Destruction and regrowth of lithospheric mantle beneath large igneous provinces

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Abstract

Large igneous provinces (LIPs) are formed by enormous (i.e. frequently $> 10^6$ km^3), but short-lived magmatic events that have profound effects upon global geodynamic, tectonic, and environmental processes. Lithospheric structure is known to modulate mantle melting, yet its evolution during and after such dramatic periods of magmatism is poorly constrained. Using geochemical and seismological observations, we find that magmatism is associated with thin (i.e. $\lesssim$ 80 km) lithosphere and we reveal a striking positive correlation between the thickness of modern-day lithosphere beneath LIPs and time since eruption. Oceanic lithosphere re-thickens to 125 km, while continental regions reach $>$ 190 km. Our results point to systematic destruction and subsequent regrowth of lithospheric mantle during and after LIP emplacement and re-cratonisation of the continents following eruption. These insights have implications for the stability, age and composition of ancient, thick, and chemically distinct lithospheric roots, the distribution of economic resources, and emissions of chemical species that force catastrophic environmental change.

1 Introduction

The thermomechanical structure of the lithospheric plate and its interaction with the underlying asthenosphere influence the evolution of Earth's topography, plate motions and deformation, the pattern of mantle convection, and the location and style of volcanism (*1–5*). Consequently, lithospheric evolution drives erosional and depositional processes, controls the distribution of economic resources, and influences climatic and environmental conditions through space and time (*6–9*). Geological and geophysical observations provide constraints on present-day lithospheric structure, but only snapshots of its variations through time [e.g. (*8, 10–14*)].

It is often assumed that the lithospheric structure of ancient continental interiors (i.e. cratons) is stable and unchanging over very long time periods, yet there is a paucity of data with which to test this hypothesis [see (*15*) for a comprehensive review]. This interpretation is generally based upon the exposure of exclusively ancient, undeformed shields at low elevations with low relief (*16*). These shields overlie a thick, depleted lithospheric mantle (i.e. up to $\sim$ 200 km in thickness) that is inherently buoyant, viscous, and difficult to convectively remove (*5, 16, 17*). However, several of these cratons are capped by Phanerozoic large igneous provinces (LIPs) — rare, but giant outpourings of lava ($> 10^6$ km^3) that were extruded within geologically short time frames ($<$ 5 Myr) — which are commonly linked to the arrival of plume heads beneath the plate (*18*). For example, the Siberian Traps, Central Atlantic Magmatic Province, and Karoo flood basalts overlie the Siberian, West African, and Kalahari cratons, respectively. This association calls into question the assumption that cratonic lithosphere is always thick and stable since a thin tectonic lid is required to generate large volumes of melt, and present-day magmatism is strongly associated with lithosphere $<$ 100 km thick (*1, 19–22*).

In this contribution, we attempt to reconcile these seemingly contradictory observations by tracking lithospheric evolution during and after magmatism. We begin by comparing the locations and ages of modern and ancient oceanic islands, seamounts and plateaux to estimates of plate age and lithospheric thickness. Subsequently, this analysis is extended into continental regions, where we develop conceptual and numerical models to explain the lithospheric response to intraplate magmatic events. Finally, we describe the implications of our results for the thermochemical structure of cratonic lithosphere and their use in addressing several key geodynamic, economic and environmental questions.

2 Results

2.1 Oceanic Intraplate Magmatism

2.1.1 Recent Intraplate Magmatism

Our starting point is to explore lithospheric thickness beneath modern-day intraplate magmatic provinces. In the oceanic realm, the lithospheric mantle cools and thickens away from mid ocean ridges in a predictable way (*10, 23, 24*). This phenomenon is one of the most widely accepted and well studied in global geophysics. Usefully, it allows expected time-dependent deepening of the lithosphere-asthenosphere boundary (LAB) to be used as a reference template against which deviations in lithospheric thickness can be defined and explored.

We wish to determine whether intraplate magmatism is associated with deviations from the plate-cooling relationship. To carry out this test, we exploit a global database of intraplate magmatic analyses (*20*). This compilation includes major and trace geochemical compositions of mafic rocks at all locations where intraplate magmatism has occurred within the last 10 Ma and has not since moved further than 400 km from the site of eruption (Figure 1a). We estimate modern-day depth to the LAB, z^1_{LAB}, beneath each recent intraplate magmatic province in three ways. First, z^1_{LAB} beneath each location is extracted from the lithospheric thickness model of Hoggard *et al.* (*8*), which is located by contouring the 1175 °C isothermal surface after converting the SL2013sv shear-wave tomographic model into temperature (Figure 1b). This temperature conversion scheme exploits a calibrated elastic-anelastic parameterisation (*25*). Secondly, we use the lithospheric thickness

estimates calculated by inverse modelling of rare earth element concentrations within mafic intraplate magmas carried out by Ball *et al.* [(*20*); see Methods]. This method is predicated upon the partitioning of rare earth elements between mantle melt and residue during partial melting, where the upper limit of melting is taken to represent the LAB. Finally, we estimate z^1_{LAB} by calculating melt equilibration pressure, P_{eq}, and temperature, T_{eq}, of intraplate melts using the thermobarometric scheme developed by Plank *et al.* (*26*). This scheme exploits the sensitivity of major oxide phase concentrations within mafic melts to these two properties within the mantle [e.g. (*26, 27*), see Methods]. The global database of Ball *et al.* (*20*) is filtered for samples where $9 < \mathrm{MgO} < 14.5$ wt% in order to mitigate removal or addition of material by fractionation or contamination. Then T_{eq} and P_{eq} are calculated for the remaining samples. We locate samples that equilibrated beneath the LAB by selecting samples where $T_{eq} > 1175$ °C. This screening procedure ensures consistency with tomographically determined LAB depths. Next, samples are geographically divided into $1 \times 1°$ bins. We exclude bins with < 10 samples since it is unlikely that a very small sample size can be used to accurately locate the base of the lithosphere. Finally, the minimum value of P_{eq} is calculated within each bin, which enables an approximation of maximum depth to the LAB (i.e. $z_{LAB} \approx 31.4 P^{min}_{eq}$; values of P_{eq} and T_{eq} are given for each sample in the Supplementary Information).

Figure 1c and Figure S1 show the three independent estimates of observed z^1_{LAB} beneath recent oceanic intraplate magmatic provinces. We present our results as a function of the age of oceanic crust, t_{crust}, atop which the magmatic provinces rest (*8, 28*). All else being equal, z^1_{LAB} beneath these recently-erupted provinces should be governed entirely by plate cooling as a function of t_{crust}. These *observed* LAB depth constraints are therefore compared to two estimates of *expected*, unperturbed LAB depth. First, we calculate expected LAB depth using a plate-cooling model constructed by Richards *et al.* [i.e. black line on Figure 1c (*29*)]. Secondly, we compare to the global median LAB depth as a function of oceanic crustal age determined by Hoggard *et al.* [i.e. black circles on Figure 1c (*8*)]. These two estimates of expected z^1_{LAB} values are in close agreement, although it is important to note that they are not entirely independent of one another [see (*8, 24, 25*)].

For magmatic provinces located above young oceanic crust (i.e. $t_{crust} < 30$ Ma), predictions and observations are in close agreement, indicating that z^1_{LAB} increases from ~ 20–50 km to 60–80 km during this time period. For crustal ages $t_{crust} > 30$ Ma, however, expected and observed LAB depths beneath intraplate magmatic provinces dramatically diverge. For an oceanic crustal age, $t_{crust} \gtrsim 30$ Ma, the plate is systematically thinner by 30–60 km than would be expected given the crustal age (compare black line and circles with coloured symbols on Figure 1c). These observations suggest that magmatism occurs exclusively where $z_{LAB} < 100$ km, and in most cases only where, $z_{LAB} \lesssim 80$ km.

To make sense of this relationship, consider first the Icelandic hotspot (ⓘ on Figure 1). This magmatic province straddles the Mid-Atlantic Ridge, where lithospheric mantle is thin or absent due to active sea-floor spreading. Consequently, all lithospheric thickness proxies suggest that melts are able to ascend to a shallow level, unimpeded by the presence of lithospheric mantle, before encountering the Moho at depths of ~ 15–45 km (*30, 31*). Next, consider the Hawaiian hotspot, which has $t_{crust} \sim 90$ Ma (ⓘⓘ on Figure 1). The LAB ought to be found at $z^1_{LAB} \approx 100$ km according to both predicted plate cooling and the global average for oceanic lithosphere of that age. However, major oxide concentrations suggest that Hawaiian melts equilibrate within the asthenosphere to depths as shallow as 30–60 km (see yellow stars on Figure 1c). Furthermore, rare earth element inverse modelling suggests that the top of the asthenospheric melt column is located at 40–60 km depth (see red squares on Figure 1c). Finally, local and global seismic imaging indicates that beneath Hawaii $z^1_{LAB} \sim 40$–90 km [e.g. (*8, 32, 33*), see green triangles on Figure 1c]. This thinner lithosphere is consistent with flexural and gravimetric observations, as well as with xenolith-based thermobarometric results which indicate that temperatures of 1000–1100 °C occur at depths of 45–55 km beneath Hawaii (*34, 35*). In contrast to the Icelandic example, this discrepancy suggests that, locally, z^1_{LAB} is significantly shallower than expected beneath Hawaii. Taken together, these considerations strongly suggest that when the plate thickens beyond 70–80 km (i.e. when $t_{crust} \gtrsim 30$ Ma), magmatism is associated with marked lithospheric thinning to depths of 60–80 km. In other words, although a precise mechanism cannot be discerned, magmatism is associated with the resetting of z_{LAB} to depths typically found beneath oceanic crust

with an age of ~ 20–40 Ma.

2.1.2 Ancient Intraplate Magmatism

Following cessation of magmatism, lithosphere that is < 80 km thick should re-thicken as a result of conductive cooling. Modern-day LAB depth, z^1_{LAB}, should therefore be thicker beneath older intraplate magmatic provinces than those erupting at present. We test this hypothesis by substantially updating a global map of oceanic plateaux, islands and seamounts [(*36, 37*); Figure 1b; Supplementary Information Section S2; Dataset S3]. This database contains polygons outlining these oceanic landforms and an associated eruption age, t_0, of the magmatic rocks that form them. Polygons are added, excluded and their age constraints updated according to recent literature (see Supplementary Information Section S3; Figure S2). We sample z^1_{LAB} at $0.1 \times 0.1°$ intervals beneath each polygon in our updated database. Once again, we exploit estimates of lithospheric thickness calculated by Hoggard *et al.* (*8*). Next, we filter our database to ensure that only the final volcanic episode in a given location is included. Thus we capture the passive response of the lithosphere to magmatism and avoid contamination by subsequent magmatic events. We remove any location that has experienced another phase of magmatism within a distance of < 500 km after > 30 Myr of the initial eruption. Note, however, that we do not control for any subsequent amagmatic tectonic thinning or thickening of the lithosphere. We calculate the moving median and interquartile range of z^1_{LAB} values beneath the remaining intraplate magmatic provinces as a function of t_0. In calculating each median value, we weight the z^1_{LAB} value of each point as a function of latitude in order to avoid biasing moving average estimates towards polar provinces (see Methods).

On Figure 1d, the black circles with error bars show observed values of z^1_{LAB} as a function of t_0. The median LAB depth beneath the youngest provinces (i.e. where eruption age, t_0 < 10 Ma) are consistent with the median thickness calculated for the database of Ball *et al.* (*20*), as well as our new thermobarometric constraints (see Methods; Section 2.1.1). Importantly, we find that there is a positive relationship between province age and LAB depth beneath the province up to an age of ~ 80–100 Ma, at which point the deepening of z^1_{LAB} flattens off (see black circles with error bars on Figure 1d). The shape of this relationship resembles a plate cooling model that flattens at a depth consistent with several well-known, published schemes [i.e. 100–130 km; Figure 1c and d; e.g. (*10, 23, 24*)].

Since older oceanic plate is associated with thicker lithosphere, it is possible that any positive relationship between z^1_{LAB} and t_0 is generated by chance. This general trend could be produced by distributing intraplate magmatism randomly upon a lithospheric plate whose thickness is governed solely by plate cooling as a function of oceanic crustal age. This relationship would arise because provinces with an eruption age of, say, t_0 = 180 Ma can only possibly be found today on old lithosphere, which in the absence of any thermal perturbation would be around 125 km thick [e.g. (*24*)]. However, recent intraplate magmatism (i.e. t_0 = 0 Ma) could theoretically be distributed over lithosphere of thickness ranging from zero at a mid-ocean ridge to 125 km on the abyssal plain. The median LAB depth of this distribution would therefore be shallower than for those locations overlying only older oceanic crust. As a result, even if z^1_{LAB} is controlled only by oceanic crustal age, younger intraplate magmatism is expected to overlie, on average, thinner lithosphere than sites of older magmatism. A sample of young seamounts, islands and plateaux would also have a broader distribution of z^1_{LAB} values than older provinces. However, if magmatic emplacement is associated with systematic thinning of the plate (i.e. shallowing of the LAB), then observed values of z^1_{LAB} beneath oceanic plateaux, islands and seamounts ought to be systematically thinner for any t_0 < 80–100 Ma than the z^1_{LAB} values expected from plate cooling alone.

To investigate whether temporal trends in z^1_{LAB} provide evidence for magmatism-related lithospheric thinning, we calculate the expected, thermally unperturbed z^1_{LAB} distribution beneath a random sample of oceanic islands, seamounts and plateaux as a function of t_0 given plate cooling alone (see Methods). As expected, we find that, indeed, there is a slight positive relationship between plate age and unperturbed lithospheric thickness beneath ancient oceanic magmatic provinces (see blue line and envelope on Figure 1d). This predicted distribution is then compared to the observed

distribution (i.e. black circles with error bars on 1d). Importantly, up to an eruption age, $t_0 \approx 90$ Ma, ancient magmatic provinces systematically overlie lithosphere that is $\sim$ 20–40 km thinner than would be expected if the plate were not perturbed since sea-floor spreading (compare black circles and blue line on Figure 1d). This result is not materially affected by using alternative lithospheric thickness models (see Supplementary Information Section S4). This finding further supports our earlier insight that intraplate magmatism resets the plate cooling process by thinning the lithospheric mantle. It is also consistent with seminal work by Detrick & Crough (*38*), who showed that ocean island subsidence cannot be explained only by seafloor age. Instead, anomalous subsidence requires either lithospheric thinning followed by time-dependent rethickening, and/or translation away from a sub-plate thermal anomaly (*39*). Intriguingly, our results suggest that the thickness of the plate beneath ancient seamounts and plateaux may be better predicted using the eruption age of the province than by oceanic crustal age.

2.2 Continental Large Igneous Provinces

The initial rejuvenation and subsequent progressive thickening of the oceanic lithosphere over time following intraplate volcanism raises an important question: does continental lithosphere behave in a similar way? Answering this question is complicated by the fact that lithospheric thickness in the continental realm — unlike in the oceans — does not generally increase as a function of crustal age. Nevertheless, Ball *et al.* (*20*) showed by extracting LAB depth from the model of Hoggard *et al.* (*8*) and modelling rare earth element concentrations in intraplate magmas, that $z^1_{LAB} \lesssim 100$ km beneath modern continental intraplate magmatic provinces. Similarly, our proxy for z^1_{LAB} based upon equilibration pressures and temperatures of major oxide phases in mafic melts agrees that almost all modern magmatic provinces are associated with $z^1_{LAB} \lesssim 80$ km [(*20*) see Methods; Supplementary Information Section S2; Figure S1]. Furthermore, White & McKenzie (*40*) modelled rare earth element concentrations in ancient magmatic provinces. They showed that, irrespective of the often substantial thickness of lithosphere beneath each province today, these ancient mafic magmatic rocks were also associated with thin lithosphere at the time of their formation. The question therefore remains as to how the lithosphere evolves following eruption of intraplate magmatic rocks in the continental realm. We now return to our updated database of ancient intraplate magmatic provinces to explore whether the relationship between z_{LAB} and intraplate magmatism also applies to the continents.

On the continents, this database is mostly comprised of large igneous provinces (LIPs), a dominance that is particularly acute prior to Cenozoic times. In the same way as our analysis in the oceanic realm, we have expanded and improved the database of Coffin *et al.* (*37*) by refining province ages and polygon outlines using recently published studies. We also now incorporate provinces up to to 750 Ma [e.g. (*18*); to access the database and find information on individual changes see Supplementary Information Section S3 and Figure S2]. Similarly, we also again exclude any location where subsequent magmatism occurred within 500 km of the eruption site after 30 Myr of the initial eruption so that only the final phase of intraplate magmatism is included in our database. We again sample z^1_{LAB} within each polygon at $0.1 \times 0.1°$ intervals.

Figure 2 shows depth to the LAB beneath continental LIPs as a function of time since their emplacement (i.e. $z^1_{LAB}(t_0)$). For the first time, we reveal a striking relationship between LAB depth and time elapsed since continental LIP emplacement. For recent provinces, the lithosphere is $\sim$ 50-70 km thick, which is consistent with that observed in the oceans and a range of geochemical proxies for lithospheric thickness [i.e. estimates calculated by ref (*20*) and new estimates derived from major element thermobarometry]. LAB depth progressively increases as a function of time since LIP emplacement to approximately 195 km at 250 Ma before flattening off. For example, the Ethiopian Flood Basalts were erupted at $t_0 \sim 30$ Ma, with magmatism continuing to recent times, and are underlain by lithosphere that is < 50 km thick (*19, 20*). The Paraná-Etendeka Traps, where $t_0 \sim$ 138-128 Ma , on the other hand are underlain by lithosphere that is 100-150 km thick. However, 195 km-thick lithosphere underlies the Siberian Traps, which was erupted at $t_0 \sim 250$ Ma. This thickness is similar to the LAB depth beneath the $\sim$ 510 Ma Kalkarindji LIP in central and northwestern Australia,

and the ~ 750-710 Ma Franklin LIP in the Canadian Arctic [e.g. see ref (*18*); Supplementary Information Section S2]. The Deccan Traps (t_0 ~ 66 Ma) mark the only significant outlier to this relationship, having markedly thicker lithosphere than LIPs of a similar age. In general, and with reference to our oceanic analysis, this result hints at lithospheric rejuvenation prior to, or during, magmatism before relaxation after magmatism ceases. Significantly, the depth at which z^1_{LAB} stops increasing and flattens off (i.e. ~ 195 km) is about 60-90 km deeper than that observed in the oceans (compare Figures 1d, black circles and 2b).

We acknowledge that lateral transport of lava flows or dykes may mean that we sample lithosphere that is far away from the eruption site. In Supplementary Information Section S4.1 we show the raw, unaveraged values of z^1_{LAB} as a function of eruption age. We further test our results by exploiting two alternative compilations of LIP eruption centres from Torsvik *et al.* (*41*) and a new compilation assembled for this study (See Supplementary Information Sections 4.2 and 4.3). We also verify that this relationship is robust against a raft of alternative lithospheric thickness models, and is consistent with gravity observations (Supplementary Information Section S4.2). Finally, it is important to note that our analysis encompasses LIPs erupted onto all types of continental environments, ranging from shields (e.g. eastern Siberian Traps) to Phanerozoic sediments (e.g. Central Atlantic Magmatic Province, i.e., CAMP, and Karoo).

2.3 Thermal Modelling

It is clear that intraplate magmatism occurs above thinner-than-expected lithosphere and that older magmatic provinces reside above progressively thicker lithosphere. Rifting, which leads to rapid thinning of the lithosphere (i.e. shallowing of the LAB), is usually followed by protracted time-dependent re-thickening as a result of conductive cooling. Similarly, the systematic increase in LAB depth beneath magmatic provinces from young to old suggests that the plate thickens as a function of time following the cessation of magmatism. This deepening eventually stalls and flattens out at a constant thickness of around ~ 195 km beneath continental provinces and ~ 125 km in the oceans. Two principal models have been developed to explain time-dependent thickening of the lithospheric plate. First, a half-space model, in which the lithosphere cools and thickens indefinitely as a function of age, and secondly a plate model, in which the lithosphere cools and thickens, initially closely matching the half-space model, before approaching a finite thickness controlled by the convective resupply of basal heat [e.g. (*10, 42, 43*)]. The impact of this heating is generally approximated by imposing a constant basal potential temperature, T_p, at a particular depth, z_p, that defines the maximum depth to which conductive cooling can penetrate (*24, 29*). Note that z_p is not the same as z_{LAB}. Here we investigate whether conductive cooling, represented in this case by a plate model, can explain the time-dependent thickening of the lithosphere, $z_{LAB}(t)$, following intraplate magmatism.

2.3.1 Oceanic Intraplate Magmatic Provinces

Again we begin with the oceanic lithosphere. We model 1-D thermal evolution, $T(z,t)$, beneath intraplate magmatic provinces and hence depth to the LAB as a function of time, $z_{LAB}(t)$, by making three simple modifications to the plate model of Richards *et al.* [(*24, 29*); Methods]. First, rather than assuming an initial temperature profile consistent with mid-ocean ridge conditions (i.e., LAB depth ~10 km below sea-level), a steady-state initial temperature profile with an LAB depth equal to 50 km is imposed, so that z^0_{LAB} = 50 km when magmatism occurs. In practice, this constraint means initially placing the 1175 °C isotherm at a depth of 50 km. This assumption is consistent with both the results of our geochemical modelling and tomographically constrained LAB depth beneath intraplate magmatic provinces. Secondly, we assume that the less conductive oceanic crustal layer is 20 km rather than 7 km thick, a choice that reflects elevated thickness beneath magmatic provinces relative to standard oceanic crust. Note that we ignore the effects of viscoelastic relaxation, which means that the topographic load generated by this thicker crust is instantaneously compensated. Finally, observational studies have strongly suggested that initial excess subsidence of ocean islands is generated by movement

away from swell bathymetry that is at least partly generated by excess mantle temperatures (*39*). Hence, to account for the excess heat provided by a putative mantle plume, initial steady-state temperature profiles corresponding to elevated initial mantle potential temperature, $(T_p)_0$, are imposed. As the thermal structure evolves, $(T_p)_0$ decays linearly over time (30 Myr) to the ambient mantle value adopted by Richards *et al.*, T_p = 1333°C [see Methods, (*29*)]. This evolving temperature excess is initially imposed at the base of the thermal boundary layer (i.e., the shallowest depth at which $\frac{dT}{dz}$ drops below 0.5°C km^{-1}), before deepening at a rate of 10 mm yr^{-1}. This progressive temperature decay and vertical advection of the thermal boundary simulate the movement of the plate away from a plume-like mantle heat source (*39*). We test the effect of varying the decay time of this thermal anomaly and its precise parametrisation, but it makes minimal difference to our results (see Supplemenatry Information Section S5). To find the optimum plate model that best describes our $z_{LAB}(t)$ observations, we carry out a grid search in which pairs of z_p and $(T_p)_0$ are systematically explored and optimized.

Our results show that cooling and re-thickening of the lithosphere from an initial thickness of ~50 km following the cessation of magmatism can explain the distribution of oceanic LAB depth as a function of time since magmatic eruption. Importantly, the optimal plate thickness is closely comparable to that obtained by fitting global oceanic bathymetry and heat flow data as a function of plate age (*24, 29*). The significant difference is that while the best-fitting value of z_p is only $\sim 17 \pm 10$ km in excess of that required to explain global oceanic subsidence and heat flow, an initial excess temperature anomaly of between 0 and +200°C relative to the global background is required to fit $z_{LAB}(t)$ beneath intraplate magmatic provinces [Figure 3a; (*29*)]. This temperature anomaly then wanes rapidly. This result is consistent with both excess asthenospheric potential temperatures associated with at least some proportion of intraplate magmatism and constraints from the drowning of ocean islands (*20, 39, 40*).

2.3.2 Continental Large Igneous provinces

Next we extend this model to the continental realm. First, we maintain the condition that the the initial LAB depth is 50 km (i.e. z^0_{LAB} = 50 km). This assumption is supported by modelling of rare earth element compositions of mafic and ultramafic melts in LIP magmas (*40*). Secondly, we assume that continental crustal thickness, z_{cc} = 35 km, and account for differences in thermal properties associated with its more felsic composition including enhanced radiogenic heat production, H^*, and higher thermal conductivity [see Methods; (*44*)]. As before, the initial temperature profile is determined by finding the steady state geotherm, given the value of $(T_p)_0$, that results in the 1175 °C isotherm at a depth of 50 km. We carry out a grid search to determine the best-fitting values of $(T_p)_0$ and z_p that are consistent with observed values of z_{LAB} as a function of LIP age. The Deccan Traps are excluded from our analysis since its two contributing temporal bins are clear outliers (see Supplementary Information Section S5).

Results show that, as for the oceans, the thickness of the lithospheric mantle beneath continental LIPs can be fitted with a simple adapted plate-cooling model, although with some significant differences. Importantly, the best-fitting potential temperature is $(T_p)_0 = 1465 \pm 120$ °C, which, while less well defined, overlaps with the range of values of $(T_p)_0$ in the oceanic realm (Figure 3). Indeed, these results suggest that significantly elevated mantle potential temperatures (i.e. $(T_p)_0 > 1360$ °C) are a systematic requirement of LIP formation, providing independent evidence for their formation above a mantle plume (*40, 45, 46*). However, the optimal plate thickness, z_p = 248 km, is significantly greater than in the oceans, which implies a steady-state LAB depth of $z^\infty_{LAB} \approx 195$ km. This depth aligns more closely with lithospheric thicknesses inferred from subsidence in intra-cratonic basins (i.e. ~ 200 km) rather than either subsidence observed in rift basins on non-cratonic continental lithosphere or the ocean floor. [i.e. ~ 100–150 km; (*10, 24, 47, 48*)]. Note that, although we have chosen to exclude the Deccan Traps outliers from our analysis, the best-fitting value of z_p is minimally affected by their inclusion (i.e., an increase of 2 km), while optimal $(T_p)_0$ remains ~ 80 °C above ambient values (i.e., 1415 °C versus 1333 °C; see Supplementary Information Section S5.1).

In summary, our analysis, in which we fit LAB depth estimates with conductive cooling models, suggests that the lithospheric mantle is rapidly thinned immediately prior to and/or during magmatism in both oceanic and continental realms. It then cools and progressively re-thickens following magmatism. For continental LIPs, this thinning is associated with a significant temperature anomaly and steady-state plate thicknesses are similar to those in intra-cratonic basins.

3 Discussion

Our global analysis of lithospheric thickness beneath intraplate magmatic provinces has shown that magmatism occurs exclusively on thin lithosphere. Following eruption, the lithospheric mantle progressively cools and re-thickens to a depth of 125 km in the oceans and 195 km on the continents. These inferences are supported by a range of observations. First, major and trace elemental compositions of recent intraplate magmas suggest that magmatism occurs only where the lithosphere is less than 80 km thick. Secondly, the thickness of the lithospheric plate calculated by exploiting seismic tomographic models and a calibrated V_S-to-T conversion scheme is in broad agreement with these geochemical estimates. Thirdly, the thickness of the lithosphere beneath ancient intraplate magmatic provinces is systematically thinner than would be expected if they were randomly distributed atop oceanic lithosphere that was following the well-documented plate-cooling relationship. Finally, lithospheric thickness defined by seismic tomographic imaging beneath both oceanic and continental intraplate magmatic provinces systematically increases as a function of eruptive age following a plate cooling relationship subject to initially elevated asthenospheric temperatures. A key observation is that several continental LIPs overlie cratons, which are regions of supposedly ancient, thick, and stable lithospheric mantle (e.g. Karoo, CAMP, Emeishan, Siberia, Kola-Dnieper, Kalkarindji, Franklin). Our results suggest that even cratonic continental interiors are significantly less stable than often assumed, with cratonic roots apparently neither able to resist being destroyed during LIP emplacement, nor remaining permanently thinned after the cessation of magmatism. Here we discuss the implications of the relationship between LIP emplacement, rapid lithospheric thinning and protracted thermal relaxation for the compositional make-up of continental lithosphere through time, the formation of economic resources, and the initiation of mass extinction events.

3.1 Re-cratonisation

Previous insights indicate that thick, cratonic lithosphere is likely to have been generated by high degrees of melt extraction at shallow depths followed by lithospheric thickening during ancient (i.e. pre-Phanerozoic) orogenic events [refs (*15, 49*) and references therein]. Our results, however, suggest that in some locations, this pre-existing lithospheric mantle is destroyed prior to, or during, LIP eruption. It is then re-emplaced over the subsequent ~ 300 Ma by thermal relaxation (Figure 2b and 3c). Compositional differences between unperturbed and re-thickened cratonic lithosphere are challenging to distinguish geophysically. However, our results allow useful constraints to be placed upon the compositional evolution of cratonic roots during and after magmatism.

LIPs often appear on ancient cratons (e.g. Siberia, Karoo, Central Atlantic Magmatic Province, Franklin), which are likely to have depleted, low-density lithospheric roots (*16*). Thinning of undepleted lithospheric mantle is a significant means of generating topographic uplift and denudation on 10^3 km lengthscales (*50*). Conversely, thickening of an undepleted lithospheric root drives time-dependent subsidence (*51*). Lithospheric depletion (i.e. density reduction, $\Delta\rho$) modulates the amplitude of this uplift and subsidence. Consider a cratonic setting where the onset of magmatism is associated with a combination of a transient asthenospheric thermal anomaly and removal of an ancient, thermally equilibrated, and depleted (i.e. low-density) cratonic root (Figure 4a–b). In this case, the initial density of the root, ρ_m^* is perhaps 50 kg m^{-3} lower than in non-cratonic lithospheric mantle. Removal of this somewhat depleted root and the sub-plate thermal anomaly together lead to significant amounts of air-loaded surface uplift (i.e. U_o = 1.8 km if

$(T_p)_0$ = 1465 °C; see Methods). Note that if ρ_m^* is significantly more depleted, then the magnitude of this uplift decreases substantially (e.g. $U_o \approx 500$ m for depletion of 80 kg m^{-3}). Subsequent thermal re-equilibration drives re-growth of the lithospheric mantle (Figure 4c–g). If this re-growing lithospheric mantle is completely undepleted, then the result is net air-loaded subsidence of 1.6 km after 350 Ma and 2.1 km after 750 Ma. No net uplift or subsidence occurs when re-thickening mantle lithosphere is depleted by 50 kg m^{-3} (i.e. same density as precursor lithospheric mantle, ρ_m^*). Finally, if depletion is increased to 80 kg m^{-3} then a total of 1.3 km net uplift is generated after 750 Ma (Figure 4f and g; Methods).

Extensive evidence exists for uplift associated with intraplate magmatism, although in cratonic regions it can be controversial [e.g. (*20, 46, 52–55*)]. This controversy perhaps indicates the relatively modest magnitude of initial uplift expected as a result of rapid thinning of depleted cratonic lithosphere. This more modest magnitude is in contrast to the kilometre-scale uplift expected to result from removing the same thickness of fertile lithosphere [e.g. (*50, 56–58*)]. A second important observation is that many LIPs that overlie re-equilibrated lithosphere (i.e older than 300 Ma) are exposed at the surface today and not buried under deep sedimentary basins. It is therefore likely that they have experienced either net uplift or at least no appreciable net vertical motion between the pre-magmatism and post-equilibration lithospheric states. Taken together with the results of our simple modelling, these observations suggest that depleted cratonic lithosphere is removed before or during magmatism, then it is replaced by similarly, or more depleted lithospheric mantle during subsequent thermal relaxation. LIP emplacement therefore provides a mechanism for cratonic destruction followed by re-cratonisation.

Additional evidence corroborates this conclusion. There are two reasons to expect that the crust may sometimes be thinned during the emplacement of an LIP, which would be expected to generate even further net subsidence after lithospheric re-equilibration (i.e. when z_{LAB} reaches z_{LAB}^{∞}). First, the initial asthenospheric thermal pulse and lithospheric removal generate excess topography that is progressively eroded resulting in crustal thinning (*59*). Secondly, many — but not all — LIPs are associated with some degree of lithospheric extension. These processes are offset to some degree by the addition of erupted, intruded, and underplated basaltic material that would tend to thicken the crust (*59*). It is important to note, however, that all that remains of many LIPs is the plumbing system, the flood basalts having been totally denuded (*60, 61*). To expose LIP plumbing systems at the surface rather than have them buried under kilometers of sediment and lava flows, post-LIP topography must acquire permanent sub-crustal support. Lithospheric depletion, potentially modulated by magmatic underplating, is a plausible mechanism for this topographic support. An important corollary is that initial uplift, erosion, and replacement of depleted lithospheric mantle with a subsequently *less* depleted (i.e. denser, in this case) lithospheric keel could be a way to generate subsidence in intra-cratonic basins.

This argument suggests that the emplacement of an LIP is likely responsible for local-to-regional destruction of cratonic lithospheric mantle on short timescales, followed by re-cratonisation on longer timescales. There is a caveat worth highlighting, however. Our analysis provides neither an estimate for the precise thickness and composition of the pre-existing lithospheric mantle beneath individual LIPs, nor a definitive explanation for the mechanism of lithospheric mantle thinning. For example, it is possible that a mantle plume could be directed by basal lithospheric topography into a pre-existing lithospheric thin spot [e.g. (*62–64*)]. There, it would generate surface uplift and extensive magmatism due to its excess potential temperature, before growth of a depleted, neutrally buoyant lithospheric root begins after magmatism, resulting in net-neutral topography. This situation may yield a similar suite of observations to those in the previously discussed case, where a neutrally buoyant cratonic root is first removed. However, independent constraints on paleo-lithospheric thickness from both the presence of kimberlites, and the reconstruction of geothermal gradients using thermobarometry of mantle xenoliths, indicate that the lithosphere was thick prior to several African magmatic events where it is now thinner [see e.g. (*19, 58, 65*)]. Similar observations also are also found in the MacKenzie Flood Basalt Province, Canada (*22*). Furthermore, our results themselves suggest that thin-spots are short lived and heal in the absence of perturbation. Basal thermal erosion by a mantle plume has been proposed as a mechanism for thinning the lithospheric mantle, but this process is challenging to achieve on short timescales (*65*). Alternatively, melt and fluid infiltration could

lead to refertilisation and subsequent destabilisation of the lithospheric keel (*15, 66, 67*).

Regardless of the precursory state of the mantle, simple geodynamic reasoning further supports re-thickening of the lithospheric mantle after magmatism by progressive cooling of depleted restites. The equilibrium thickness of undepleted continental lithospheric mantle is around 100–150 km (*47, 51*). We find that the equilibrium thickness of the thermal boundary layer in continental regions affected by large igneous province emplacement is closer to 180–220 km. McKenzie & Richter (*68*) showed that the thickness of the upper thermal boundary layer, δ, is proportional to $\nu^{\frac{1}{3}}$, where ν is kinematic viscosity (i.e. η/ρ, where η is dynamic viscosity and ρ is density). At a given temperature, melt depletion leads to an increase in viscosity of the mantle residue [e.g. (*69, 70*)]. Hence a doubling of ν leads to an increase in δ of around one quarter. Geochemical depletion of the uppermost mantle therefore leads to a thickening of the upper thermal boundary layer relative to ambient mantle.

Importantly, although time-dependent deepening of the LAB is a globally robust relationship, excess deepening is not observed in the oceanic realm, where z^{∞}_{LAB} = 125 km. In the oceans, the LAB seldom exceeds depths of > 150 km, regardless of the presence or absence of LIPs. Furthermore, in the oceans, significant post-magmatic subsidence is observed (*38, 39*). A mechanism is therefore needed to re-thicken depleted mantle on the continents while perhaps not in the oceans. Numerical modelling carried out by Liu *et al.* (*22*) indicates that lithospheric thin spots surrounded by thick cratonic lithosphere can trap accumulations of low-density, depleted mantle material. This material can be generated by *in situ* melt extraction within the plume head during LIP formation, or can be composed of remnant parcels of previously removed cratonic mantle. Their results show that in these circumstances the lithosphere can recover up to 90% of its original thickness as depleted residue cools over the subsequent 300 Ma, almost entirely healing any putatively destroyed precursory cratonic lithospheric mantle. By contrast, their analysis implies that when LIPs are emplaced in oceanic regions distant from thick cratonic lithosphere, melt-depleted material is more easily entrained into the sub-lithospheric mantle, limiting the eventual re-thickening of the oceanic lithosphere. Our observations strongly support this idea and suggest that, despite their rarity, where plumes are able to erode the cratonic lithospheric mantle, plume-driven re-cratonisation is routine in the continental realm.

3.2 Economic Resources

Our results provide a predictive framework for modelling the thermal evolution of undeformed continental interiors in response to magmatic events. The thinning and re-thickening pattern can act as a useful tool to highlight and evaluate prospects for economic resources that are dependent on the thermal evolution of the lithosphere, including critical minerals [e.g. (*8*)]. For example, diamonds primarily form within cold, thick continental lithospheric roots (i.e., 1130 ± 120 °C at 5.3 ± 0.8 GPa) and are brought to the surface within kimberlite pipes (*71*). Our results suggest that thinning of the plate during/prior to emplacement of an LIP will suppress diamond formation and entrainment for ~ 300 Myr as the plate rethickens beyond $\gtrapprox$ 170 km (Figure 2b). Several LIPs are spatially proximal to older, diamondiferous and/or younger, barren kimberlite fields. Such locations include the Siberian Traps, which erupted 75–45 Myr after the diamondiferous Alakit, Upper Muna and Daldyn fields and 0–30 Myr before the barren, traps-related Kharamay field (*72*). The Keweenawan LIP on the southern Superior craton, North America, at 1150-1100 Ma preceded/coincided with the barren 1100 Ma Kyle Lake kimberlite field but, crucially, the much more recent Attawapiskat kimberlites, which occurred in the same location at 180–150 Ma, were diamondiferous (*73*). Similar logic applies to the Artemisia kimberlites, which became diamondiferous 600 Ma after emplacement of the MacKenzie Flood Basalts above the Slave Craton, Canada (*22*). Previous authors have tied these relationships to the heating of lithospheric roots out of the diamond-bearing window by mantle plumes during LIP events (*19, 73–75*). We contend that rapid and wholesale lithospheric removal may be a key mechanism (*19*). Moreover, our observations reveal a predictive timescale over which cratonic keels will return to a state favourable to diamond formation, which is similar to results of previous modelling studies [~ 300 Myr, (*22*)].

3.3 Climate Change and Mass Extinctions

Finally, our results have implications beyond the solid Earth since LIP eruptions release large quantities of chemical species that can lead to global environmental change. When carbon, sulfur, and halogens are expelled into the atmosphere, they warm the planet, cool the planet, and deplete the ozone layer, respectively (*76*). The rapid and large-scale eruption of these climate-forcing gases during LIP emplacement is thought to be responsible for many of Earth's mass extinctions (*7, 77, 78*). The Siberian Traps ($\sim$ 1–2 $\times$ 10^6 km^3) coincides with the largest known extinction event, which defines the Permian-Triassic boundary and represents the loss of > 80% of marine species (*79*). However, conventional mantle plume-derived melts are insufficiently concentrated in C, S, and halogens to generate the catastrophic environmental change required to drive such a mass extinction (*4, 76*). Metasomatized continental lithospheric mantle, on the other hand, is rich in these elements (*80*). Analysis of mantle xenoliths emplaced during and after eruption of the Siberian Traps demonstrates that the sub-continental lithospheric mantle stored abundant halogens, $\sim$ 70% of which were scavenged by ascending melts (*80*). Therefore, destruction of the sub-continental lithospheric mantle and release of the volatiles it contains during LIP formation is a viable trigger for the Permian-Triassic boundary extinction, and was probably enhanced by release of S and C from sediments as a result of magmatic heating (*80–82*). While it has been previously argued that destruction of the lithospheric mantle beneath the very largest LIPs contributes to the most significant mass extinction events, a temporal, if not mechanistic link has also been drawn between mass extinctions and smaller LIPs. For example, the Emeishan Traps, which are around $\sim 3 \times 10^5$ km^3 in volume (i.e. $< 1/3$ Siberian Traps), and erupted above the Yangtze Craton, have been linked to the end-Guadalupian mass extinction event (*83, 84*). Alteration and removal of the lithospheric mantle beneath even the smaller LIPs could therefore be a contributing factor to the environmental forcing that leads to mass extinction events.

Materials and Methods

Large Igneous Province Database

Coffin & Eldholm (*36*) and Coffin *et al.* (*37*) mapped the outlines of Phanerozoic large igneous provinces (LIPs) and oceanic seamounts, and provided approximate eruption ages for each location. We have updated this database to include LIPs up to 750 Ma. We further amended the ages and outlines of a number of polygons contained in the original database to honor updated radiometric dates. A list of sources for the updated database can be found in the Supplementary Material.

Seismologic Estimates of Lithospheric Thickness

Throughout this study we exploit the lithospheric thickness model of Hoggard *et al.* (*8*). This model converts shear-wave velocities, V_S, into temperature, T using the calibration method of Richards *et al.* (*25*) and the upper-mantle V_S tomographic model of Schaeffer & Lebedev (*85*). This temperature conversion scheme relies on the experiment-based parameterisation of Yamauchi & Takei (*86*), which accounts for the effects of anelasticity on V_S as the mantle melting temperature is approached. The equations include seven material constants that are calibrated by exploiting the thermal structure beneath mid-oceanic ridges in four different ways. First V_S profiles are extracted along oceanic plate flow lines and are averaged to yield V_S as a function of plate age and depth. These stacked profiles are then compared to predicted values of V_S from applying the V_S-to-T conversion to a plate-cooling model (*24*). Misfit is calculated between observed and modeled V_S at depth slices of 87.5 and 112.5 km. Since oceanic plate thickness is generally $\leq$ 125 km, it is not possible to use this approach at greater depths. Secondly, V_S profiles are averaged globally at depths beneath the thermal boundary layer (i.e. between 225 and 400 km) and are compared to values of V_S predicted by applying the V_S-to-T conversion to a mantle isentrope with a potential temperature, T_p = 1333 °C, which is assumed to represent ambient mantle. Thirdly, the globally averaged shear wave attenuation as a function of depth beneath old (> 100 Ma) oceanic lithosphere is calculated from V_S using the same conversion and compared to published estimates at depths > 150 km. Finally, the misfit between published and calculated depth-averaged mantle viscosity beneath the thermal boundary layer is evaluated.

These four misfit functions are combined, weighted and minimised to optimise the values of the seven unknown material constants in the anelastic parameterisation. Finally, lithospheric thickness is calculated globally by extracting the 1175°C isotherm, which is assumed to be a proxy for the depth to the base of the mechanical boundary layer. For further detail on the methodology, see Hoggard *et al.* (*8*) and Richards *et al.* (*25*).

Geochemical Estimates of Lithospheric Thickness

Major Element Thermobarometry

Partitioning of major elements between melt and solid mantle is pressure and temperature dependent. The Plank & Forsyth (*26*) thermobarometer exploits a comprehensive suite of melt-equilibration experiments to parameterise the response of major-element composition to pressure and temperature changes. To estimate lithospheric thickness, we apply this thermobarometer to a global database of Neogene–Quaternary mafic intraplate magmatic samples compiled by Ball *et al.* (*20*). These calculations are carried out using meltPT – an open-source Python library designed for whole-rock thermobarometric analysis [https://github.com/fmcnab/meltPT; (*27*)].

Prior to estimating melt equilibration conditions, the effects of fractional crystallisation are mitigated by filtering our magmatic database so that all samples have 9 < MgO < 14.5 wt%. Samples must also include a measurement of Ce concentration so that H_2O can be estimated assuming that H_2O/Ce = 200 (*87, 88*). We calculate the primary melt compositions of each sample by adding olivine that is in equilibrium with the melt until olivine forsterite content is 0.9 (*89*).

Since only Fe^{2+} is compatible within olivine, an estimate of $Fe^{3+}/\Sigma Fe$ is required. We assume that $Fe^{3+}/\Sigma Fe = 0.2$, which is halfway between the average values for theoliitic and alkali basalts (*90*).

These primary melt compositions are used to generate estimates of the pressure and temperature at which mantle melts equilibrated [i.e. P_{eq} and T_{eq}, respectively (*26*)]. Samples may have last equilibrated within the lithosphere or asthenosphere, and the boundary between these layers is defined throughout our study by the 1175 °C isotherm. In a given location, we assume that the maximum lithospheric thickness coincides with the equilibration pressure of the shallowest sample that has an equilibration temperature $T_{eq} > 1175$ °C since temperatures cooler than this value are likely to indicate equilibration occurred within the lithosphere. To estimate lithospheric thickness globally, we subdivide our thermobarometric results into $1° \times 1°$ geographic bins. We only estimate lithospheric thickness for bins with ≥ 10 samples since smaller sample sizes are less likely to capture the lithosphere-asthenosphere boundary.

Rare Earth Element Inverse Modelling

We exploit the results of rare earth element (REE) inverse modeling presented by Ball *et al.* (*20*), who use an adapted form of the INVMEL-v12 forward model to simulate partitioning of of trace elements between melt and residue within the mantle during melting (*91*). REE concentrations along adiabatic melt paths are calculated and misfit between observed and synthetic REE concentrations is minimised using a grid search procedure. Ball *et al.* (*20*) collect their data into 1°×1°geographic bins and calculate an optimum melt path for each bin. The upper limit of the melting region is assumed to represent the lithosphere-asthenosphere boundary (*20*). Note that this definition is slightly different to the isothermal definition used elsewhere in our study. The partitioning of each REE depends upon the bulk source composition, the aluminous phase present in the source, as well pressure and temperature. Ball *et al.* (*20*) filter their database for $8.5 < MgO < 14.5$ wt% to mitigate fractionation effects, and they use ϵNd as a proxy for source composition. Finally, they place the garnet-spinel transition zone at 63–72 km (*92*).

Statistical Analyses

Weighted Median

In order to mitigate skewness and suppress outliers, we calculate the median when averaging data. For ease of analysis, we have divided Earth's surface into 1°×1° bins. To avoid oversampling higher latitudes and biasing median values, we weight each lithospheric thickness datapoint, z_i, by latitude, ϕ_i. To achieve this weighting, we construct a data vector, $\mathbf{z}$ and a weights vector, $\mathbf{w}$, where each weight is given by $\cos\phi_i$. The weighted median is calculated by sorting $\mathbf{z}$ and $\mathbf{w}$ in ascending order by the values of $\mathbf{z}$. The median value is given by z_j, where the index, j, of the sorted data vector is calculated by minimising

$$j = \min_k \left[\sum_{i=1}^{k} w_i z_i > \frac{1}{2} \sum_{i=1}^{n} w_i z_i \right]. \tag{1}$$

For even-length data vectors, we use the upper-weighted median, so that z_j is given by $(z_j + z_{j+1})/2$. We calculate the weighted interquartile range of the distribution by changing the prefactor of $\frac{1}{2}$ in Equation (1) to $\frac{1}{4}$ and $\frac{3}{4}$ for upper and lower quartiles, respectively.

Predicted Oceanic Lithospheric Thickness Distribution

Seamounts can only form on oceanic crust that is the same age, or older, than themselves. For example, a suite of seamounts erupting at the present day (i.e. 0 Ma) can cap oceanic crust that formed between the present day and Jurassic times. The median age of the sea floor beneath these seamounts will therefore lie somewhere between the two, and be skewed towards

younger ages since there is much more young oceanic crust than old on Earth (Figure 1a). Moreover, a seamount that is 200 Ma can only be located on crust ≥ 200 Ma, the majority of which has been subducted and lost from the surface record. Consequently, since lithospheric plate thickness increases with age, we expect to observe a positive relationship between seamount age and LAB depth.

Richards *et al.* (*24, 25*) constructed an oceanic plate-cooling model that simultaneously minimises the misfit between global databases of heat flow and bathymetric observations that have been corrected for sediment and crustal loading. Their approach includes both pressure- and temperature-dependent thermal properties. We assume that the LAB within this plate-cooling model coincides with the 1175 °C isotherm [i.e. consistent with the LAB model of Hoggard *et al.* (*8*)]. We predict the unperturbed depth to the modern-day LAB, z^1_{LAB}, by extracting LAB depth as a function of age from this plate model using the oceanic crustal age model of Muller *et al.* (*28*), which was updated to include additional data by Richards *et al.* (*24*). The predicted value is compared to thickness implied by seismic tomography, which is in close agreement [see Figure 1c; (*8, 85*)].

Oceanic crustal ages, t_{crust}, and LAB depths range from 0 Ma to t_{max} and from $z_{LAB}(0)$ to $z_{LAB}(t_{max})$, respectively. Seamounts cannot form on plates younger than their eruption age, t_0, and if $t_0 > t_{max}$, then they would not be recorded on Earth's surface today. Therefore, the thickness of lithosphere on which seamounts are located ranges from $z_{LAB}(t_0)$ to $z_{LAB}(t_{max})$. For each eruption age, we record the lithospheric thickness distribution of oceanic crustal ages where $t_0 < t < t_{max}$. The median and interquartile range of predicted lithospheric thickness as a function of t_0 (i.e. the blue line in 3d) is calculated using Equation (1).

Lithospheric Modelling

Parameterisation

To model lithospheric re-thickening following periods of active magmatism we solve the one-dimensional heat equation expressed as

$$\rho(P,T,X)C_P(T,X)\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left[k(P,T,X)\frac{\partial T}{\partial z}\right] + H^*(X), \tag{2}$$

where t is time, z is depth, T is temperature, P is pressure, X is composition, ρ is density, C_P is the isobaric specific heat capacity, k is the thermal conductivity, and H^* is the internal radiogenic heat production.

Equation (2) is solved numerically with an unconditionally stable time- and space-centered Crank-Nicholson finite difference scheme and a predictor-corrector step (*93*). Accordingly, Equation (2) is recast as

$$\begin{aligned} &T^{n+1}_j + A\left(-\frac{k^m_{j+\frac{1}{2}}}{\Delta z^m_j}T^{n+1}_{j+1} + \left(\frac{k^m_{j+\frac{1}{2}}}{\Delta z^m_j} + \frac{k^m_{j-\frac{1}{2}}}{\Delta z^m_{j-1}}\right)T^{n+1}_j - \frac{k^m_{j-\frac{1}{2}}}{\Delta z^m_{j-1}}T^{n+1}_{j-1}\right) \\ &= T^n_j + A\left(\frac{k^m_{j+\frac{1}{2}}}{\Delta z^m_j}T^n_{j+1} - \left(\frac{k^m_{j+\frac{1}{2}}}{\Delta z^m_j} + \frac{k^m_{j-\frac{1}{2}}}{\Delta z^m_{j-1}}\right)T^n_j + \frac{k^m_{j-\frac{1}{2}}}{\Delta z^m_{j-1}}T^n_{j-1}\right) + AH^*\left(\Delta z^m_j + \Delta z^m_{j-1}\right), \end{aligned} \tag{3}$$

where

$$A = \frac{\Delta t}{\left(\rho^m_j {C_P}^m_j\left(\Delta z^m_j + \Delta z^m_{j-1}\right)\right)}, \tag{4}$$

and Δt is the time step, Δz is the depth spacing between nodes, and n and j are the time and depth indices, respectively. Equation (3) is solved by tridiagonal elimination (*93*). For the initial predictor phase of each time step, $m = n$, while in the subsequent corrector phase, $m = n + \frac{1}{2}$. We use a Lagrangian reference frame, whereby Δz^m_j is initially set to 1 km (i.e., when $m = 0$) then scales with thermal contraction in subsequent timesteps. These timesteps are calculated using a

Courant-Friedrichs-Lewy condition calculated according to

$$\Delta t = \min_j \left[\frac{\left(\Delta z_j^0\right)^2 \rho_j^0 {C_P}^0_j}{2.2 k_j^0} \right] \sim 5 \text{ kyr}. \tag{5}$$

T^{n+1} typically converges to within a tolerance of 0.001°C after the corrector phase.

Boundary Conditions

All models consist of crustal and mantle layers. Crustal thickness for oceanic and continental regions are set to z_{oc} = 20 km and z_{cc} = 35 km, respectively. In both cases the underlying mantle extends from the Moho to an assumed equilibrium plate thickness, z_p. The initial depth to the LAB, after the cessation of magmatic activity and lithospheric thinning, is assumed to be z^0_{LAB} = 50 km. To account for the possible presence of a mantle plume beneath the plate, initial asthensopheric potential temperature, $(T_p)_0$, can vary between 1333°C, which is assumed to be the temperature of ambient mantle, and 1633°C (i.e. an excess temperature of 300 °C).

Separate parameterisations are used to define the thermophysical properties of the crust and mantle (k, C_P, ρ, and H^*). Crustal radiative thermal conductivity and ρ are determined using the Richards *et al.* (*29*) parameterisation, but the reference density, ρ_0, is dependent on whether oceanic or continental lithosphere is to be modelled (2950 kg m^{-3} and 2700 kg m^{-3}, respectively). In the continental crust, C_P and lattice thermal diffusivity, κ_{lat}, are calculated using the parameterisation of Whittington *et al.* (*44*), and H^* is assumed to be 0.7 μW m^{-3} (*8*). In oceanic crust, we use the Richards *et al.* (*29*) parameterisation for C_P and κ_{lat}, and assume H^* = 0 mW m^{-3}. In both the continental and oceanic mantle, the temperature- and pressure-dependent formulations specified in Richards *et al.* (*29*) are adopted, with radiogenic heat production assumed to be negligible (H^* = 0 μW m^{-3}).

Initial temperature profiles are obtained by combining the parameters outlined above with the equations of McKenzie *et al.* (*94*). In all cases kinematic viscosity, ν, is assumed to be 9×10^{19} m^2 s^{-1}. For each combination of z_{oc} or z_{cc}, and $(T_p)_0$, we find the steady-state geotherm consistent with z^0_{LAB} by iterating through a range of mechanical boundary layer thicknesses (2–60 km). We select the temperature profile with depth to the 1175°C isotherm equal to z^0_{LAB}.

To simulate the waning of a plume-derived heat source through time, we impose an evolving basal boundary condition. At t = 0, $T(z)$ is given by the initial isentrope defined by $(T_p)_0$ below the base of the thermal boundary layer (i.e., the shallowest depth at which the geothermal gradient, $\frac{\partial T}{\partial z}$, drops below 0.5°C km^{-1}). In later timesteps, the depth at which this boundary condition is imposed, z_b, increases according to a prescribed plume sinking rate, v_{plume} = 10 mm yr^{-1}, until the deepest model node (i.e. $z_b = z_p$) is reached, whereupon the basal boundary depth remains fixed. Simultaneously, from t = 0–30 Myr, the temperature applied at the basal boundary decays linearly to that of the ambient mantle isentrope (i.e. T_p = 1333 °C) at the relevant depth. Beyond t = 30 Myr, the basal temperature is assumed to remain equal to that of the ambient mantle isentrope at the appropriate depth.

Optimisation Strategy

We carry out a two-parameter sweep to find the combination of plate thickness, z_p, and initial potential temperature, $(T_p)_0$, that best explains our constraints on $z_{LAB}(t)$. For each $(z_p, (T_p)_0)$ pair we find the initial steady-state thermal structure that yields z^0_{LAB} = 50 km. Note that, if the lithospheric mantle is instantaneously thinned, it is likely that the remaining mantle will not be thermally equilibrated. Nonetheless, since the conductive layer is thin, it will rapidly reach steady state and so the effect of thermal disequilibration is likely to be minor. We then calculate $z_{\text{LAB}}(t)$, for $0 < t < t_{max}$ Ma, where t_{max} = 750 Ma, using Equation (3). To minimise the misfit between tomographically determined LAB depth and predicted LAB depth as a function of age beneath LIPs, we use the trial function

$$\chi = \sqrt{\frac{1}{M}\sum_{i=1}^{M}\left(\frac{z_i^o - z_i^c}{\sigma_i}\right)^2}, \tag{6}$$

where z_i^o and z_i^c are observed and calculated LAB depth, σ_i is the standard deviation of the ith measurement, and M is the number of estimates. We used weighted binned median lithospheric thickness estimates, where

$$\sigma_i = \frac{\sqrt{(\text{IQR}_1^i)^2 + (\text{IQR}_2)^2}}{1.349}, \tag{7}$$

and IQR_1^i is the interquartile range of the ith measurement and $\text{IQR}_2 = 25$ km is the approximate vertical uncertainty of the lithospheric thickness grid. Note that this relationship assumes that errors are normally distributed.

Subsidence

Air-loaded subsidence of the surface for our pressure- and temperature-dependant plate model is given by

$$S(t) = \int_0^{z_p}\left[1 - \frac{\rho(0,z')}{\rho(t,z')}\right]dz', \tag{8}$$

where z' is the Lagrangian depth coordinate that deepens as the column is compressed. The effect of lithospheric depletion is explored by adding $\Delta\rho$ (i.e. $\Delta\rho$ is negative for more depleted lithosphere) to all values of $\rho(t,z')$ that are located at depths, z, within the lithospheric mantle (i.e. $z_{cc} < z < z_{LAB}$). Since many ancient LIPs are located above continental shields, we assume that the pre-LIP lithospheric mantle was in thermal equilibrium, initially cratonic, and depleted by $\Delta\rho = 50$ Mg m^{-3} (Figure 4a). We explore initial geotherms that are out of equilibrium in the Supplementary Material. For simplicity, we assume that the steady-state thermal structure of the lithospheric column before thinning and after re-thickening are identical and given by the results of our thermal evolution modelling (i.e. z_{LAB}^{∞}; Figure 3d). Initial uplift, $U_o = S(0)$, after thinning is calculated using Equation(8), thence $S(t)$ is calculated using a range of values of $\Delta\rho$ to explore the effect of lithospheric melt depletion upon subsidence of the Earth's surface (Figure 4).

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Data Availability Statement

All data needed to evaluate the conclusions in the paper are present in the Supplementary Materials. On acceptance, data will be placed in an open-source repository whose doi will be stated here.

Figure 1: **Distribution of oceanic magmatic provinces.** (a) Oceanic crustal age (*28, 95*). Yellow circles = locations of intraplate magmatic provinces (*20*); (i) & (ii) = locations of Icelandic and Hawaiian hotspots. (b) Oceanic LAB depth (*8*). Coloured polygons = locations of oceanic magmatic provinces (i.e. ocean islands, seamounts and plateaux) coloured by age. (c) Constraints on modern-day LAB depth beneath intraplate magmatic provinces as a function of underlying oceanic crustal age. Red squares = LAB depth determined by inverse modelling of rare earth element compositions of mafic rocks (*20*); green triangles = LAB depth determined by tomographic-derived LAB depth model shown in panel (b) (*8, 85*); yellow stars = LAB depth estimated by finding magmatic equilibration pressure and temperature that corresponds to base of lithosphere (this study, see main text & Methods); (i) & (ii) = age of crust underlying Icelandic and Hawaiian hotspots. (d) Black circles with error bars = modern LAB depth beneath magmatic provinces as a function of province age; moving median at 10 Ma intervals with a ±10 Myr window; error bars = interquartile range; blue line = median expected distribution of LAB depth beneath LIPs if they were distributed randomly in time and space across oceanic lithosphere assuming plate model of Richards *et al.* (*29*) and plate age model of Muller *et al.* (*28*); blue envelope = interquartile range; (i) & (ii) = age of Icelandic and Hawaiian hotspots.

Figure 2: **Global distribution of continental large igneous provinces.** (a) Distribution of continental large igneous provinces coloured by time since their eruption, overlain on map of continental LAB depth (*8*). Scale bars located on axes of panel (b); labels indicate names of LIPs mentioned in the main text. (b) Modern-day LAB depth beneath large igneous provinces as a function of time since eruption. Labels indicate age of LIPs mentioned in main text.

Figure 3: **Lithospheric thermal modelling.** (a) Observed and modelled LAB depth beneath oceanic intraplate magmatic provinces as function of time since eruption. Grey bar = interquartile range of LAB depth distribution; red line = best-fitting modelled LAB depth. $(T_p)_0$ = initial potential temperature; z_p = plate thickness; z^{∞}_{LAB} = asymptotic depth to the LAB as $t \rightarrow \infty$. (b) Misfit between observed and calculated LAB depth as function of plate thickness, z_p, and initial potential temperature, $(T_p)_0$, for oceanic intraplate magmatism (see methods for details). Red circle = misfit minimum; red dashed line = contour marking 1.5× value at misfit minimum. Optimum values given in lower left-hand corner of panel (a). (c) Observed and modelled LAB depth beneath continental LIPs as function of time since eruption. Red bars = Deccan Traps, excluded from the analysis. (d) As panel (b) but for continental LIPs. Optimum values of $(T_P)_0$, z_p and z^{∞}_{LAB} given in lower left-hand corner of panel (c).

Figure 4: **Thermal, depletion and subsidence modeling.** (a) Pair of cartoons showing continental column and temperature profile of precursory lithosphere before LIP emplacement. Lithosphere initially in thermal steady-state. z_{LAB}^{∞} = steady-state lithosphere-asthensphere boundary (LAB) depth, defined by 1175°C isotherm; z_{cc} = crustal thickness; z_p = plate thickness; ρ_{cc} = steady-state crustal density; ρ_m^* = steady-state initial lithospheric mantle density; ρ_a = steady-state asthenospheric density; T_p = ambient mantle potential temperature; BL = base level (i.e. initial elevation of top of lithospheric column). (b) Perturbed (i.e. thinned) lithospheric column during LIP emplacement at $t = 0$, where LAB depth, z_{LAB}^0 = 50 km; U_o = initial uplift driven by thinning of the plate and sub-plate thermal anomaly; ρ_m^0 = initial lithospheric mantle density; ρ_a^0 = initial asthenospheric mantle density; $(T_p)_0$ = initial (i.e. elevated) mantle potential temperature. (c) Subsidence as LAB depth, $z_{LAB}(t)$, deepens from z_{LAB}^0 back to z_{LAB}^{∞}. $\rho_{cc}(t)$ = crustal density as function of time; $\rho_m(t)$ = lithospheric mantle density as function of time; $\rho_a(t)$ = asthenospheric mantle density as function of time; $\Delta\rho$ = difference as result of depletion between precursor, ρ_m^*, and final lithospheric mantle density, ρ_m^{∞}; $S(t)$ = time-dependent net subsidence (i.e. above or below base level, BL), where $S(\infty) = 0$ km if $\Delta\rho = 0$ Mg m^{-3}. (d) Density as function of depth for thickening of undepleted lithospheric mantle coloured by time since eruption (i.e. red to blue through time). (e) As for (d) but with depleted lithospheric mantle (i.e. density reduction relative to ambient mantle, $\Delta\rho = -50$ kg m^{-3}). (f) Air-loaded subsidence as function of time since eruption, $S(t)$, for various values of $\Delta\rho$ after thinning precursory lithosphere from z_{LAB}^{∞} = 200 km to z_{LAB}^0 = 50 km with initial depletion of -50 kg m^{-3}. U_o = 1.76 km is transient initial uplift generated by combined effect of thinning depleted precursory lithosphere and sub-plate thermal anomaly. Note that if re-thickening lithosphere is more depleted than original lithosphere then uplift is generated after steady-state is re-achieved. Dotted line = no net uplift or subsidence. Labelled lines indicate values of $\Delta\rho$ where negative values are density reductions due to increased depletion. (g) Depth to LAB as a function of time since eruption, $z_{LAB}(t)$.

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