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The long-term strength of Europe and its implications for plate-forming processes

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Field-based geological studies show that continental deformation preferentially occurs in young tectonic provinces rather than in old cratons[1]. This partitioning of deformation suggests that the cratons are stronger than surrounding younger Phanerozoic provinces. However, although Archaean and Phanerozoic lithosphere differ in their thickness[2–4] and composition[4,5], their relative strength is a matter of much debate. One proxy of strength is the effective elastic thickness of the lithosphere, T_e. Unfortunately, spatial variations in T_e are not well understood, as different methods yield different results. The differences are most apparent in cratons, where the 'Bouguer coherence' method yields large T_e values (>60 km)[6–9] whereas the 'free-air admittance' method yields low values (<25 km)[10]. Here we present estimates of the variability of T_e in Europe using both methods. We show that when they are consistently formulated[11], both methods yield comparable T_e values that correlate with geology, and that the strength of old lithosphere (≥1.5 Gyr old) is much larger (mean T_e > 60 km) than that of younger lithosphere (mean T_e < 30 km). We propose that this strength difference reflects changes in lithospheric plate structure (thickness, geothermal gradient and composition) that result from mantle temperature and volatile content decrease through Earth's history.

The current debate concerning the strength of continental lithosphere is focused on the values of T_e estimated using both forward and inverse (spectral) methods based on the Bouguer coherence and free-air admittance. For example, forward models in the Slave craton (Canadian shield) reveal a T_e of about 12 km (ref. 12), which is much smaller than the ~100 km obtained using Bouguer coherence in the same area[6]. This strength difference arises because, in cratons, forward models and Bouguer coherence yield estimates of the strength at different times[12]. Forward models are generally based on the reconstruction of the original (surface and subsurface) loads and their associated flexures and reveal the strength at the time of a specific loading event (for example, orogeny or rifting), while Bouguer coherence is based on present-day topography and gravity anomaly data and yields the current strength of thick, cooled, cratonic lithosphere[12]. Because the free-air admittance is also based on present-day topography and gravity anomaly data, it should yield a T_e similar to that obtained using the Bouguer coherence. However, in some regions, it yields $T_e < 25$ km (ref. 10), while the Bouguer coherence yields values in excess of 60 km (refs 6–9).

Recently, it has been shown that this discrepancy occurs because the free-air admittance and Bouguer coherence methods have not been consistently formulated[11]. The calculation of both functions involves the estimation of the spectra of finite and non-periodic data. Therefore, the wavelength dependence of the admittance and coherence varies with the size of the data window analysed. In the admittance method, analytical solutions of the predicted spectra (which correspond to infinite data windows) have been compared to observed spectra of the data within a finite window[11]. When the underlying T_e is large (>30–40 km), this can lead to an underestimation of T_e by a factor of 2 (ref. 11). However, when the observed and predicted admittance functions are calculated in the same data windows, as is usually the case in the coherence method, then the results from the two techniques are equivalent[11].

We present here estimates of the T_e structure of Europe using both Bouguer coherence and free-air admittance. The resulting T_e structures are similar, and correlate well with the tectonic provinces in Europe inferred from geological and other geophysical data[13] (Fig. 1). High-T_e regions correlate with Precambrian Baltica and Avalonia. Low-T_e regions, in contrast, correlate with the younger provinces accreted during the Caledonian, Variscan and Alpine orogenies. These results are in agreement with local studies of T_e in Europe[14–17]. Figure 1 also shows that the largest changes in T_e occur at the sutures that separate different provinces. Deep seismic data indicate that these sutures coincide with major tectonic boundaries in the crust and shallow lithospheric mantle (for example, Iapetus suture[18], Thor suture[19]). Our results show that, in addition, these sutures correlate with major changes in lithospheric strength.

The recovered T_e is generally consistent with other physical properties of the lithosphere. High-T_e regions generally correlate with areas of large thermal thickness (as derived from heat flow data[3]) and fast seismic S-wave velocities[20] and vice versa (Fig. 1c, d). This indicates that T_e is high in regions where the lithosphere is thick and cold. In addition, a close correlation exists between T_e and seismicity: high- and low-T_e regions show sparse and abundant seismicity, respectively (Fig. 1d). This suggests that in high-T_e areas, the strength of the lithosphere is large enough to reduce the background level of seismicity[21].

The dependence of T_e on age is shown in Fig. 2. The figure shows a plot of the mean T_e obtained from Bouguer coherence for each of the main tectonic provinces versus their age (Table 1). Unlike T_e values in oceanic lithosphere, spectrally derived continental T_e estimates do not necessarily reflect the strength at the time of loading. This is because spectral estimates reflect the present-day response to loads, which resulted from specific geological events (for example, orogeny and rifting) as well as from subsequent sedimentation and erosion. The relatively young Caledonian, Variscan and Alpine provinces still have significant topography such that their spectrally derived T_e also mainly reflects the strength of the basement at the time of these orogenies (that is, it is the T_e at the time of loading which is recovered). However, cratons have subdued topography, and so their spectrally derived T_e mainly reflects the response to post-orogenic erosion and sedimentation, both of which may continue up to the present day. Because forward models reflect the T_e at the time of loading[12], they generally coincide with estimates of spectrally derived T_e in young tectonic provinces (see, for example, ref. 17) but differ in cratonic areas[12].

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Figure 2 shows that Archaean and Early/Middle Proterozoic tectonic provinces (≥1.5 Gyr old) are much stronger than younger ones. This strength increase with tectonic province age cannot be explained by conductive cooling of a lithospheric plate with a given thermal thickness as in oceanic lithosphere. For example, both the Sveconorwegian (~1 Gyr old) and Karelian (~3 Gyr old) tectonic provinces have had sufficient time (>1 Gyr) to conductively cool and thermally equilibrate, yet T_e in the former is at least half that of the latter (Fig. 2). Therefore, there must be fundamental differences in the mechanical structure between old and young continental lithosphere in Europe.

We suggest that these structural differences may reflect changes in continental plate forming processes related to the decrease in temperature and volatile content in the sublithospheric mantle during Earth's history. In the Archaean, higher mantle temperatures and/or volatile content probably favoured a larger degree of melting to greater depths than today and, hence, the formation of a thick lithosphere, with a highly depleted, buoyant root[4,22]. At that time, the

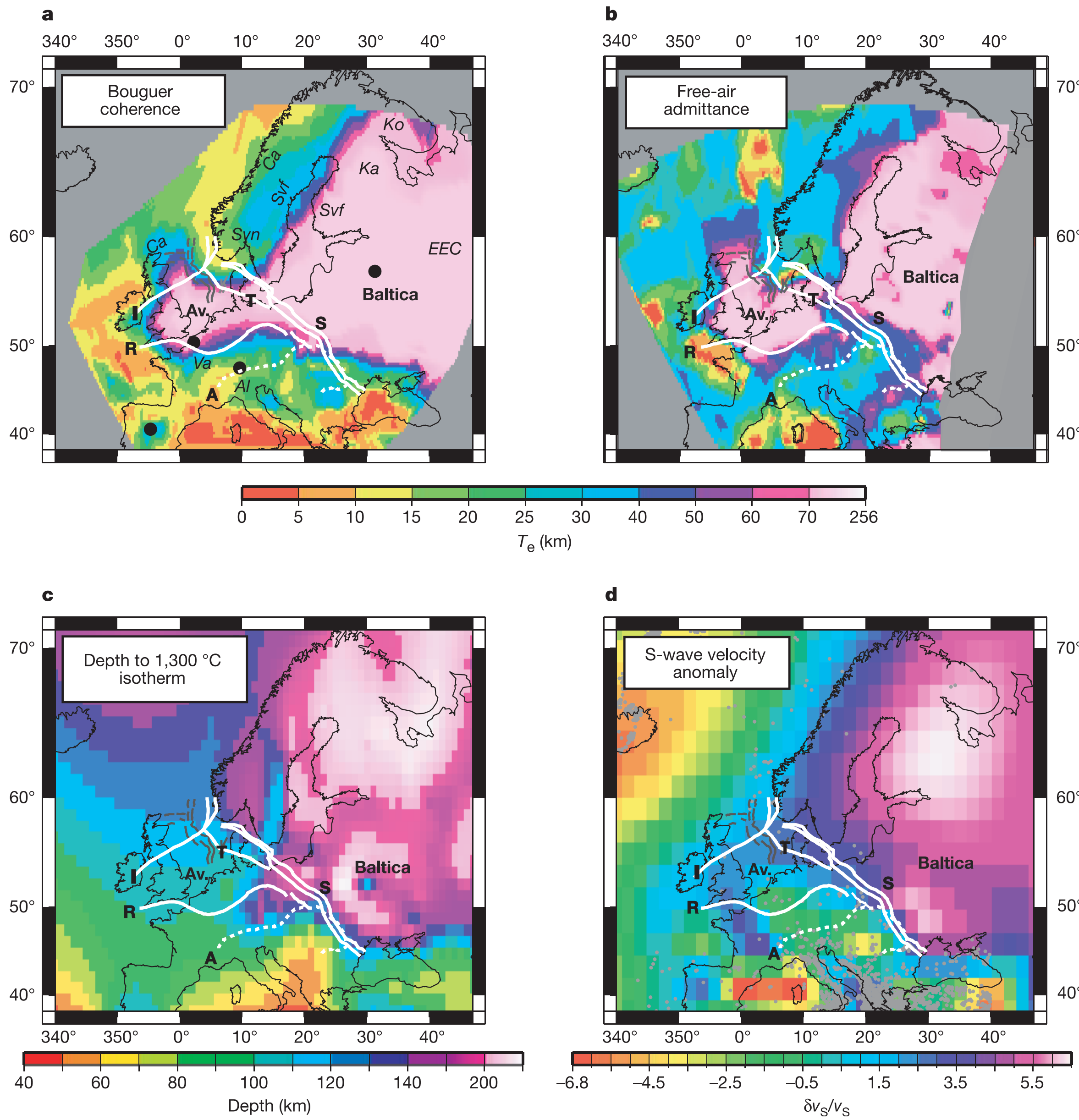

Figure 1 | T_e structure of Europe obtained using two different methods, and comparison with other geophysical data. a, **b**, T_e obtained from Bouguer coherence (**a**) and free-air admittance (**b**). **c**, **d**, Other geophysical data: **c**, thermal thickness[3] (defined as depth to the 1,300 °C isotherm); and **d**, S-wave velocity anomaly, $\delta v_S/v_S$, at 100 km depth[20] given in per cent with respect to an isotropic version of PREM (decribed in ref. 20). Note that even if T_e values do not exactly coincide in **a** and **b**, the general pattern of T_e variation is equivalent. Free-air admittance has a poorer T_e recovery ability than the Bouguer coherence (Methods); we consider the results obtained with the latter more reliable. The T_e structure in **a** and **b** is based on CRUST 2.0[30], Poisson ratio = 0.25, and Young's modulus = 100 GPa. Because the maximum T_e that can be recovered with confidence is 60 km, larger T_e estimates exceed this value by an undetermined amount (Methods). Hence the relative strength between the Precambrian Baltica and Avalonia (Av) can not be resolved. White lines define the sutures: I, Iapetus; T, Thor; R, Rheic; and S, Sorgenfrei-Tornquist and Teisseyre-Tornquist zones, between the Precambrian provinces and younger ones. Italic labels show the approximate location of the tectonic provinces: Ko, Kola; Ka, Karelia; Svf, Svecofennian; EEC, East European Platform; Svn, Sveconorwegian; Ca, Caledonian; Va, Variscan; and Al, Alpine. Dashed white line (A) shows the Alpine deformation front. Grey lines in indicate the main boundary faults of the North Sea rift, which borders the high-T_e area in Avalonia. Grey dots in **d** are earthquake locations (US Geological Survey Earthquake Data Base, $M_b > 4.0$). Black dots in **a** show locations of Bouguer coherence analysis shown in Supplementary Information.

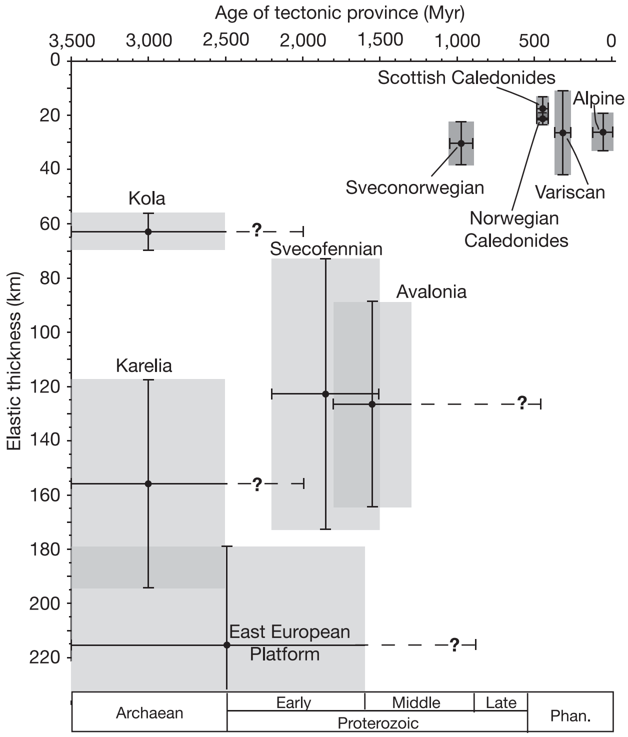

Figure 2 | Mean T_e determined from Bouguer coherence within each of the major tectonic provinces of Europe versus their age. Filled circles show data points; age is taken from Table 1. The age of the tectonic province corresponds to the age of those portions of the province that have not been deformed by later orogenic events at their edges. For example, the East European Platform consists mainly of Archaean and Early Proterozoic basement that are largely undeformed, so we assign this age instead of that of the later orogenies that have modified its edges. For tectonic provinces comprising only orogenic belts such as the Alps, we assign the age of the orogeny (see Table 1). To distinguish between tectonic province and orogenic age, we have coloured them in light and dark grey respectively. Error bars indicate the standard deviation in T_e and the age span of the tectonic provinces. Dashed error bars and question marks indicate uncertainty in the age of the tectonic province (see Table 1). As mentioned in Fig. 1, the maximum T_e estimate that we can recover is 60 km (Methods). Hence, variations in strength between tectonic provinces with $T_e > 60$ km cannot be interpreted.

response of the lithosphere to loading would have reflected a combination of weakening due to the high geothermal gradients and any strengthening due to compositional effects (related to melt extraction[22]). These two competing factors might have resulted in an Archaean lithosphere, which despite its thickness, had a low strength and hence low T_e. However, with time, conductive cooling of the lithosphere to a stable state, as well as the Earth's secular cooling itself, would have increased its strength. This probably explains the high present-day T_e values that have been determined (both in this Letter and elsewhere) using spectral methods in old tectonic provinces (for example, North America[6–8], Australia[9], South America[23] and Africa[24]).

The decrease in temperature and volatile content in the sublithospheric mantle through time[22] has resulted in Late Proterozoic and Phanerozoic lithosphere that is associated with smaller degrees of melting at shallower depths[5,22]. Hence, this lithosphere is thinner[5], has a higher geothermal gradient[5], and is less depleted in basaltic constituents than Archaean lithosphere[5]. This makes Late Proterozoic and Phanerozoic lithosphere intrinsically weaker than older lithosphere, even after conductive cooling. This weakness is reflected, we believe, in the low T_e values obtained here in young tectonic provinces using spectral methods (Fig. 2).

It is interesting to note that the youngest ages for lithosphere

Table 1 | Age of European tectonic provinces

Tectonic province	Age range (Gyr)
Kola*	3.5–2.5
Karelia*	3.5–2.5
Svecofennian†	2.2–1.5
East European Platform‡	3.5–1.6
Avalonia§	1.3–1.8
Sveconorwegian	1.05–0.9
Caledonian	0.488–0.416
Variscan	0.359–0.27
Alpine	0.12–0

*The Kola and Karelian tectonic provinces comprise Archaean basement with a U–Pb age in the range 2.5–3.5 Gyr (ref. 13) and were amalgamated 2 Gyr ago. Hence, we assign them an age span of 3.5–2.5 Gyr. In Fig. 2 we extend the age error bar to 2 Gyr, as the extent to which the amalgamation 2 Gyr ago modified the Kola and Karelia lithospheres is uncertain.
†The Svecofennian tectonic province contains material younger than 2.2 Gyr and was accreted and underwent collision 2.0–1.8 Gyr ago. It was later locally reworked by melting 1.8–1.5 Gyr ago[13]. We assign an age range of 2.2–1.5 Gyr to this tectonic province.
‡The East European Platform consists mainly of basement of Archaean and Early Proterozoic age and locally middle Proterozoic age[31]. We assign an age range of 3.5–1.6 Gyr. In Fig. 2 we extend the age error bar to 0.9 Gyr, with a question mark to indicate that the extent to which the local middle Proterozoic basement is represented in the T_e of this province is uncertain.
§The Avalonia basement is structurally very heterogeneous and rarely exposed. Therefore, we have based our age range on Sm–Nd model ages, which vary from 1.3 to 1.8 Gyr (ref. 32). Note that in Fig. 2 we have extended the age error bar up to 443 Myr ago (when Avalonia docked against Baltica[13]), with a question mark to indicate the uncertainty about the extent to which later tectonic events affecting Avalonia's edges modified its lithospheric core.

with seismic and thermal thicknesses greater than 200 km are around 1.6–1.7 Gyr (refs 2, 3), a similar age to that at which T_e, in this study, appears to change markedly, except, perhaps, for Avalonia (Fig. 2). However, given that Avalonia is structurally very heterogeneous, its highest T_e values may reflect the areas where the basement ages are oldest (see Table 1).

Although it is not yet known exactly when cratons acquired their strength and stability[12,25], it is likely that, at least, they were strong throughout the Phanerozoic. If so, continental deformation would preferentially occur at the edges of stable cratonic provinces. Figure 1 shows some weakening of Baltica, for example, where it is juxtaposed to the Caledonian, Variscan and Alpine orogenic belts. We follow Dixon *et al.*[26] and speculate that water introduced during subduction may have further weakened the Phanerozoic tectonic provinces. The net result is a tectonic province that because of its weakness acts as a focus, such that sites of orogeny become repeatedly involved in both compressional and extensional deformation, as is predicted in the Wilson cycle.

Finally, our results indicate that the T_e of old tectonic provinces ($\geq$1.5 Gyr) is significantly larger (>60 km) than their mean crustal thickness (~40 km)[27]. This suggests that the lithospheric mantle is strong, consistent with dynamical models that indicate that the stability of cratons is due not only to the chemical buoyancy of their root but also, importantly, to root strength[28,29]. It appears that only when the cratonic lithosphere is subjected to processes such as enrichment by hot upwelling mantle will the old cratons be weakened enough to deform[5].

METHODS

Calculations and data. The Bouguer coherence and free-air admittance are statistical methods that determine the relationship between the Bouguer and free-air gravity anomaly and the topography as a function of wavelength[7]. Both methods consist in finding a 'best fit' T_e by minimizing the root-mean-square difference between observed and predicted coherence and admittance functions[7].

We base our analysis on a continent-wide 8×8 km grid of gravity anomaly (Bouguer onshore and free-air offshore) and topography data. These were compiled by GETECH (UK) as part of their West-East Europe Gravity Project (WEEGP). The gravity anomalies have been corrected for terrain, and we estimate that the data are accurate to better than 1–2 mGal. The calculation of the Bouguer anomaly offshore and the free-air anomaly onshore is described elsewhere[11].

The calculation of the Bouguer coherence and free-air admittance follows refs 7 and 11. The only differences from the methods followed in ref. 11 is that we:

(1) deconvolve the loads in the same data window as the observed functions, (2) assume that subsurface loads occur at the boundary of the upper and middle crust, and (3) combine the T_e results obtained using windows of different sizes. The model input parameters are summarized in Fig. 1.

The load deconvolution consists of extracting the contribution to the observed topography and gravity anomaly of the surface and subsurface loads that were initially emplaced on the lithosphere[7,11]. The deconvolution requires information on the density structure of the crust, which we deduced from CRUST 2.0[30].

The subsurface loads occur at the upper/middle crust boundary, which, according to CRUST 2.0[30], is at ~10–15 km depth. Because the predicted gravity anomaly due to subsurface loading increases with T_e and decreasing depth of loading, any given gravity anomaly can be modelled by a combination of either a small loading depth and small T_e or a larger depth and larger T_e. However, we have found that the T_e variation resulting from different assumed loading depths is not large (±5 km).

To recover a spatially varying T_e, the window size used needs to be large enough to recover the maximum flexural wavelength, but small enough to recover the spatial variation in T_e (Supplementary Fig. 3 illustrates how T_e varies with window size). To obtain an optimal solution, we used overlapping windows spaced 56 km apart and assigned the resulting T_e to the window centre. The study region was analysed four times using window sizes of 400 × 400 km, 600 × 600 km, 800 × 800 km and 1,000 × 1,000 km. We therefore generated our 'final' T_e structure using a combination of different window sizes. In particular, for $T_e < 20$ km, $20 < T_e < 40$ km, $40 < T_e < 60$ km and $T_e > 60$ km, we used the results obtained with windows of 400 × 400 km, 600 × 600 km, 800 × 800 km and 1,000 × 1,000 km, respectively.

Limitations. On the basis of tests with synthetic topography and gravity anomaly data, it has been found that, for a window of 1,000 × 1,000 km, the largest T_e that can be recovered with confidence is 60 km (ref. 11; see also Supplementary Information). Hence, the amount by which any particular T_e estimate exceeds this value is undetermined. In addition, in small areas where T_e is high (for example, central Finland), the use of large windows results in a reduction of T_e due to the inclusion in the analysis of low-T_e areas that flank the high-T_e areas. We therefore only place a lower limit on the largest T_e obtained (60 km), and attach little significance to local spatial variations of T_e within the Baltica and Avalonia tectonic provinces.

The performance of the free-air admittance in recovering T_e values is poorer than that of the Bouguer coherence owing to the relatively low power of the free-air anomaly at wavelengths where the isostatic compensation occurs (see, for example, ref. 11). Hence, we consider the results obtained with Bouguer coherence to be more reliable.

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1. McConnell, R. B. Geological development of the rift system of Eastern Africa. *Geol. Soc. Am. Bull.* **83,** 2549–2572 (1972).
2. Polet, J. & Anderson, D. L. Depth extent of cratons as inferred from tomographic studies. *Geology* **23,** 205–208 (1995).
3. Artemeieva, I. M. & Mooney, W. D. Thermal evolution of Precambrian lithosphere: A global study. *J. Geophys. Res.* **106,** 16387–16414 (2001).
4. Jordan, T. H. Composition and development of the continental lithosphere. *Nature* **274,** 544–548 (1978).
5. O'Reilly, S. Y., Griffin, W. L., Poudjom Djomani, Y. H. & Morgan, P. Are lithospheres forever? Tracking changes in subcontinental mantle through time. *GSA Today* **11**(4–10) (2001).
6. Bechtel, T. D., Forsyth, D. W., Sharpton, V. L. & Grieve, R. A. F. Variations in the effective elastic thickness of the North American lithosphere. *Nature* **343,** 636–638 (1989).
7. Forsyth, D. Subsurface loading and estimates of the flexural rigidity of continental lithosphere. *J. Geophys. Res.* **90,** 12623–12632 (1985).
8. Audet, P. & Mareschal, J.-C. Variations in elastic thickness in the Canadian Shield. *Earth Planet. Sci. Lett.* **226,** 17–31 (2004).
9. Swain, C. J. & Kirby, J. F. The effect of 'noise' on estimates of the elastic thickness of the continental lithosphere by the coherence method. *Geophys. Res. Lett.* **30,** 1574–1578, doi: 10.1029/2003GL017070 (2003).
10. McKenzie, D. Estimating T_e in the presence of internal loads. *J. Geophys. Res.* **108,** doi:10.1029/2002JB001766 (2003).
11. Pérez-Gussinyé, M., Lowry, A., Watts, A. B. & Velicogna, I. On the recovery of the effective elastic thickness using spectral methods: examples from synthetic data and the Fennoscandian shield. *J. Geophys. Res.* **109,** doi: 10.1029/2003JB002788 (2004).
12. Grotzinger, J. & Royden, L. Elastic strength of the Slave craton at 1.9 Gyr and implications for the thermal evolution of the continents. *Nature* **347,** 64–66 (1990).
13. Windley, B. in *A Continent Revealed. The European Geotraverse* (eds Blundell, D., Freeman, R. & Mueller, S.) 139–214 (Cambridge Univ. Press, Cambridge, UK, 1992).
14. Kogan, M. G., Fairhead, J. D., Balmino, G. & Makedonskii, E. L. Tectonic fabric and lithospheric strength of northern Eurasia based on gravity data. *Geophys. Res. Lett.* **21,** 2653–2656 (1994).
15. Poudjom Djomani, Y. H., Fairhead, J. D. & Griffin, W. L. The flexural rigidity of Fennoscandia: reflection of the tectonothermal age of the lithospheric mantle. *Earth Planet. Sci. Lett.* **174,** 139–154 (1999).
16. Cloetingh, S. *et al.* Lithospheric memory, state of stress and rheology: neotectonic controls on Europe's intraplate continental topography. *Quat. Sci. Rev.* **24,** 241–305 (2005).
17. Stewart, J. & Watts, A. B. Gravity anomalies and spatial variations of flexural rigidity at mountain ranges. *J. Geophys. Res.* **102,** 5327–5352 (1997).
18. McBride, J. H., Snyder, D. B., England, R. W. & Hobbs, R. W. Dipping reflectors beneath old orogens: A perspective from the British Caledonides. *GSA Today* **6,** 1–6 (1996).
19. Mona Lisa Working Group, Closure of the Tornquist Sea: constraints from MONA LISA deep reflection seismic data. *Geology* **22,** 617–620 (1997).
20. Boschi, L., Ekstroem, G. & Kutowski, B. Multiple resolution surface wave tomography, the Mediterranean basin. *Geophys. J. Int.* **157,** 293–304 (2004).
21. Watts, A. B. & Burov, E. Lithospheric strength and its relationship to the elastic and seismogenic layer thickness. *Earth Planet. Sci. Lett.* **213,** 113–131 (2003).
22. Pollack, H. Cratonization and thermal evolution of the mantle. *Earth Planet. Sci. Lett.* **80,** 175–182 (1986).
23. Ussami, N., Cogo de Sa, N. & Molina, E. C. Gravity map of Brazil. 2. Regional and residual anomalies and their correlation with major tectonic provinces. *J. Geophys. Res.* **98,** 2199–2208 (1993).
24. Hartley, R., Watts, A. B. & Fairhead, J. D. Isostasy of Africa. *Earth Planet. Sci. Lett.* **137,** 1–18 (1996).
25. Pearson, D. G. *et al.* Archaean Re-Os age for Siberian eclogites and constraints on Archaean tectonics. *Nature* **374,** 711–713 (1995).
26. Dixon, J. E., Dixon, T. H., Bell, D. R. & Malservisi, R. Lateral variation in upper mantle viscosity: role of water. *Earth Planet. Sci. Lett.* **222,** 451–467 (2004).
27. Christensen, N. I. & Mooney, W. D. Seismic velocity, structure and composition of the continental crust—global view. *J. Geophys. Res.* **100,** 9761–9788 (1995).
28. Doin, M. P., Felitout, L. & Christensen, U. Mantle convection and the stability of depleted and undepleted continental lithosphere. *J. Geophys. Res.* **102,** 2771–2787 (1997).
29. Lenardic, A., Moresi, L. N. & Mulhaus, L. N. Longevity and stability of cratonic lithosphere: Insight from numerical simulations of coupled mantle convection and continental tectonics. *J. Geophys. Res.* **108,** doi: 10.1029/2002JB001859 (2003).
30. Laske, G., Masters, G. & Reif, C. The Reference Earth Model Website. ⟨http://mahi.ucsd.edu/Gabi/rem.html⟩ (2000).
31. Goodwin, A. M. *Principles of Precambrian Geology* (Academic, San Diego, California, 1996).
32. Brendan Murphy, J. & Nance, D. Sm-Nd isotopic systematics as tectonic tracers: an example from West Avalonia in the Canadian Appalachians. *Earth Sci. Rev.* **59,** 77–100 (2002).
33. Wessel, P. & Smith, W. H. F. Free software helps map and display data. *Eos* **72,** 441–446 (1991).

Supplementary Information is linked to the online version of the paper at www.nature.com/nature.

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