheric keel will reduce basal drag and therefore resistance to plate motion. This will permit faster movement of the craton with respect to the underlying mantle, which, in turn, will enhance lithosphere erosion by the basal drag. If an Archean keel is thinned to significantly less than 200 km, the remaining lithospheric column will be a candidate for strong deformation in a collisional environment and for modification by metasomatism (i.e., invasion by volatiles and relatively enriched mantle magmas).

4.8. Secular cooling of the mantle and mantle drag

We recognize that the basal drag model suggests a long-term preservation of thick (300–350 km) Archean keels only if the cratons have never experienced a period of high plate velocity. In view of the long (~ 4 Ga) existence of the keels, this scenario seems unlikely. Though it has been suggested that plate velocities in Archean could have been even slower than at present, it is difficult to imagine that the Archean cratons of West Africa, Baltica, and Siberia were never a part of a fast-moving plate. However, if the viscosity of the mantle to a depth of ~ 450 km was one or two orders of magnitude lower during the Archean, corresponding to mantle temperature some 100–200 °C higher than today, the basal drag, and along with it, basal erosion would have been much smaller than today. Thus, the thick keels of Archean cratons would have been

preserved, even if Archean plate velocities were high.

Acknowledgements

We thank Norm Sleep, Brad Hager, and Rick O'Connell for insightful discussions on these general topics. Reviews and constructive comments by Dallas Abbott and Tom Parsons are appreciated. The comments of Hans Thybo, Fred Pollitz, and James Cahalane helped to improve the clarity of the presentation.

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