

Investigating the high seismic velocities and buoyancy of cratonic roots – the possibility of diamond

Eric Debayle  · Yanick Ricard

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Abstract

High seismic velocity keels extending to depths greater than 200 km underlie the oldest parts of continents, the Precambrian cratons. These keels have probably been formed early in Earth's history, and the preservation of these deep, cold, and highly viscous roots in a convective mantle remains enigmatic. A classical view is that the excess density due to colder temperatures is compensated for by a light composition. Here, we map the magnesium number (Mg#, a proxy for mantle depletion) within cratonic keels, based on the thermochemical interpretation of a global shear velocity model. Our interpretation suggests that depletion is strong above 150 km ($\text{Mg\#} > 92$), and decreases with depth down to the lithosphere-asthenosphere boundary (LAB). However, changing only the major element composition and the Mg# does not fully meet the requirements implied by seismology or gravity observations. We explore various hypotheses, varying the temperature of the lithosphere, considering the presence of eclogite, accounting for chromium or for a minute quantity of diamonds. We also discuss the uncertainties in seismological observations of velocity and radial anisotropy as well as in mineralogical models. Changes in temperature, the presence of eclogite, of Cr incorporated in garnet, or radial anisotropy are inconsistent with shear velocity and gravity constraints when considered individually. Below the graphite/diamond transition, the combination of depletion with a low volume fraction of diamond ($< 1\%$ at 150 km) appears to be the simplest way to explain the very high shear velocities, while maintaining the cratonic lithosphere close to neutral buoyancy. This small amount of diamond accounts for only 0.5% of the mantle's carbon and can be formed within a reasonable geological timescale (> 180 Ma) through the subduction of organic carbon.

Introduction

Our results come in the context of the thermochemical interpretation of a recent S-wave tomographic model of the upper mantle, 3DLGL-TPESvoigt (Fig. S1). This model is based on the waveform inversion of a massive dataset of vertical component Rayleigh waves, recorded since 1976 and regularly updated up to November 2022¹. Rayleigh waves constrain vertically polarized shear waves (Sv waves), and we obtain isotropic velocities by computing the Voigt average of this Sv model and the Sh (horizontally polarized shear waves) model obtained by adding the radial anisotropy of PREM² to our Sv model (see Methods). The originality of our approach lies

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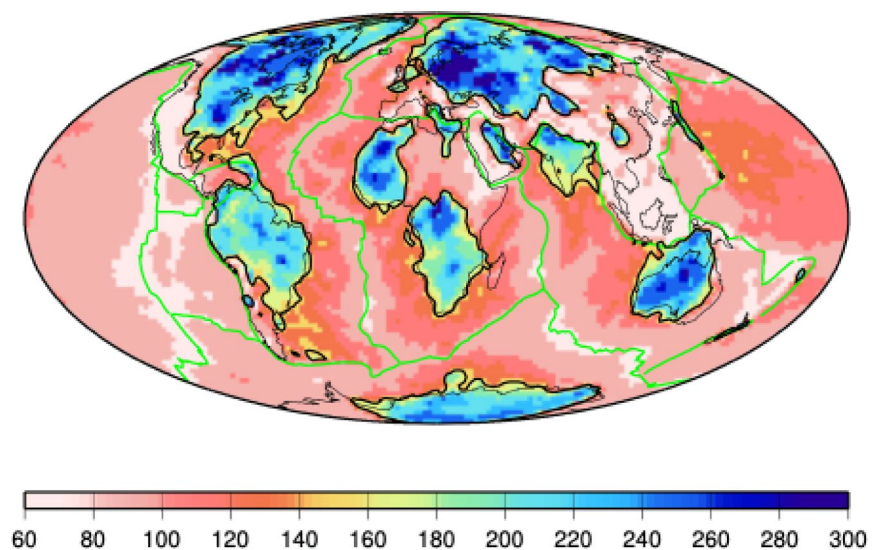
in combining data from the resulting S-wave model with geodynamic constraints on density, in an inversion designed to map petrological parameters relevant to measuring Precambrian root depletion. These parameters are the magnesium number Mg# defined as the molar ratio of Mg/(Mg+Fe), and the modal mass proportion of olivine M_{Ol} . However, depletion alone does not fully explain the velocity and density of the craton keels. The remaining discrepancy is discussed and interpreted in terms of radial anisotropy, eclogite content, Cr included in the garnet phase, presence of diamonds, or a combination of these factors.

Lithosphere-asthenosphere boundary (LAB) and cratonic geotherms

Cratons are coherent blocks of Precambrian lithosphere, stable for long periods of time (several hundreds or thousands of millions of years), and with deep lithospheric keels^{3–7} (>150 km). We first identify the cratons by transposing our tomographic S-wave model to temperature (Fig. S2), using experimental relationships^{8,9} and assuming a pyrolitic composition at their base (see Sect. [Methods](#)). A pyrolitic composition is reasonable for the well-mixed convective mantle beneath the cratons. We assume that the adiabatic gradient in the shallow mantle is $0.57\text{ }^{\circ}\text{C km}^{-1}$ with mantle potential temperature of 1315°C ¹⁰. We estimate the upper limit of the convective mantle, associated with the LAB, when the temperature derived from tomography is closest to the adiabat. We obtain a LAB depth map (Fig. 1), where cratons are associated with parts of the continents resting on a thick lithosphere (>150 km), with an average thickness of around 220 km, but locally reaching up to 375 km. These regions are well associated with Precambrian crustal provinces^{11,12}. In contrast, Phanerozoic regions are thinner, with an average lithosphere thickness of 125 km. In oceanic regions, the lithosphere thickens with sea-floor age, reaching 120 km in the oldest oceanic basins (>170 Myr).

Using our knowledge of the LAB depth beneath the cratons, we compute a geotherm^{11,13} at each geographical location (Fig. S3) and estimate the shear velocity V_s and density ρ along this geotherm using mineralogical phase diagrams computed by the Perple X software¹⁴. The cratonic geotherms used in this study are consistent with temperature and pressure estimates from nearby xenoliths¹⁵. The difference in density between cratons and ridges is bounded using gravity considerations¹⁶. We show in [Methods](#) that, since the geoid above cratons is within 10 m that above ridges, we can conservatively conclude that the lithospheric mantle beneath cratons has a mean density excess of between $0.5\% \pm 0.5\%$ compared with the pyrolitic mantle beneath ridges. Accounting for the thick crust of cratons, an excess density in the cratonic mantle that decreases linearly from 0.80% at 100 km depth to 0% at 200 km explains geoid anomalies and gives a density profile in isostatic equilibrium with that

Fig. 1 Global Lithosphere-Asthenosphere boundary (LAB) depth map. The map is obtained by estimating the upper limit of the convective mantle, when the temperature derived from the tomographic model 3DLGL-TPE. v2022-11¹ is closest to the adiabat. This is done by assuming a surface potential temperature of 1315°C ¹⁰ and an adiabatic gradient in the upper mantle of $0.57\text{ }^{\circ}\text{C km}^{-1}$. Colour scale indicates LAB depths in km. Cratons are associated with parts of the continents that rest on a thick lithosphere (>150 km) and are highlighted with the black contour lines. Green lines mark plate boundaries.



under a ridge (Density 2 profile in Table S1). We therefore compare the density predicted for a pyrolytic composition with that required to explain the geoid and maintain a relative isostatic equilibrium with a ridge (see Fig. S4). We also compare in this figure the Vs predictions for a pyrolytic composition with the seismic velocity according to our tomographic model.

Inversion for petrological parameters

The tomographic velocities are faster than those predicted from pyrolite under the cratons and the densities are larger than those required by isostasy and gravity (Fig. S4). On average, densities are 1.1% too large at 100 km, decreasing to 0.4% at 150 km and 0.1% at 200 km. Tomographic shear velocities are on average faster than those expected for a pyrolite by 2.3% at 150 km decreasing to 1.2% at 200 km and 0.9% at 250 km. Locally, the velocity differences at 150 km exceed 3.0% over large parts of the cratons beneath Australia, Antarctica, India North America and Greenland. Reducing these differences requires cratonic roots that are more depleted than pyrolite, and/or the presence of other mineral assemblages¹⁷.

We parameterize the cratonic mineralogy using three parameters, the Magnesium and Silicium numbers (Mg# and Si#), and the proportion of minor oxides R# (see Sect. Methods). We also predicted ρ and Vs for a harzburgite composition (Table S2) and found that while harzburgite is within a reasonable density range, it is far too slow - even slower than pyrolite -, compared with tomographic velocities (Table S3). We identify compositions that would correspond to the tomographic seismic velocities at 125 km depth and the density required to achieve isostatic equilibrium with a ridge, by computing the probability density functions of Mg#, Si# and R# (Fig. S5). In the case of a 220 km thick lithosphere, neither harzburgite or pyrolite are in the most favorable zone (Mg# $\approx$ 92, Si# < 37 and R# > 75). We therefore invert for composition, using a faster least square inversion at each location and depth, under cratons.

The least-square inversion (see Methods) minimizes at each depth and geographical point the differences in velocity and density displayed in Fig. S4. The phase content (such as orthopyroxene, clinopyroxene, spinel, garnet or olivine proportion), and their composition are consistently deduced from the elemental composition using Gibbs minimization with Perple X. Figure 2 displays maps of Mg# and the modal mass proportion of olivine within the cratonic lithosphere, M_{Ol} . The resolution of these maps is limited by the horizontal (600 km) and vertical (50 km) resolution of our seismic model, so that sharp contrast in the variations of Mg# and M_{Ol} cannot be resolved. The largest partial derivative of the misfit is with respect to Mg#, which is the parameter that is best resolved. R# is associated with small partial derivatives, at least for a pyrolite composition (Table S3). However, Fig. S5 indicates that the proportion of minor oxides, R#, cannot be too low as is the case with harzburgite. The minor oxides are necessary to stabilize the pyroxenes. We therefore use pyrolite as the starting model of our inversion and keep R# around the value appropriate for pyrolite, R# = 85.

At depths shallower than 150 km, Mg# values of between 92 and 94 are found inside the cratons and decrease toward 90–91 near their periphery (Fig. 2, see also Fig. S6 that shows the a posteriori errors on Mg#, relatively constant and close to 1). These high Mg# values, with an average close to 92.5 at 150 km depth (Fig. 3), are consistent with petrological data of mantle xenoliths and indicate that the shallow cratonic mantle is highly depleted, and close to a monomineralic olivine residue resulting from mantle melting to the point of orthopyroxene exhaustion¹⁸. At 175–200 km, average Mg# decreases toward 91 and down to 90 at 250 km depth (Fig. 3), suggesting a lower level of depletion. Although the M_{Ol} maps in Fig. 2 are less resolved than the Mg# maps, they give an overall indication of the lateral variations of this parameter. The highest M_{Ol} values ($M_{Ol} \approx 90\%$) are obtained at 150 km depth in the center part of cratons. Lower values near 80% at 100 km depth suggest Si addition, probably due to pyroxene M_{Ol} enrichment (Fig. 3). Deeper than 150 km, the average modal olivine content also decreases until the base of the lithosphere (Fig. 3). Despite the lower resolution of M_{Ol} , Mg# and M_{Ol} values are consistent with petrological results for mantle peridotites^{18,19} in the shallow upper mantle. They suggest a strong depletion in the upper parts of cratonic roots, that probably occurred at the time of craton formation, followed with a

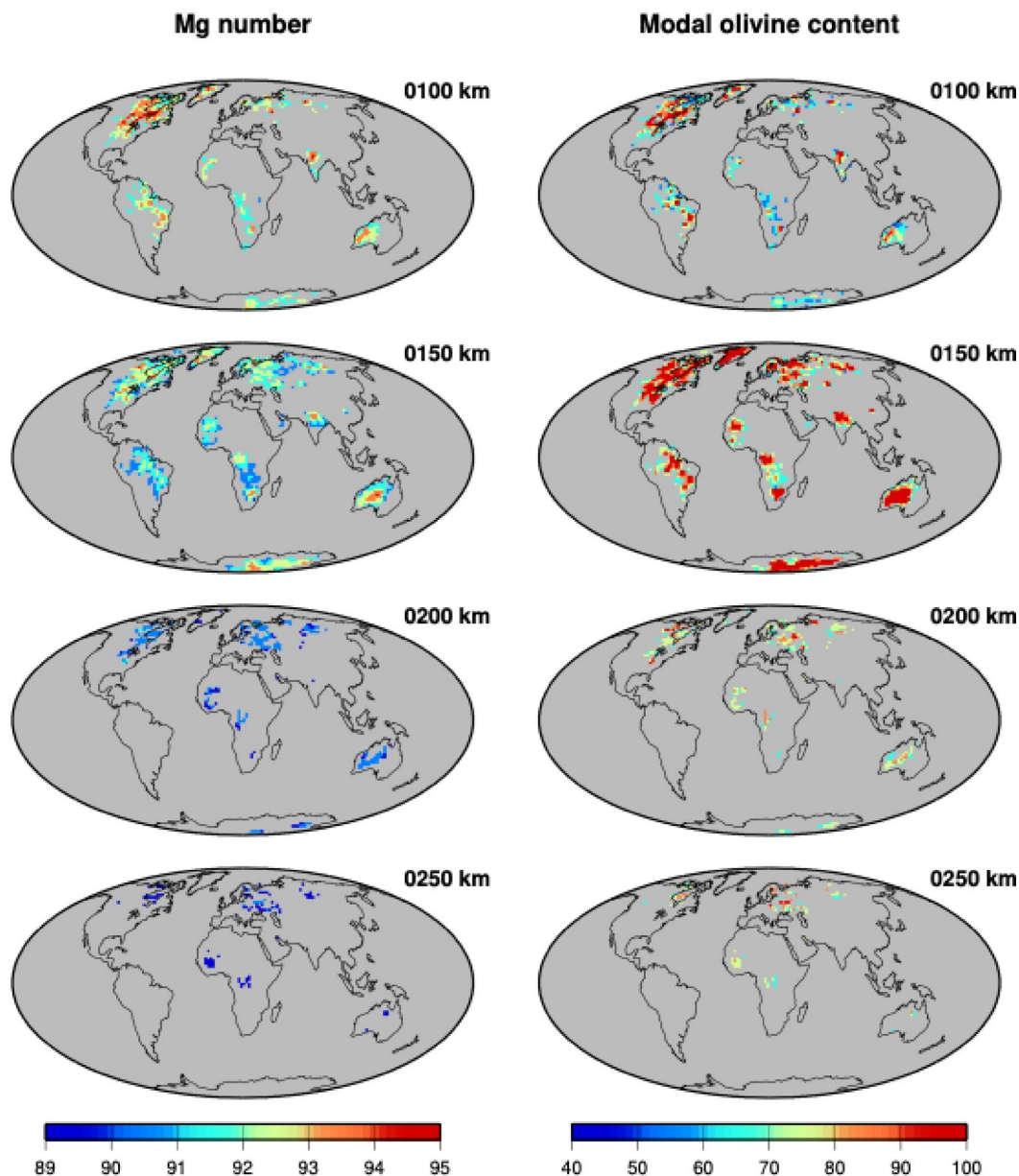


Fig. 2 Maps of Mg# number (left column) and modal olivine content M_{Ol} (right column) in the cratonic lithosphere. These maps are obtained after a least-square inversion assuming a pyrolytic starting model. M_{Ol} is derived from the Si# parameter (see [Methods](#)). Cratons are associated with parts of the continents resting on a thick lithosphere (> 150 km) and are isolated from the black contour lines of Fig. 1. Values are displayed only in the cratonic lithosphere. In the asthenosphere, our seismic models can be explained by a pyrolytic composition, assuming a small amount of melt in oceanic regions⁸.

potential pyroxene enrichment (Si addition and decrease in M_{Ol}) at depths shallower than 150 km. In the cratonic keels deeper than 150 km, depletion decreases and composition tends toward that of a pyrolite.

A remaining misfit

Figure S7 shows the percentage mismatch for density and shear velocity after the least square inversion. At all depths, the inversion predicts densities within 0.5% of that required by isostasy and gravity, providing rough confirmation of the isopycnicity hypothesis necessary for the long-term stability of cratonic roots³. However,

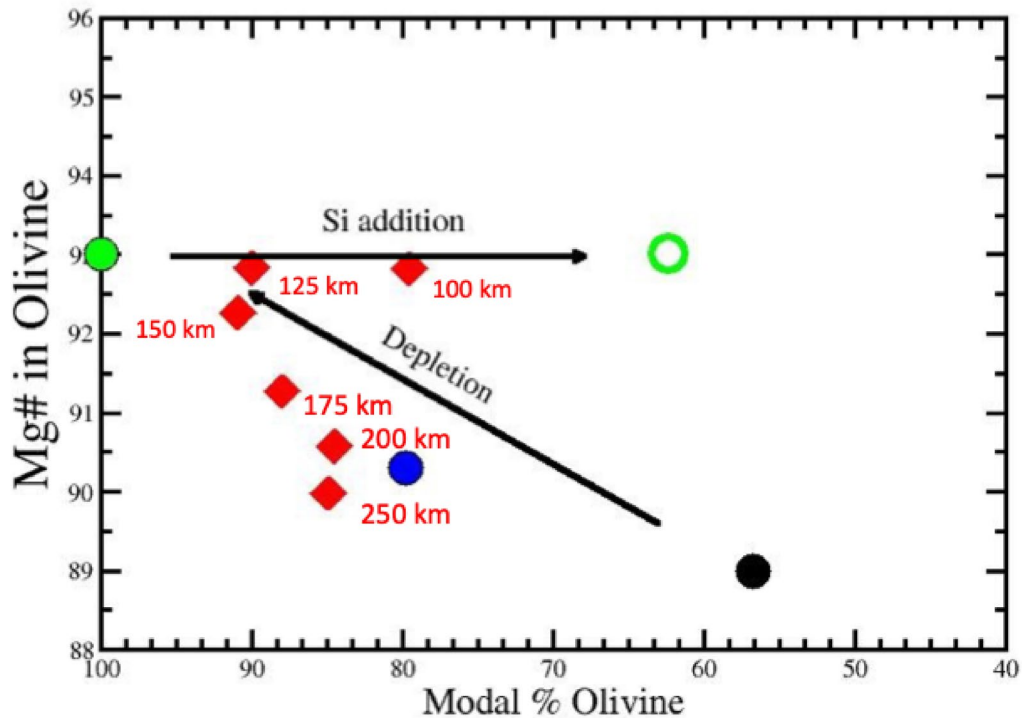


Fig. 3 Mean values of Mg# number and modal olivine content M_{Ol} in the cratonic lithosphere. Red diamonds indicate mean cratonic values of the maps in Fig. 2 for different depths. Black and blue dots indicate pyrolite and harzburgite composition values⁵¹ respectively. The full green dot corresponds to a pure olivine with a Mg# number of 93, while the open green dot shows the effect of Si addition to this pure olivine.

although our inversion reduces the velocity differences observed in Fig. S4 by at least 60% at all depths, it does not fully explain the high seismic velocities inferred from tomography in some parts of the cratonic roots. The remaining velocity differences are at their maximum at 150 km depths, where the mineralogical velocities remain slower than those suggested by seismology by between 1% and 3.5%, beneath Australia India, South Africa, West Africa, Amazonia, Sao Francisco, Antarctica, North America, Greenland and the Siberian cratons. This residual misfit implies that other possibilities should be considered to explain the properties of the fastest cratons.

Figure 4 summarizes how the V_s and ρ misfit between seismic observations/gravity constraints and predictions can be explained. The origin of the graph, is the prediction for pyrolite at 150 km depth under a cold craton, the green circle shows the observation requirements, i.e., a velocity 2.5% faster and a density 1.25% lighter. Figure 4a shows that increasing Mg# has a greater effect on ρ than on V_s , while fitting the data would require a greater effect on V_s than on ρ . Changing Si# or R# has a smaller effect that goes in the wrong direction. Other possibilities to consider for modifying velocity and density include adding eclogite, Cr or decreasing temperature. However, Fig. 4 shows that they do not change the system in the favourable way since an increase in velocity is accompanied by a significant increase in density. A weak or negative radial anisotropy at 150 km depth could also reduce the apparent velocity of the cratonic roots (cyan circle). These possibilities are discussed in the next section. Diamond is the only candidate that can even in small quantities, would increase the velocity with no significant effect on density.

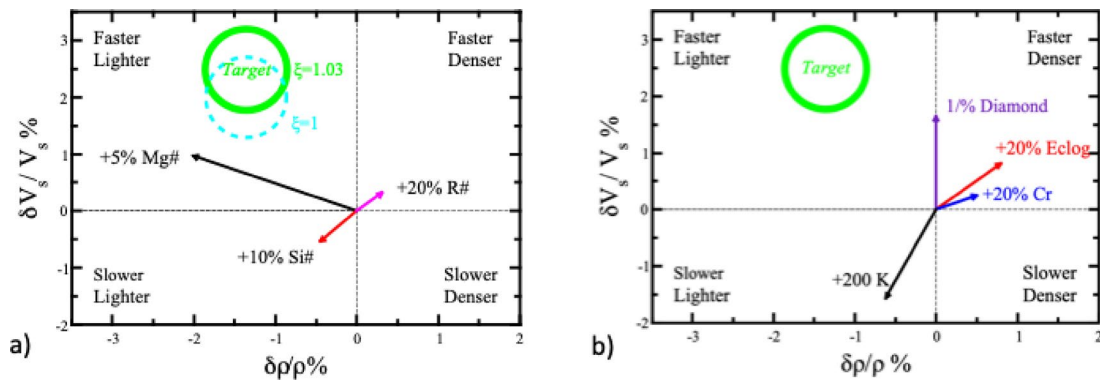


Fig. 4 Cartoons illustrating the sensibility of velocity and density to various variations of the reference model. The target (green circle) is to obtain near 150 km depth, a cratonic lithosphere faster and lighter than a pyrolite (arbitrary at the origin). This requires to increase V_s by about 2.5% and to decrease ρ by about 1.4% (Table S2). The arrows are proportional to the sensitivities of the different parameters with respect to V_s and ρ . Panel (a) shows the effect of composition. Increasing Mg# (by 5%) has a greater effect on ρ than on V_s , while reaching the target would require a greater effect on V_s than on ρ . Changing Si# or R# (by 10 and 20%) has a smaller effect but in the wrong direction (faster but denser or slower but lighter). A reduction of radial anisotropy ($\xi = 1.03$ in PREM) would displace the target (cyan circle with $\xi = 1$) but is not compatible with seismic observations. Panel (b) shows that the effect of eclogite, Cr and temperature cannot align with the target. Diamond is the only candidate that can increase the velocity with a very small effect on density.

The effect of radial anisotropy

One way of reducing the misfit is to adjust the level of radial anisotropy, to lower the overall tomographic velocities based on vertically polarized S waves (see Methods). However, this would require V_{sh} to be significantly lower than V_{sv} , ($V_{sh} < V_{sv}$ or $\xi = (V_{sh}/V_{sv})^2 < 1$) beneath most cratons (Fig. S8d).

A $\xi < 1$ contradicts the $V_{sh} > V_{sv}$ radial anisotropy of 1D models like PREM or the recent REM1D²⁰ as well as the radial anisotropy present in some recent 3D tomographic models^{21–24}, all with $\xi \sim 1.05$ at 150 km depth. A recent model²⁵, however, proposed that little radial anisotropy is present deeper than 150 km depth beneath several locations above cratons. This model confines the anisotropy in a thin layer between 50 and 150 km depth, with extreme values ($\xi > 1.2$). At any rate, we do not know of seismic observations that would lead to anisotropy as low as 0.95, at 150 km depth that would explain the velocity beneath cratons (see Fig. S8d). In Fig. 4a, a cyan circle shows how the velocity misfit would be reduced if no radial anisotropy is present (i.e., $\xi = 1$).

The presence of eclogite

It has been shown that eclogite provide S-wave velocities exceeding seismic velocities inferred from tomography¹⁷. In order to consider the presence of eclogite in the depleted craton keels, it is necessary to assume that the two phases are unmixed and not in thermodynamic equilibrium. Adding eclogite increases density and velocity in similar proportions (Figs. 4b, S9). We show in Fig. S10 that explaining tomographic S-wave velocities requires unrealistic amounts of eclogite above 200 km (up to 94% at 150 km and 43% at 175 km). In addition, the corresponding increase in density (up to 4.7% at 150 km and 2% at 175 km) would far exceed the requirements of gravity observations. Fig. S11 shows the maximum amount of eclogite that can be incorporated for increasing density by no more 1% relative to a density profile in isostatic equilibrium with that under a ridge. Although at a depth of 150 km, the necessary amount of eclogite reaches 40%, exceeding previous estimates of between 2% and up to 20% eclogite in the cratonic mantle^{17,26–28}, a significant V_s misfit remains, in some locations larger than 2%.

The role of chromium

The number of oxides used in the mineralogical modelling is limited, and one might wonder whether adding some minor oxides could explain the velocities and densities. For instance, chromium oxide (Cr_2O_3) has not been considered in the database, even though it is present in the subcontinental lithospheric mantle, where it is the most abundant oxide after those considered^{29,30}. Chromium mainly replaces aluminum in the octahedral sites of garnet, which is stable at 150 km depth. Since chromium is twice heavier than aluminum, it increases the density of garnet. In the peridotite, dominated by the olivine phase, 20 wt% Cr increases the density by 0.3%³¹. The presence of chromium also increases the rigidity. At low Cr concentrations, the increase in rigidity appears to be larger than the increase in density and the Vs velocity could increase by up to 0.5%. However, with higher Cr concentrations, the density increase exceeds the rigidity increase which reduces the S-waves velocity^{32–34}. Therefore, in the best-case scenario, the presence of chromium might help explaining the fast velocity of cratons, but would not be in agreement with the gravity observations (case shown in Fig. 4). In the worst case, a larger concentration would be opposite of what we are looking for, decreasing the velocity and increasing the density.

The possibility of diamond

The occurrence of diamonds in the cratonic mantle is limited by the depth at which the transition from graphite to diamond occurs, which increases with temperature³⁵. If a small graphite volume fraction is present above this transition, its low Vs (about 4.0 km/s) represents a negligible decrease in velocity. However, below the transition, diamond has Vs above 12 km/s and is less dense than pyrolite (Fig. S9). Therefore, the addition of a small volume fraction of diamond (<1%) should significantly increase velocity, while having a negligible effect on density. Figure S12 shows the depth of the graphite to diamond transition for the temperatures derived from our seismic model as well as the thickness of the potential diamond layer. This is obtained by subtracting the depth of the graphite to diamond transition to the LAB depth. Beneath craton cores, which are associated with the coldest geotherms, the deepest lithosphere and the thickest potential diamond layer, the transition is shallower than 150 km and diamonds are stable at this depth. We show in Fig. 5 that adding small volume fractions of diamond (see Methods) explains high seismic velocities beneath cratons, without affecting the density and the approximate isopycnicity condition. At 150 km depth, these volume fractions only exceed 1% in small areas (blue regions in Fig. 5a) of central Australia, South Africa, north India, Greenland and north America. These regions represent 3.6% of the surface of the cratons and the diamond volume fraction can reach 1.95%. Elsewhere, the diamond volume fraction is less than 1%, dropping to below 0.2% over 54% of the surface of the cratons (white regions in Fig. 5a). The volume fraction of diamond stays below 0.2% over 81% of the surface of the cratons at 175 km, and vanishes at greater depths. These volume fractions are likely to be upper limits, since a small volume of eclogite, of chromium, a radial anisotropy lower than in PREM, or the effect of uncertainties, could also contribute to the high velocities. These values are however up to two orders of magnitude higher than those generally observed in kimberlite xenoliths³⁶, and have only been observed in rare diamond-rich “seams” xenoliths^{37–39}.

The effect of decreasing the temperature

Cooler temperatures result in faster predicted velocities that will reduce the Vs misfit, although it will also increase the density. We show in Fig. S13 the result of a simulation where we have increased the lithosphere thickness by 50 km beneath cratons. This produces continental roots with thicknesses reaching 425 km instead of 375 km in Fig. 1. This appears extreme as cratonic keels never reach 400 km in recent seismological models^{1,40,41}. Roots thicker than 300 km would also be difficult to maintain in a convective mantle and appear in large contradiction with gravimetric observations⁴². Fig S13 shows that using unrealistic thick roots reduces, but does not suppress the need of diamond presence at 150 km depth.

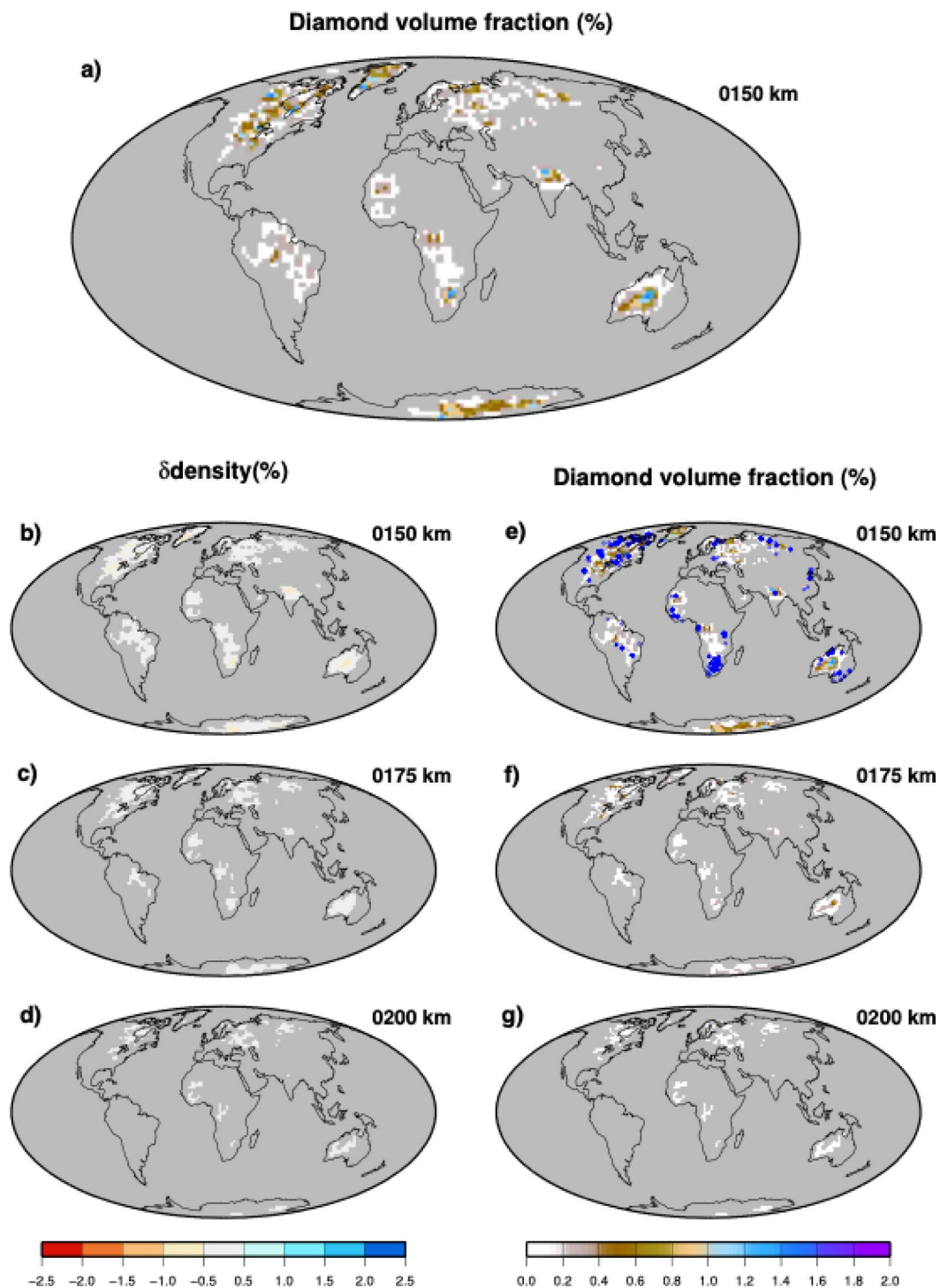


Fig. 5 Explaining data with the diamond hypothesis : density anomalies and diamond volume fraction at different depths in cratonic lithospheric roots. Blue squares on the map in panel (e) indicate where diamonds are found at the surface⁵⁸. Volume fractions of diamonds are indicated with colours from white ($<0.2\%$) to brown ($<1\%$) and up to purple ($<2\%$). At 150 km (panel a), maximum values up to 1.95% are locally observed, but the volume fraction generally stays below 1% and even 0.2%. Below 150 km (panels f–g), the amount of diamond is less than 0.2%. Density differences are in percent with respect to the density required to explain the geoid and achieve isostatic equilibrium with a ridge.

Uncertainties of the tomography

Uncertainties on the observed and predicted shear velocities, on the predicted LAB thickness and on anisotropy, can mitigate the need of diamond. Using our tomographic approach, we computed V_s error maps beneath various cratons and found that these errors are smaller than 1% in the uppermost 300 km^{43–47}. For 1% error and gaussian uncertainties that are classical in linearized travel-time tomography, the probability that the actual V_s is reduced by 1% is around 0.15. As shown in Fig. S14, diamond is less widespread in this case, and its maximum amount is reduced to 1.3% at 150 km.

Uncertainties in the mineralogical predictions

The mineralogic properties are predicted using the Perple X software¹⁴. Perple X predicts the most stable assemblage in a given bulk-rock composition and P-T conditions. Other codes exist^{48–50}, with different strategies for minimising the free energy, but the differences in their predictions is mostly due to different databases (i.e. the thermodynamic properties of the phase end-members). We tested different database^{51–54} without finding significant differences in the results. The figures of this paper were computed with the database used in⁵¹. We show in Fig. S15 that a different database^{53,54} produces very little changes in the amount of diamond that can be estimated.

Results and conclusions

Our results support the idea that cratonic roots consist of highly depleted harzburgite/dunite ($Mg\# \approx 92.5$) down to 150 km, underlain by more fertile compositions. This depletion produces cratonic roots that are almost neutrally buoyant. This explains their long-term stability. However, depletion alone cannot explain locally their high seismic velocities. This is hardly a novelty for petrologists but it is the first time that depletion maps are derived from seismology.

However, changing only the major elements does not fully explain the observations for all cratons, even when considering the uncertainties of tomographic and mineralogical models. We have tested various hypotheses that could account for this discrepancy. As shown in Fig. 4, the changes in prediction induced by these additions to our simple model do not align with the observations. Changing the temperature or the lithospheric thickness, or considering the presence of chromium and of eclogite, improves the predicted velocities but deteriorates the predicted densities, or vice-versa. Seismic radial anisotropy where horizontally polarized S-waves are slower than vertically polarized S-waves, would decrease the velocity misfit, but this seems to contradict a majority of observations. Due to their extremely large velocities, the presence of diamonds, around or below the 1% volume, could eliminate the velocity mismatch without affecting the density balance. This hypothesis may appear improbable. Alternatively, one could conclude that tomographic velocities and radial anisotropy of cratonic roots are over-estimated. In addition, depletion would be so strong that the presence of Cr or of eclogites would be required to increase the density of a lithosphere consisting of pure forsterite. The combined effect of temperature, anisotropy, presence of Cr, eclogite and the uncertainties of the thermodynamic database might provide a solution that does not necessarily appear more probable.

Although our approach is hopefully more systematic, other authors have shown that a depleted composition is necessary to explain the velocity and/or density of cratonic roots^{3,15,17,55}. Some of these authors have proposed that depletion alone can explain their observations^{15,55}. However, these models neglect radial anisotropy, and generally propose a rather low SV shear velocity for the cratonic roots (~ 4.65 km/s) (see Fig. S26 or S27 of ref⁵⁵, or Fig. S7 of ref¹⁵). In contrast, our tomographic model and those of^{22–24,40,56} predict much higher velocity (~ 4.8 km/s). These latter models account for radial anisotropy that further increases the values of the isotropic velocities. The $\sim 1.5\%$ excess velocity of the cratonic lithosphere in the latter models compared to the former (and locally more) is why we cannot explain the high seismic velocities and neutral buoyancy of the cratonic

keels locally, in agreement with¹⁷. We strongly believe that these high velocity values are indeed required by the surface wave observations.

If the hypothesis of diamond is correct, only a small part of this diamond reservoir is brought to the surface by kimberlite magmas (Fig. 5). Such a cratonic mantle reservoir is estimated to contain around 10^{19} kg of diamonds. Although such a quantity is certainly impressive, the mantle is known to contain about 3×10^{21} kg of carbon⁵⁷, meaning that 0.5% of carbon would be in the form of diamond. This carbon was either present in the primitive Earth or was emplaced over reasonable geologic timescales by organic carbon subduction (> 180 Myr)¹⁷.

Methods

Seismic tomographic model

In Debayle et al.⁸, we simultaneously used the attenuations^{59,60} and velocities^{1,6} to discuss the mineralogical properties of the asthenosphere. This was appropriate for the mantle under oceans or young continents where the variations of attenuation are meaningful. Under cratons however, the attenuation is very low⁶⁰ (the quality factor Q exceeds 600 and is often larger than 1000) and has a very small effect on the amplitude of surface waves. This means that it is impossible to determine whether Q is equal to 1000 or much greater, at least using surface wave data. Therefore, under cratons we cannot use two independent parameters, Q and V_s , since only V_s is quantitatively meaningful.

We use our recent V_s upper mantle seismic velocity model, 3DLGL-TPESv⁸, version 2022-11, constrained by a massive surface wave dataset and regularly updated with new available data. The version we use is based on the waveform analysis of 2,008,820 Rayleigh waves seismograms and has been updated to November 2022. Its construction is well documented^{1,6} and it has been shown that it achieves a lateral resolution of ~ 600 km, and a vertical resolution of ~ 50 km. This model is consistent with other recent models up to degree 60 in the uppermost 200 km of the mantle¹. Since Rayleigh waves constrain vertically polarized shear waves (S_v waves), we obtain isotropic velocities by computing the Voigt average of this S_v model and the S_h (horizontally polarized shear waves) model obtained by adding the radial anisotropy $\xi = (V_{sh}/V_{sv})^2$ of PREM⁶¹ to our S_v model. The obtained isotropic model is shown in Fig. S1 and hereafter referred as 3DLGL-TPESvoigt.

Building a LAB depth map

Our V_s seismic model is interpreted in terms of temperature using a model derived from experimental results⁹. These experimental results are valid at temperatures larger than 1000°C and can be applied to the asthenosphere. Using the parameters listed in Debayle et al.⁸ and assuming a pyrolitic composition⁵¹ in the asthenosphere, we can relate the seismic velocity of the tomographic model to the temperature. We obtain temperature maps Fig. S2. We define the LAB (Lithosphere-Asthenosphere Boundary) as the depth where the asthenosphere temperature reaches the adiabat (1315°C at the surface with an adiabatic gradient of $0.57^\circ\text{C km}^{-1}$). The resulting LAB depth map is shown in Fig. 1.

Isostasy and Geoid

Isostatic equilibrium, which is valid for wavelengths greater than the elastic length of the lithosphere⁶², indicates that the integrated density calculated vertically across cratons cannot be very different from that calculated beneath an oceanic ridge. In other words, it is similar to the density calculated along an adiabatic mantle covered by 3 km of water

$$\int_0^L \delta \rho_{c-r}(z) dz \approx 0,$$

(z is depth, $\delta \rho_{c-r}(z)$ is the density difference, at depth z , between craton and ridge, L is larger than the lithospheric thickness). As the geotherm under the cratons is much colder than the adiabat (the geotherm under ridges), this led to the conclusion that chemical depletion occurred in the continental roots to reduce their density and balance their thermal contraction³.

Another strong constraint on craton density comes from the height of the geoid. For the shallow lithospheric anomalies considered here, the geoid anomaly approximately follows the moment law derived from isostasy⁶³:

$$\delta N = -\frac{2\pi G}{g} \int_0^z \delta \rho_{c-r}(z) z dz$$

It was noted very early on that the height of the geoid above cratons is comparable to that over young oceanic lithosphere, $\delta N_{c-r} \approx 0$. More precisely the geoid over cratons is only around 5–10 m below that over ridges⁶⁴ (to be compared with the overall geoid variations of order 250 m). This again indicates a weak excess density of the cratonic lithosphere. The z -term in the geoid moment law suggests furthermore that the excess density should be located at shallow depth.

Although the isostatic and geoid rules are only approximate, they constrain the possible density of cratons quite tightly. In Table S1, we compare a density profile under a ridge, with adiabatic temperatures, to two possible density profiles under cratons. All these profiles are in exact isostatic equilibrium (i.e., their lithostatic pressures at 200 km depth are the same). The crustal densities are in agreement with⁶⁵. To compensate for the light crust of cratons, their lithosphere is, on average, denser and we consider two cases, one (Density 1), where isostasy is obtained by assuming a uniform excess density in the mantle beneath the cratons of 22.2 kg.m^{-3} , down to 200 km, the other (Density 2) where the excess density beneath cratons decreases linearly with depth and cancels out at 200 km depth. In the second case, the density in the mantle below the continental Moho is 44.5 kg.m^{-3} larger than the adiabatic density. Only the Density 2 profile leads to a geoid anomaly of -9.7 m which is acceptable (although already large). We can therefore consider that the mean excess density of the craton keel is only of $0.5\% \pm 0.5\%$, significantly lower than the value predicted for pyrolite, as discussed below. In this paper we assume that the excess density of cratons follows the Density 2 profile and decreases linearly from 0.80% at 100 km depth to the adiabatic density at 200 km.

Mineralogical considerations

In order to compute density and velocity, we choose a simple model for the lithospheric mineralogy consisting of a mixture of 6 oxides MgO , FeO , SiO_2 , Al_2O_3 , CaO and Na_2O . We will further simplify the problem by considering that the molar proportion of the different oxides only depends on three parameters:

a bulk Mg# number

$$\text{Mg\#} = 100 \cdot \frac{\text{MgO}}{\text{MgO} + \text{FeO}},$$

the Si# number

$$\text{Si\#} = 100 \cdot \frac{\text{SiO}_2}{\text{SiO}_2 + \text{MgO} + \text{FeO}},$$

and the ratio

$$\text{R\#} = 100 \cdot \frac{\text{Al}_2\text{O}_3 + \text{CaO} + \text{Na}_2\text{O}}{\text{FeO}}.$$

These quantities constrain the composition to be

$$Mg0 = \frac{Mg\#(1 - Si\#)}{1 + R\#(1 - Mg\#)(1 - Si\#)} ,$$

$$Fe0 = \frac{(1 - Mg\#)(1 - Si\#)}{1 + R\#(1 - Mg\#)(1 - Si\#)} ,$$

$$SiO_2 = \frac{Si\#}{1 + R\#(1 - Mg\#)(1 - Si\#)} .$$

The total proportion of the minor oxides is

$$Al_2O_3 + Na_2O + CaO = \frac{R\#(1 - Mg\#)(1 - Si\#)}{1 + R\#(1 - Mg\#)(1 - Si\#)} ,$$

and each minor oxide is kept in the same proportions as in pyrolite⁵¹. This assumes that the incompatibility of these three oxides is comparable. This is a rough approximation, but the effect of these minor components on the resulting density and velocity is not crucial. The ratio $R\#$ in a melt decreases with pressure, so that a higher $R\#$ in the cratonic lithosphere should indicate that this lithosphere is the residue of melt extraction that occurred at a greater depth⁶⁶ than under a ridge.

The typical value of the $Mg\#$ is around 89 in pyrolite. Neglecting the minor oxides, $Si\#$ varies between 1/3 (2 cations 2+ per silicium in olivine) to 1/2 (1 cation 2+ per silicium in pyroxene). The $R\#$ is 85 in pyrolite, 22 in harzburgite. In Table S2, we report the values of the three parameters and the resulting composition for pyrolite and harzburgite, identical to those used in⁵¹. In the last two columns we also report the Mg number in olivine, $Mg\#_{ol}$ and the weight proportion of olivine, Mod_{ol} . Bulk $Mg\#$ and $Mg\#_{ol}$ are close together as the partitioning of Fe between olivine and pyroxene is ≈ 1 . The molar proportion of olivine is close to $M_{ol} = 3 - 6Si\#$ since, neglecting Al , Ca , and Na , only olivine can be formed if $Si\# = 1/3$ and no olivine can be formed if $Si\# = 1/2$. As the olivine mass is larger than that of pyroxene ($\approx 147 \text{ g.mol}^{-1}$ versus $\approx 103 \text{ g mol}^{-1}$), the weight proportion of olivine is somewhat larger than its molar proportion.

For each composition, we then calculate the density and S-wave velocity using *Perple X*¹⁴. In Table S3, we show the values found at a depth of 125 km for a 200 km thick craton, using the temperature from Fig. S2. The pyrolite is far too dense (1.36% denser than adiabat, second column) while harzburgite is within a reasonable density range (0.05% denser). Both are too slow, and harzburgite is even slower than pyrolite (in the third column we give the excess velocities compared with adiabat, in the fourth column, the deficit velocity compared with seismologic velocities). None of these compositions is therefore acceptable for a cratonic root.

To understand how we can explain the density and S-wave velocity of cratons by changing the composition, we show in Table S3 the derivatives of these quantities when our three mineralogical parameters are varied. For a better understanding, these values are also plotted in Fig. 4a. We have chosen the density and velocity calculated with pyrolite as the origin. We want to modify the composition so that the density is lower and the velocity faster. Clearly, increasing the $Mg\#$ would move our predictions in the right direction. However, increasing the $Mg\#$ content tends to decrease the density more than it increases the seismic velocity. This must be counterbalanced by decreasing $Si\#$ (and as a minor effect increasing $R\#$) that will still increase the velocity but moderately decrease the density. The same reasoning applies whether we compute the derivative starting from a pyrolitic or a harzburgitic composition (dashed arrows); harzburgite having already a higher $Mg\#$ and lower $Si\#$ than pyrolite, the $Mg\#$ derivatives are somewhat better in increasing the velocity without decreasing the density too much.

To better identify the mineralogical composition that would be acceptable, we can compute the probability density function (PDF) of the three mineralogical parameters. Assuming that the a priori distribution of these parameters is uniform over [89, 94] for $Mg\#$, over [34, 42] for $Si\#$, and over [0, 100] for $R\#$, their PDF is easily computed from their likelihood ψ that we choose to be:

$$\psi = \exp \left(-\frac{(\delta \rho_{c-r} - \delta \rho_0)^2}{2a^2} - \frac{(\delta V_{c-p}^s - \delta V_{c-T}^s)^2}{2b^2} \right)$$

$\delta \rho_0 = 0.6\%$, where $\delta \rho_0$ is the excess density targeted at 125 km depth to keep a craton in isostatic equilibrium with a ridge. $\delta V_{c-T}^s = 2.49\%$, where δV_{c-T}^s is the expected shear velocity contrast at 125 km depth between tomographic seismic velocities and prediction along a cratonic geotherm for a pyrolytic composition (Table S3). The difference between the density beneath a craton with its specific composition and geotherm, and the density of an adiabatic pyrolite, is $\delta \rho_{c-r}$. The difference between the S velocity for a given cratonic composition and the velocity of a pyrolite, both along a craton geotherm is V_{c-p}^s . Parameters a and b are both fixed at 0.5%. The resulting PDFs are shown in the 9 panels of Fig. S5, again at a depth of 125 km. The PDFs for each variable are depicted along the diagonal panels (1D marginals), while the trade-off between the parameters (2D marginals) is shown in the other panels (symmetrically). An appropriate mineralogy should be close to the maximum of the probability of each parameter and in the red zones of the 2D marginals. The red and blue dots represent the values for pyrolite and harzburgite: although harzburgite appears somewhat better than pyrolite, neither of these compositions is located in the most favorable zone (in red). A favorable composition has a higher Mg# and less silicium. Its R# remains quite high; this suggests that the cratonic lithosphere is a residue of melting that occurred at larger depth than under a ridge.

Least-square inversion for composition

The data of our inversion are the seismic velocities from our tomographic model¹ (Fig. S1) and the Density 2 profile (Table S1), which is in isostatic equilibrium with that under mid ocean ridge. Therefore, the inversion aims to explain seismic velocities by respecting isostatic equilibrium between ridges and cratons, with a small excess density in the mantle, compatible with long term craton stability. We minimize the differences between these data and the velocity and density predicted for a pyrolytic (Fig. S4) composition⁵¹ along cratonic geotherms computed at each geographical location using a least square inversion⁶⁷. We parametrize composition using the magnesium number Mg# and the “silicium number” Si# as defined in the previous section. The R# parameter, is associated with small partial derivatives (see Table S3), and has been fixed to the pyrolite value. Knowing Si#, it is straightforward to retrieve the modal mass proportion of olivine, M_{Ol} close to $M_{Ol} = 3-6 \cdot Si\#$ (see Sect. Methods: [Mineralogical considerations](#)). For a pyrolite, the a priori Mg#, Si#, and M_{Ol} values are 89, 41 and 55%. The inversion is regularized using a priori covariance matrices in the model (C_m) and data (C_d) spaces⁶⁷. These covariance matrices are diagonal and their diagonal terms are the variance of the model and data, i.e. the square of the model and data a priori errors. After some trials, we set the model a priori error σ_m to 3% and the data errors σ_d to 1% of the V_s values and 0.5% of the density values. A $\pm 1\%$ data error on V_s is on the conservative side of recent V_s estimates⁶⁸ and a $\pm 0.5\%$ data error on density is consistent with the idea that the excess density of cratonic roots is within of $0.5\% \pm 0.5\%$ (see Methods: Isostasy and Geoid). The chosen σ_m value allows us to recover the expected variations in Mg# and M_{Ol} . Larger σ_m values result in unrealistic Mg# and M_{Ol} values without improving significantly data fit. Partial derivatives of the velocity and density with respect to the inverted parameters, Mg#, and Si# are shown in Fig. S16. The $\frac{\delta V_s}{\delta Mg\#}$ and $\frac{\delta \rho}{\delta Mg\#}$ are much larger than the $\frac{\delta V_s}{\delta Si\#}$ and $\frac{\delta \rho}{\delta Si\#}$, indicating that Mg# is better constrained than M_{Ol} . This is indeed what we observe. The inverted Mg# values are insensitive to the starting model, while M_{Ol} values are affected by the starting composition. Using a pyrolytic or harzburgite starting model produces the same Mg# values and similar M_{Ol} perturbations, but with respect to a different starting model.

Explaining the misfit with radial anisotropy

Our seismic model (Fig. S1) is an isotropic S-wave model constructed using the radial anisotropy $\zeta = (V_{Sh}/V_{Sv})^2$ of the PREM⁶¹ model. Here, we compute the radial anisotropy ζ required to eliminate the misfit shown in Fig. S7,

without adding diamonds at depths of 150 km or below. Figure S8 shows that neither the pattern nor the amplitude match. The global mean ξ values over cratonic regions would be close or smaller than 1 below 125 km depth, in contrast with SEMUM2⁵⁶ or S362ANI²³, which exceed PREM and reach values close to 1.05 at 150 km depth. Consequently, radial anisotropy is unlikely to explain seismic velocities inferred from tomography.

Explaining the misfit with diamond and eclogite

The diamond and eclogite volume fraction maps displayed in Figs. 4 and S10 explain the velocity difference that remains after inverting for Mg# and M_{Ol} (Fig. S7). The shear velocity and density after inversion are used to extract the predicted shear modulus μ_p of the bulk rock after inversion. In the same way, we extract from our data (the velocity from our tomographic model¹ and the Density 2 profile of Table S1) a “tomographic” shear modulus, μ_t . We then consider that μ_t is a Voigt-Reuss-Hill average of the predicted rigidity with that of a volume fraction x of another component (eclogite or diamond) with shear moduli μ_c . This Voigt-Reuss-Hill average is expressed as:

$$\mu_t = \frac{(\mu_p + x(\mu_c - \mu_p)) + 1/\left(\frac{1}{\mu_p} + x\left(\frac{1}{\mu_c} - \frac{1}{\mu_p}\right)\right)}{2}$$

and μ_t is such that it accounts for the residual velocity anomalies. We solve this equation for x to obtain the maps shown in Figs. 4 and S10. Knowing x , we then compute the effect on density (expected to be small according to Fig. S9). We use a Voigt average, which is more adapted for a volume change¹⁷, to express the final density ρ_f as a function of the predicted density (after inversion) ρ_p and the component (eclogite or diamond) density ρ_c :

$$\rho_f = (1 - x)\rho_p + x\rho_c$$

The density differences in % with the Density 2 profile of Table S1, are displayed in Figs. 4 and S10.

Supplementary Information The online version contains supplementary material available at <https://doi.org/10.1038/s41598-026-74399-2>.

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Author contributions E.D. and Y.R. collaborated in developing the concept of this paper. E.D. wrote the codes for the interpretation of the seismic models and wrote the drafts of the manuscript. Y.R. developed the concepts of the mineralogical interpretation, contributed to all mineralogical aspects and to the writing of the manuscript.

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Data availability The 3D Vs model used in this study is available as IRIS data products at <https://ds.iris.edu/ds/products/cm-3dgl-tpesv/>.

Code availability Requests about the numerical modelling codes associated with this paper should be sent to Eric.Debayle@ens-lyon.fr. Most figures were created with open software GMT 4.5.13.

Declarations

Competing interests The authors declare no competing interests.

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Authors and Affiliations

Eric Debayle¹  · Yanick Ricard¹

✉ Eric Debayle
Eric.Debayle@univ-lyon1.fr

¹ ENS de Lyon, Université Lyon 1, Université Jean Monnet, CNRS, LGL-TPE, UMR 5276, Villeurbanne, Lyon, France