Intraplate volcanism originating from upwelling hydrous mantle transition zone

Most magmatism occurring on Earth is conventionally attributed to passive mantle upwelling at mid-ocean ridges, to slab devolatilization at subduction zones, or to mantle plumes. However, the widespread Cenozoic intraplate volcanism in northeast China\(^1\)–\(^3\) and the young petit-spot volcanoes\(^4\)–\(^7\) offshore of the Japan Trench cannot readily be associated with any of these mechanisms. In addition, the mantle beneath these types of volcanism is characterized by zones of anomalously low seismic velocity above and below the transition zone\(^8\)–\(^12\) (a mantle level located at depths between 410 and 660 kilometres). A comprehensive interpretation of these phenomena is lacking. Here we show that most (or possibly all) of the intraplate and petit-spot volcanism and low-velocity zones around the Japanese subduction zone can be explained by the Cenozoic interaction of the subducting Pacific slab with a hydrous mantle transition zone. Numerical modelling indicates that 0.2 to 0.3 weight per cent of water dissolved in mantle minerals that are driven out from the transition zone in response to subduction and retreat of a tectonic plate is sufficient to reproduce the observations. This suggests that a critical amount of water may have accumulated in the transition zone around this subduction zone, as well as in others of the Tethyan tectonic belt\(^8\) that are characterized by intraplate or petit-spot volcanism and low-velocity zones in the underlying mantle.

The Cenozoic intraplate volcanism in northeast China is located more than 1,000 km westward of the Japan Trench\(^1\), while the young alkaline basalts (0–6 Ma) known as petit-spots outcrop up to 600 km eastward of the trench\(^1\) (Fig. 1). The formation mechanism of these types of onshore and offshore volcanism is still debated, as there is no geological and geophysical correlation with mantle plumes or arc volcanism\(^6\)–\(^14\).

Seismic tomography models indicate that in this region the Pacific Plate is currently stagnant in the mantle transition zone (MTZ), extending continuously up to nearly 1,000 km to the inland of northeast China\(^3\)–\(^8\),\(^10\). Thus, it has been proposed that the Cenozoic intraplate magmatism is related to the dehydration of the Pacific slab in the MTZ\(^8\)–\(^15\).

The primary petit-spot magma has been determined to be volatile-rich with extremely enriched mantle (EMI-like) isotopic compositions\(^6\)–\(^8\). The lack of hotspot tracks in this region excludes a contribution from a mantle plume. It has been postulated that the petit-spot magma forms in the asthenosphere and migrates upward through the oceanic lithosphere by reactive porous flow in response to plate flexure\(^6\)–\(^8\),\(^11\).

Based on electrical conductivity surveys, the MTZ probably holds about 0.1 wt% water\(^16\). The MTZ below northeast China and Japan is particularly wet, with at least 0.5–1 wt% water\(^15\). The MTZ is primarily composed of wadsleyite and ringwoodite minerals that can accommodate 1–3 wt% water, which is 1 to 2 orders of magnitude higher than the water (hydrogen) solubility in upper- and lower-mantle minerals. Given the large contrast in water solubility between the MTZ and upper- or lower-mantle, it is reasonable to expect deep dehydration melting when subducting slabs excite vertical flow in the nearby wet MTZ\(^8\). Indeed, seismic low-velocity zones (LVZs) above 410 km and below 660 km have been observed not only in Japan\(^8\)–\(^12\),\(^14\),\(^15\), but also around subduction zones in Europe\(^9\) and the western United States\(^20\),\(^21\).

To test this hypothesis, we construct two-dimensional numerical experiments in which a self-sustained oceanic plate subduction is characterized by trench retreat and slab stagnation into a homogeneously or heterogeneously wet MTZ (see Methods). The subducting plate and entrained dry upper mantle push the adjacent wet MTZ downward to the lower mantle such that a partially molten layer forms between 700 km and 800 km depth (Fig. 2a, region labelled M2) (Supplementary Video 1). On the other hand, MTZ material uplifted to the upper mantle starts to partially melt above 410 km (Fig. 2a, M3). Slab stagnation and retreat is accomplished by sub-slab MTZ upwelling and new melting (M1). These partially molten regions above and below the MTZ cause large seismic LVZs (Fig. 2c). When melt percolation is active (see Methods), extraction to the surface occurs, forming intraplate and petit-spot volcanism ahead of and behind the trench, respectively.

Figure 3 shows the spatial and temporal trend of modelled volcanics for the reference model in Fig. 2. The first intraplate volcanism occurs about 500 km away from the trench, then spreads in two opposite directions. The mantle water content decreases after melt extraction, which precludes further deep (>200 km) melting of the residual peridotite (Fig. 2b). As the slab rolls back, more distal wet MTZ is sucked into the upper mantle wedge, such that partial melting and volcanoes will form further away from the trench. The new generated volcanism is not homogeneously distributed as it is strongly influenced by mantle flow and trench movement. Furthermore, a heterogeneous distribution of water in the MTZ would prevent the formation of any...
temporal-spatial magmatic sequence as wetter portions would melt earlier than drier regions at the same pressure and temperature (P–T) conditions. It is noteworthy that intraplate volcanism also occurs a few hundred kilometres in front of the slab tip. After about 12 Myr of modelled subduction, petit-spot volcanoes appear behind the trench. They are located up to about 300 km seaward of the trench and exhibit a similar magmatic activity trend to the intraplate volcanism.

We further test the influence of initial water content in the transition zone and other parameters on the genesis of asthenospheric melting (Extended Data Fig. 5). Melting commences 40 km above the transition zone, and no petit-spot volcanoes are formed for 0.2 wt% initial water. However, the results also hold for a more realistic heterogeneous distribution of the water distribution of water in the MTZ, these models provide upper-bound estimates on the volumes of volcanics and melt. However, the results also hold for a more realistic heterogeneous distribution of the water in this mantle level (Extended Data Fig. 5e).

When comparing the model results with seismic and geological observations, we note that around the Pacific slab, three remarkable seismic LVZs outside the transition zone are clearly imaged (Fig. 1b). These are well correlated with the locations of intraplate and petit-spot volcanoes, and the modelled partially molten zones (Figs. 1a and 2). Although seismic low-velocity anomalies are generally attributed to thermal effects or to the presence of water, melt and/or major element compositional heterogeneities, it has recently been argued that some of these LVZs could be artefacts induced by seismic anisotropy. Nevertheless, the authenticity of the sub-slabs LVZ1 and LVZ2 appearing in tomographic models has been confirmed by other independent studies using, respectively, an accurate scrutiny of the seismic ray paths sampling the LVZ1, and receiver functions in the case of LVZ2. The LVZ3 sits below the active Changbai volcano and appears to extend down to 410 km as revealed by multiple high-resolution tomography models. A thermal anomaly from a non-hotspot upwelling, if it hypothetically exists, is difficult to reconcile with the large velocity drop of LVZ1. The hot material will rapidly cool when flowing upward, because of adiabatic decompression and the latent heat of the wadsleyite-to-olivine reaction. Laboratory experiments show that seismic wave speeds are insensitive to moderate (<1 wt%) water contents for olivine and wadsleyite; thus, the LVZs are very likely to be caused by partial melting and/or compositional heterogeneities. The presence of basalts at the bottom of the upper mantle can be excluded as it would generate a positive seismic anomaly. On the other hand, basalts accumulating at the base of the MTZ could be effectively dragged by the slab into the uppermost lower mantle and generate the LVZ2. However, receiver functions indicate that the lower-mantle LVZs are within the 750–780 km depth range, which is likely to be below the post-garnet phase transition where basalts are seismically faster than mantle rocks.

The presence of melt in the deep mantle, which is mostly catalysed by the involvement of volatiles, decreases seismic velocities and provides a magmatic source for intraplate/petit-spot volcanism. Our numerical models suggest that a hydrous transition zone with at least 0.2–0.3 wt% water beneath northeast China and offshore Japan can
comprehensively explain the LVZs and the intraplate and petit-spot volcanism. This model does not exclude the devolatilization of the stagnant Pacific slab as a mechanism to explain the LVZ3 region and the overlying intraplate volcanism, which favours the upwelling of volatile-rich plumes from the MTZ as envisaged by the Big Mantle Wedge model. However, the same slab-derived volatiles cannot obviously be the cause of both the LVZ1 and LVZ2 and of the petit-spots, implying the presence of a metasomatized MTZ before the last subduction episode. The accumulation of water in the MTZ could be caused by, for example, delamination of volatile-rich lithospheric roots, or by previous slab dehydration episodes in the MTZ and subsequent absorption of the water by wadsleyite and ringwoodite. Alongside with water, reduced (by redox-freezing) carbonated sources and restitic K-hollandite-bearing sediments are required to explain the volatile-rich, alkaline and EM1-type petrological and geochemical signature of the basalts. This is not surprising, as the MTZ, a graveyard for stagnating slabs, is the most likely candidate to host volatiles and subducted sediments, and long-term isolation of these MTZ domains would be consistent with the ancient metasomatizing episodes estimated for intraplate basalts. Subsequent subduction events would mobilize the wet and (carbon + alkali)-bearing MTZ rocks, promoting the formation of silica-undersaturated magmas in the upper mantle. It is important to note that the addition of these components is not critical to our results, because the location and amounts of partial melting above and below the MTZ will still be dictated by the distribution of wet MTZ domains, while reduced carbonated sources are expected to experience redox melting at shallower depths (<250 km).

Fig. 2 | Dynamics of subduction-induced dehydration melting above and below the mantle transition zone. a, Composition field. A colour key indicating different rock types is given at the bottom. Two horizontal black lines mark depths of 410 km and 660 km. Three partially molten regions (M1, M2 and M3) are indicated. b, Water content with temperature contours. c, Seismic P-wave velocity anomalies. An initial water content of 0.3 wt% is assumed in the MTZ, and the reference melt extraction timescale $t_{\text{ref}} = 6$ kyr (see Methods).
The process proposed here could potentially explain also the Cenozoic anorogenic volcanism in the Mediterranean\cite{13} and intraplate volcanism in the Turkish–Iranian Plateau\cite{36} regions characterized by the long-term subduction of the Tethys Ocean. Together with surface intraplate/petit-spot volcanism, constraints on deep seismic low velocities and/or high electrical conductivity may thus indicate a volatile-rich and/or partially molten mantle within and around the transition zone.

Online content

Any methods, additional references, Nature Research reporting summaries, source data, extended data, supplementary information, acknowledgements, peer review information; details of author contributions and competing interests; and statements of data and code availability are available at https://doi.org/10.1038/s41586-020-2045-y.

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Methods

Modelling approach
The 2D petrological–thermomechanical numerical code I2VIS used in this study is based on a finite difference method using a marker-in-cell technique on a staggered grid. It solves mass, momentum and energy conservation equations (1)–(3) on the Eulerian grid and interpolates physical properties to the markers for advection accordingly.

\[
\frac{\partial \upsilon}{\partial t} = 0
\]

\[
\frac{\partial P}{\partial x} + \frac{\partial}{\partial x} \left( \eta \left( \frac{\partial \upsilon}{\partial x} + \frac{\partial \upsilon}{\partial y} \right) \right) + \rho g = 0
\]

\[
\rho C_p \frac{\partial T}{\partial t} + \upsilon_i \frac{\partial T}{\partial x_i} = \frac{\partial}{\partial x} \left( k \frac{\partial T}{\partial x} \right) + H_e + H_s + H_b
\]

where \(v_i\) is velocity, \(x_i\) coordinate, \(P\) dynamic pressure, \(\rho\) density, \(g\) gravity acceleration, \(c_h\) heat capacity, \(T\) temperature, \(k\) thermal conductivity, \(H_e\) radioactive heating, \(H_s\) shear heating, and \(H_b\) adiabatic heating, where \(g\) is the material time derivative. The latent heat is implicitly considered by computing the effective thermal expansion and heat capacity.

Model configuration
The initial model set-up (6,000 × 1,000 km discretized with 1,501 × 501 nodes) is composed of a 3,500-km subducting plate and a 2,500-km overriding plate. The model imposes free-slip mechanical boundary condition at the top with 30-km-thick and viscosity of 10^{18} Pa s 'sticky-air' to mimic free surface; the bottom boundary is no slip, and side boundaries are periodic. The bottom no-slip condition is needed to define an initial horizontal velocity from which finite differences can be computed for this variable. Comparison with results from a model with a bottom free-slip condition and closed vertical walls indicate that the bottom no-slip boundary condition does not affect the subduction dynamics at all, as it is confined above the lower mantle. The initial thermal structure is defined by the half-space cooling age for the plates (50 Myr old) and an adiabatic thermal gradient of 0.5 K km^{-1} for the underlying mantle. The thermal boundary conditions are isothermal on the top and bottom, while side boundaries are periodic, consistent with the mechanical boundary conditions.

To initiate subduction, the subducting slab extends down to about 200 km in the upper mantle together with a rheologically weak zone on top of it which lubricate the initial contact between the plates. The high numerical resolution (4 km × 2 km) used here is needed to ensure plate contact lubrication at shallow depths and localized, bending-related hydration at the trench outer-rise. Tests at a lower resolution (4 km × 4 km) result in less-localized slab mantle hydration, whereas with a resolution of 8 km × 4 km, self-sustained subduction and slab rollback do not appear spontaneously.

Viscous–plastic rheological model
The rock mechanical behaviour is represented by the effective viscosity, which combines ductile (dislocation, diffusion and Peierls creep) and brittle (Drucker–Prager) deformation. The effective ductile viscosity is given by the harmonic average of the combined rheologies (parameters and physical meaning are defined in Extended Data Table 1):

\[
\eta_{\text{ductile}} = \left( \frac{1}{\eta_{\text{diss}}} + \frac{1}{\eta_{\text{diff}}} + \frac{1}{\eta_{\text{Peierls}}} \right)^{-1}
\]

where the dislocation and diffusion creep are given by:

\[
\dot{\epsilon} = A (\sigma / \mu)^n (b / d)^m \exp \left( -\frac{E + PV}{RT} \right) \exp (\alpha \phi)
\]

\[
\eta = \frac{\sigma}{2\dot{\epsilon}}
\]

For hydrated (wet) mantle, viscosity is reduced by \(\eta_{\text{wet}} = \eta_{\text{dry}} \left( \frac{C_w}{C_{w_0}} \right)^{1-r/n} \), and \(C_w, C_{w_0}\) are water content and reference water content (100 ppm, which is the water content for the dry upper mantle), respectively. The Peierls creep \(\eta_{\text{Peierls}}\) is given by:

\[
\eta_{\text{Peierls}} = 0.5A \sigma^{\gamma-1} \exp \left[ \frac{E_{\text{Peierls}} + P V_{\text{Peierls}}}{RT} \left( 1 - \frac{\sigma^{\gamma}}{\sigma_{\text{Peierls}}} \right)^{\gamma/\sigma_{\text{Peierls}}} \right]
\]

Parameters are defined in Extended Data Table 1.

Brittle behaviour occurs when stresses are above the plastic yield stress \(\tau_y\):

\[
\eta_{\text{ductile}} \leq \frac{\tau_y}{2\dot{\epsilon}_g}
\]

\[
\tau_y = C + \mu P
\]

Petrological modelling
Petrological solid–solid phase changes are included through the density and enthalpy look-up tables for basalt and pyrolite obtained from PERPLE_X. Therefore, phase transition boundaries at 410 km and 660 km have been considered.

The solids (\(T_s = f(P, T, H_2O)\)) and liquidus (\(T_l = f(P, T)\)) temperatures for the upper mantle and MTZ are taken from high-pressure experiments Extended Data Fig. 1). At lower mantle conditions, \(T_s\) and \(T_l\) vary considerably among different experiments. Here we adopt the dry solidus and liquidus of chondritic mantle, as these are more compatible with the results of KLB-1 peridotite, while the wet solidus was measured on samples with an estimated water content of 400 ppm wt.

A conservative estimate of the melt fraction in the wet upper mantle is applied:

\[
\phi = \frac{W_m - W_a}{W_m - W_{l0}}
\]

where \(W_m, W_a\) and \(W_{l0}\) (10 wt%) are water mass fraction of the ambient mantle, olivine and melt, respectively. Note that the water solubility in olivine increases with pressure, so the melt fraction will decrease with depth if \(W_m\) remains constant (Extended Data Fig. 3).

The silicate melt density (Extended Data Fig. 2) is taken from high-pressure sink–float experiments, which show that the melt becomes denser than the surrounding mantle at 400 km (refs. 45,46) owing to the increased compressibility. However, the presence of water generally reduces the melt density such that it becomes buoyant relative to solid mantle (see Extended Data Fig. 2), rendering melt extraction at this depth possible. We also test the melt density from molecular dynamics simulations at high-pressure conditions Extended Data Figs. 4f, 5d).

Melt extraction timescale
The distance over which the compaction rate decreases by a factor of e is the characteristic length scale of the compaction process and is known as the compaction length, \(\delta_c\):

\[
\delta_c = \frac{K(z + \frac{1}{2} \eta)}{\eta_f}
\]
where $\zeta$ and $\eta$ are the effective bulk and shear viscosities, respectively, of the partially molten rock; $\eta_i$ is the fluid viscosity; $K$ is the permeability given by the empirical equation:

$$K = K_0 \left( \frac{\phi}{\phi_0} \right)^n$$

(12)

where $\phi$ is the porosity (melt fraction); $K_0$, ($10^{-12}$ m$^2$) is the permeability at the reference porosity $\phi_0$, (0.01); and $n = 3$.

The relative migration velocity between the melt and the solid matrix is $w$:

$$w = \frac{K \Delta \rho g}{\eta_0 \phi}$$

(13)

Thus, the extraction timescale $t$:

$$t = \frac{\zeta}{w} = \frac{\eta_0}{\eta_0 + \phi} \left( \zeta + \frac{\phi}{\eta} \right)$$

(14)

where $\zeta = \frac{\eta}{\phi}$, so $\zeta + \frac{\phi}{\eta} = \frac{\eta_0}{\phi_0}$; if $\eta_l = 1$ Pa s, then

$$t = \frac{10^4 \Delta \rho g \phi}{\Delta \rho g \phi}$$

(15)

where $\Delta \rho$ is the density difference between the solid and melt, ranging from about $-100$ kg m$^{-3}$ to $270$ kg m$^{-3}$ and typically about $70$–$180$ kg m$^{-3}$ in the lowermost upper mantle if most of the water is partitioned into melt, and the surrounding mantle viscosity $\eta = 10^{29}$–$10^{30}$ Pa s in the upper mantle. For melt fraction $\phi = 0.02$, the estimated timescale would be $t_{\text{er}} = 3$–20 kyr. But only this timescale is the not the time for melt migration to the surface. On the other hand, dehydration of the underlying mantle within the transition zone is thought to cause intracratonential magmatism. Consequently, we allow for serpentinization by bending-related deformation when the strain of mantle rocks is greater than 0.1 (ref. 50).

Upon partial melting and extraction, the water is partitioned into the extracted melt and water in the residual peridotite as:

$$V_i = V_i^{\text{ref}} - \frac{\nabla P - \rho_i \mathbf{g}}{(\rho_s - \rho_i) g_y} V_0$$

(16)

where $V_i$, $V_i^{\text{ref}}$ are the velocities of solid and fluid phases, respectively; $\rho_s$, $\rho_i$ are the densities of solid and fluid, respectively; $V_0$ is a constant percolation velocity; $g$ is the gravity acceleration vector as defined in equation (2), and $g_y$ is its vertical component.

Falling block tests. The validity of the petrological model used here can be easily tested with a simple model in which a falling block (simulating the subducting slab) sinks into the wet MTZ, exciting wet upwellings to the upper mantle and squeezing water into the lower mantle (Extended Data Fig. 3). These tests indicate that the melt layer gets thicker (>100 km) when melt extraction is not efficient, owing to very small amounts of melt/water and dense melt phase. This might explain the thick low-velocity layers above 410 km in many regions. After melt extraction, less water remains above the transition zone, causing higher viscosity and less melt fraction, which yields a larger extraction timescale: that is, the melt preferentially ponds above 410 km depth.

Seismic velocity anomalies. The seismic velocity perturbation in the upper mantle and squeezing water into the lower mantle (Extended Data Fig. 4). These tests indicate that the melt layer gets thicker (>100 km) when melt extraction is not efficient, owing to very small amounts of melt/water and dense melt phase. This might explain the thick low-velocity layers above 410 km in many regions. After melt extraction, less water remains above the transition zone, causing higher viscosity and less melt fraction, which yields a larger extraction timescale: that is, the melt preferentially ponds above 410 km depth.

The melt migration process is illustrated here with more realistic models accounting for visco-elastoplastic deformation in a two-phase flow regime. These models demonstrate that melt migration from the deep upper mantle to the surface should occur through several mechanisms: viscous diapirism, viscoplastic decoupling channels, and elastoplastic melting (ref. 46). For weak host rocks where viscous deformation dominates, such as the asthenosphere, magma migrates by diapirism. When the magma moves through the lithosphere–asthenosphere boundary (or the lower crust in continents) where both ductile and brittle deformation occur, the fluid compaction pressure might reach the tensile strength, and magma could migrate by channelling. If the host rock is completely elastoplastic, such as the core of lithospheric mantle and upper crust, magma migrates by dyking.
The change of seismic wave velocities caused by the existence of a fluid phase is given by:

$$\frac{V_s}{V_p} = \sqrt{\frac{N/\mu}{\sqrt{p/\rho}}}$$  \hspace{1cm} (20)

and

$$\frac{V_s}{V_p} = \frac{K_{\text{eff}}}{k} + (4y/3)N/\mu$$  \hspace{1cm} (21)

where

$$\frac{K_{\text{eff}}}{k} = \frac{K_b}{k} + \frac{(1 - K_b/k)^2}{1 + 4y/3}$$  \hspace{1cm} (22)

and $y = \frac{\mu}{\rho} = \frac{3(1 - 2\nu)}{2(1 + \nu)}$. $V_s$, $V_p$ are the shear and compressional wave velocities of the solid phase; $k$, $\mu$, $\nu$, and $\rho$ are the bulk modulus, shear modulus, Poisson’s ratio and density of the solid phase, respectively. $p = (1 - \phi) + \rho_{\text{ff}}$ is the effective density when fluid (for example, melt) exists. $K_b$ and $N$ are the bulk and shear moduli, which are dependent on melt fraction and dihedral angle $\gamma$:

$$K_b = (1 - \phi)(1 - \phi)$$  \hspace{1cm} (23)

$$N = (1 - \phi)(1 - \phi)$$  \hspace{1cm} (24)

where

$$n_k = a_1\phi + a_2(1 - \phi) + a_3\phi(1 - \phi)^{1.5}$$  \hspace{1cm} (25)

$$n_\rho = b_1\phi + b_2(1 - \phi) + b_3\phi(1 - \phi)^{2}$$  \hspace{1cm} (26)

and $\phi$ is the dihedral angle

$$\phi = \frac{2A_s}{A_1 + A_d}$$  \hspace{1cm} (27)

with $A_s, A_d$ the area of solid–solid contact and solid–liquid contact, respectively.

Extended Data Fig. 7 shows $K_b/k$ and $N/\mu$ for the equilibrium geometry model at various dihedral angles.

Data availability

The dataset generated during the current study is available at https://figshare.com/articles/Yang_Faccenda_Nature2019/9933056.

Code availability

Requests about the numerical modelling codes associated with this paper should be sent to the main code developer (taras.gerya@erdw.ethz.ch). The map in Fig. 1a is created with open software GMT 5.4.3 which is under a GNU Lesser General Public License. The numerical 2D finite element code MVEP2 (https://bitbucket.org/bkaus/mvep2) was used for the two-phase flow model in Extended Data Fig. 6.
Extended Data Fig. 1 | Solidus and liquidus of basalt and mantle. a, The solidus and liquidus of basalt are obtained from experimental data42,59,60. The solidus from ref. 60 fits well within the uncertainty region of ref. 42 and is thus adopted. b, Solidus and/or liquidus of mantle collected from literature. Sol, solidus; Liq, liquidus; BrPe, MgSiO₃-MgO (bridgmanite + periclase); Fiqul, MgSiO₃-SiO₂ (bridgmanite + stishovite). Experimental data are from refs. 41,44,61-66.
Extended Data Fig. 2 | Melt density of basalt and mantle for different temperatures and/or water contents. a, Basalt. PREM, density profile from Preliminary Reference Earth Model; dry melt density at temperatures of 1,673 K, 2,073 K and 2,473 K (ref. 47) and 2,735 K (ref. 67); dry and wet with 2 wt% and 8 wt% H2O melt density at 2,473 K (ref. 45); the modelled basalt (MB), hydrated basalt (hyMB) and basalt (MORB)68. b, Mantle. Melt density of dry peridotite47; dry and wet (2 wt% and 8 wt% H2O)45; wet peridotite (3 wt%, 5 wt%, 7 wt% H2O)46; dry peridotite and komatiite69 and perovskite (pv)70. Note the density crossover at around 13 GPa (refs. 45,46). All the profiles are fitted by third or fourth order of the Birch–Murnaghan equation of state.
Extended Data Fig. 3 | Phase diagram of H₂O-peridotite, after ref. 53. The solidus/liquidus curves are the same as in Extended Data Fig. 1. The grey lines are olivine–wadsleyite (Ol–Wd) and ringwoodite–perovskite (Rd–Pv) phase boundaries. The abbreviations of major hydrous phases are as follows: Chl, chlorite; Serp, serpentine; A, phase A; E, phase E; shyB, superhydrous phase B; D, phase D.
Extended Data Fig. 4 | Falling block simulations with different parameters.

a, Reference model with initial MTZ water content of 0.3 wt%, melt density from ref. 45 and reference extraction timescale $t_{\text{ref}} = 6$ kyr. b–f, Other tests are similar to this model except for (b) initial water content $C_w = 0.2$ wt%, (c) $C_w = 0.3$ wt% and $t_{\text{ref}} = 4$ kyr, (d) $C_w = 0.3$ wt% and $t_{\text{ref}} = 8$ kyr, (e) $C_w = 0.2$ wt% and $t_{\text{ref}} = 15$ kyr, (f) $C_w = 0.3$ wt% and $t_{\text{ref}} = 4$ kyr by using the melt density from ref. 47. Note that the extraction timescale is calculated only when the melt is less dense than the solid matrix.
Extended Data Fig. 5 Additional parameter tests. **a, b,** Extraction timescales of 4 kyr (**a**) and 8 kyr (**b**), with 0.3 wt% initial water content in both. **c,** Initial water content 0.2 wt%. **d,** Melt density from ref. 47 and \( t_{\text{ref}} = 4 \) kyr. **e,** Wet inclusions in the transition zone with \( t_{\text{ref}} = 6 \) kyr. Note that all the models differ by only one parameter from the reference model (Fig. 2), except **d.**
Extended Data Fig. 6 | Visco-plastic shear viscosity for melt percolation in two-phase flow. a–c, Melt percolation at three typical stages as (a), diapirism (b), channelling and (c), dyking from deep mantle to the surface. The numerical 2D finite element code MVEP2 was used to simulate melt migration dynamics. A small background strain rate ($10^{-15}$ s$^{-1}$; the model domain was extended by only 0.75 km after 12.2 kyr) was applied at the side boundaries. The top boundary is free surface. An initial porosity at the bottom boundary with Gaussian distribution (resulting in an average porosity of 0.127) was applied. The details of the approach allowing for its reproduction are provided elsewhere$^{51,52}$. 


Extended Data Fig. 7 | Normalized bulk modulus $K_b/k$ and shear modulus $N/\mu$ of skeleton (solid porous matrix) versus melt fraction. The ratios of both bulk and shear modulus decrease with melt fraction. The numbers shown on the lines are dihedral angles.
## Extended Data Table 1 | Physical properties of rocks used in this study

| Property                        | Symbol | Unit    | Value   |
|---------------------------------|--------|---------|---------|
| Gravity                         | g      | m/s²    | 9.81    |
| Water content                   | C_w    | wt. %   | -       |
| Reference water content         | C_{w0} | wt. %   | 0.01    |
| Melt fraction                   | φ      | -       | -       |
| Melt-weakening factor           | α      | -       | 28      |
| Shear modulus                   | μ      | GPa     | 80      |

### Diffusion creep

| Property                        | Symbol | Unit     | Value     |
|---------------------------------|--------|----------|-----------|
| Prefactor                       | A      | s⁻¹      | 8.7×10¹⁵  |
| Activation energy               | E      | kJ mol⁻¹ | 300       |
| Activation volume               | V      | cm³ mol⁻¹| 6         |
| Burgers vector                  | b      | nm       | 0.5       |
| Grain-size exponent             | m      | -        | 2.5       |
| Water exponent                  | r      | -        | 0.8       |

### Dislocation creep

| Property                        | Symbol | Unit     | Value     |
|---------------------------------|--------|----------|-----------|
| Prefactor                       | A      | s⁻¹      | 3.5×10²²  |
| Activation energy               | E      | kJ mol⁻¹ | 540       |
| Activation volume               | V      | cm³ mol⁻¹| 20        |
| Stress exponent                 | n      | -        | 3.5       |
| Water exponent                  | r      | -        | 1.2       |

### Peierls creep

| Property                        | Symbol | Unit    | Value   |
|---------------------------------|--------|---------|---------|
| Prefactor                       | A_p    | Pa² s   | 10⁴.²   |
| Activation energy               | E_{Peierls} | kJ mol⁻¹ | 532     |
| Activation volume               | V_{Peierls} | cm³ mol⁻¹| 12      |
| Peierls stress                  | σ_{Peierls} | GPa    | 9.1     |
| Exponent                        | p, q   | -, -    | 1, 2    |

### Yield stress τ_y

| Property                        | Symbol | Unit    | Value   |
|---------------------------------|--------|---------|---------|
| Cohesion                        | G      | MPa     | 10      |
| Friction coefficient            | μ      | -       | 0.6     |