Reconciling surface plate motions with rapid three-dimensional mantle flow around a slab edge

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The direction of tectonic plate motion at the Earth’s surface and the flow field of the mantle inferred from seismic anisotropy are well correlated globally, suggesting large-scale coupling between the mantle and the surface plates1,2. The fit is typically poor at subduction zones, however, where regional observations of seismic anisotropy suggest that the direction of mantle flow is not parallel to3–7 and may be several times faster than8 plate motions. Here we present three-dimensional numerical models of buoyancy-driven deformation with realistic slab geometry for the Alaska subduction–transform system and use them to determine the origin of this regional decoupling of flow. We find that near a subduction zone edge, mantle flow velocities have magnitudes of more than ten times the surface plate motions, whereas surface plate velocities are consistent with plate motions9 and the complex mantle flow field is consistent with observations from seismic anisotropy10. The seismic anisotropy observations constrain the shape of the eastern slab edge and require non-Newtonian mantle rheology. The incorporation of the non-Newtonian viscosity11–13 results in mantle viscosities of $10^{17}$ to $10^{18}$ Pa s in regions of high strain rate ($10^{-12}$ s$^{-1}$), and this low viscosity enables the mantle flow field to decouple partially from the motion of the surface plates. These results imply local rapid transport of geochemical signatures through subduction zones and that the internal deformation of slabs decreases the slab-pull force available to drive subducting plates.

In olivine grains, the dislocation creep deformation mechanism leads to lattice-preferred orientation (LPO)14 and a non-Newtonian rheology15, that is, a power-law dependence between strain rate and stress. In the mantle, LPO can lead to anisotropy in seismic velocities11,12. Measurements of seismic anisotropy near subduction zones show trench-parallel fast directions beneath slabs3,7 and above slabs4,6, as well as fast directions that are perpendicular to the edges of slabs3. If the fast seismic direction tracks mantle flow16,12, then there is spatially variable flow at subduction zones and the motion of surface plates is partially decoupled from mantle flow9.

Numerical models and laboratory experiments of subduction can provide a fluid dynamics explanation for the patterns of mantle flow inferred from observations of seismic anisotropy. Three-dimensional (3D) models of subduction–transform systems, commonly in the context of slab rollback, indicate a significant component of toroidal flow around the slab edge and a small component of trench-parallel flow in the mantle wedge beneath the slab14–19. In addition, trench-parallel seismic fast directions match stretching directions generated by along-strike changes in slab geometry (dip and curvature) in models with a non-Newtonian rheology14. These models provide valuable insights into the dynamics governing mantle flow in subduction zones; however, they are limited because they prescribe a velocity boundary condition for the subducting plate, which is kinematically driven, do not include an overriding plate or use a Newtonian rheology.

The southern Alaska subduction zone is a good location to study to understand how the 3D structure of a slab drives plate motion and couples to mantle flow, because sufficient data exist to characterize the slab shape, there are seismic observations constraining the pattern of mantle flow and it is isolated from other subduction zones (Supplementary Information). We construct 3D models of buoyancy-driven flow for this subduction–transform plate boundary system and compare model results with observations of Pacific plate motion (PPM) and SKS fast-axis seismic directions9 (SKS waves are S waves converted to compressional waves where they pass through the outer core) (Fig. 1). These models test how mantle rheology, the slab geometry, yield strength ($\sigma_y$) and the viscosity of the plate boundary shear zone (PBSZ) influence the surface plate motions and underlying mantle flow (Supplementary Information). For the mantle rheology, we use either Newtonian-only viscosity or a composite (Newtonian and non-Newtonian) viscosity (Methods). The slab geometry, constrained by seismicity and tomography, varies along the length of the subduction zone, forming a flat slab beneath south-central Alaska and steepening close to the transform boundary. Because seismic observations do not constrain the exact depth of the slab edge, we construct two slab shapes, slab_{E325} and slab_{E115}, where the subscript refers to the depth of the slab in kilometres at the eastern slab edge (Fig. 1).

**Figure 1** Schematic of full model domain and slab geometry. Outline of the high-resolution mesh region (dashed grey line); the portion of the slab geometry that is varied (short-dashed black line); the PBSZ and southern mesh boundary shear zone (SMSZ) (thick dark-grey lines); Juan de Fuca Ridge (JdFR, double black line); and the locations of the cross-sections shown in Fig. 3 (AA’ and BB’, black). NAM, North American plate; PAC, Pacific plate; $T$, temperature. See Methods and Supplementary Information.

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We first compare PPM predicted from the models with the observed PPM, to verify that models producing high flow velocities in the mantle also produce surface plate motions consistent with first-order observations. Although the models do not include the full Pacific plate or its slabs, they are designed such that the region of the Pacific plate is free to move in response to the local driving forces, and this motion should be generally consistent with PPM (Supplementary Information).

Off the shore of south-central Alaska, the observed PPM with respect to North America is approximately N 16° W at 5.2 cm yr⁻¹ (ref. 8). In models with a Newtonian viscosity and slab₁₁₅, the predicted PPM is ~N 27° W at 1.9 cm yr⁻¹ (Fig. 2). The predicted PPM is more northerly and slightly faster (~N 20° W at 4.4 cm yr⁻¹) in models that use the composite rheology, owing to the reduced viscous support of the slab and to weakening and widening of the plate interface. A deeper slab along the eastern boundary (slab Eb₁₁₅) increases the slab-pull driving force, resulting in a more northerly and slightly faster PPM (~N 14° W at 5.3 cm yr⁻¹). For both slab₂₁₅ and slabEb₁₁₅, increasing the PBSZ viscosity from 10¹⁷ to 10²¹ Pa s decreases the speed of the Pacific plate by ~2.5 cm yr⁻¹. Therefore, models with a composite viscosity, a PBSZ viscosity of 10¹⁷ Pa s and either slab shape provide good agreement with observed PPM. More importantly, none of the composite-viscosity models, which have fast mantle flow, produce plate motions that are inconsistent with surface observations.

Subduction of the Pacific plate beneath southern Alaska induces a spatially variable mantle flow and localized flow rates that can be more than ten times the speed of surface plate motions (Fig. 2). In the composite-viscosity flow models, with σₕ = 500 MPa, mantle speeds close to the slab can be up to 90 cm yr⁻¹. These high flow rates occur where strain rates are high (10⁻¹¹ to 10⁻¹² s⁻¹), and viscosities are therefore low (10⁷ to 10⁸ Pa s) owing to the non-Newtonian rheology. For Newtonian viscosity models, mantle flow rates near the slab are 1.0–5.0 cm yr⁻¹. In all the models, far from the slab edge and the driving force of the sinking slab, mantle velocities decrease and are comparable to plate motions.

The pattern of flow in the mantle depends strongly on the slab shape, yield strength and mantle viscosity. The sinking of the dense slab draws flow into the mantle wedge towards the slab, resulting in a poloidal component of flow (Fig. 3). The velocity vectors in the slab and mantle dip more steeply than the slab, which is indicative of steepening of the slab with time or retrograde slab motion. Steepening of the slab pushes material from underneath the slab around the slab edge and into the overlying mantle wedge, forming an anticlockwise toroidal component of flow (Fig. 3). A stronger slab (σₕ = 1,000 MPa) steepens more slowly, generating a proportionally smaller toroidal component of flow. The locus of toroidal flow depends on the depth of the slab edge and the mantle rheology (Fig. 2). The strength of the toroidal component is significantly greater for models with a composite viscosity. In general, the highest velocities occur in those regions of the mantle wedge that are affected simultaneously by poloidal and toroidal flow.

To determine which mantle rheology and slab shape are consistent with observations, infinite-strain axes (ISAs) are calculated from the predicted mantle flow and compared with SKS fast-axis directions indicative of seismic anisotropy (Fig. 4). The pattern of seismic anisotropy is consistent with toroidal flow around the slabEb₁₁₅ edge beneath south-central Alaska and a component of trench-parallel flow beneath western Alaska, assuming A-type LPO fabric in the mantle wedge and B-type fabric in the wedge nose (Supplementary Information). None of the slab₂₁₅ models, in which the toroidal flow is located further to the east, match the anisotropy north of the slab nose; this provides a strong constraint on the slab shape. West of longitude 210°, where slabEb₁₁₅ and slabEb₂₃₀ have the same shape along the mantle wedge, the flow is dominantly trench-perpendicular. However, for models with composite viscosity, the ISAs have regions of trench-parallel orientation that match the observed trench-parallel SKS fast-axis. In this region, stretching occurs as a result of shearing (that is, velocity gradients) in the trench-parallel component of flow due to pressure gradients formed by the westward steepening of the slab. None of the models using the Newtonian viscosity or a higher

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**Figure 2 | Maps of flow field.** a–c, Surface velocity field and viscosity (colour scale) for three models (σₕ = 500 MPa; PBSZ viscosity, 10⁷ Pa s). The observed NUVEL-1A Pacific motion vector, assuming North America to be fixed, is indicated using a white arrow. AVO, Alaska Volcano Observatory (Supplementary Information). d–f, Mantle velocity field at 100-km depth and vertical velocity magnitude (colour scale). The implied strong vertical gradient in velocity illustrates the significant decoupling of the overriding plate from the mantle flow.
and the low-viscosity regions in the mantle wedge and beneath the slab. The low-viscosity regions correlate with regions of high strain rate. b, Isosurface of viscosity showing an oblique, cross-sectional, radial slice through the velocity field. The cross-section BB’ shows poloidal flow and along-strike flow. The plan view shows anticlockwise toroidal flow and an upward component of flow east of the slab edge.

Figure 3 | 3D mantle flow field and viscosity structure. Calculated using the model with slab_{E115} and composite rheology (\(c = 500 \text{ MPa}; \text{PBSZ viscosity, } 10^{20} \text{ Pa s}\)). Figures show subset of modelled domain. a, Isosurface and cross-sections (AA’ and BB’) through composite viscosity show the strong slab and the low-viscosity regions in the mantle wedge and beneath the slab.

yield strength (1,000 MPa) are able to match the overall pattern of seismic anisotropy.

The broad spatial distribution of seismic anisotropy observations provides strong constraints on the slab shape, mantle rheology and slab yield strength, and comparison with the surface plate motions demonstrates that rapid mantle flow, which is strongly decoupled from the surface plates, is consistent with surface observations. We find that for the Alaska subduction zone, both observations are best fitted by a single model using slab_{E115} and a composite viscosity with a yield strength of 500 MPa and a PBSZ viscosity of 10^{20} Pa s. The difference between the surface motion and the sinking rate of the subducting plate requires strain within the plate. This is primarily accommodated by yielding within the slab hinge, which can reduce the effective viscosity from 10^{24} to 10^{22} Pa s and is more significant in models with a composite viscosity, because the non-Newtonian viscosity weakens the outer portions of the slab and the surrounding mantle, requiring more of the slab weight to be supported by the slab strength.

The depth of the Alaskan slab and its predicted steepening are consistent both with the transient state of slab evolution found before a slab reaches the more viscous lower mantle in time-dependent subduction models\(^{16–18,22}\) and with global observations, which show that slab dip increases with slab length for slabs in the early stages of subduction\(^{22,27}\). Geochemo constraints on mantle flow above the short slab in Costa Rica suggest a trench-parallel flow component 50% greater than the convergence rate\(^6\). Thus, although localized high velocities may not occur in all subduction zones, short slabs may be in a transient state characterized by slab steepening, localized fast mantle flow rates and a minimum in viscous resistance to sinking.

Rheologically controlled decoupling of plate motions from rapid mantle flow in subduction zones has implications for a range of processes and for developing a more dynamic understanding of plate tectonics. Rapid transport through the mantle wedge implies that material spends little time in the active melt zone beneath the arc and therefore experiences only small degrees of melting. Locally, high strain rates and complex 3D flow at the slab edge should form regions of efficient mixing of geochemical signatures, and the significant component of upwelling along slab edges may contribute to adakitic volcanism (for example, the Wrangell volcanics in southern Alaska\(^{24}\)). Because the mantle flow is fast (for example, 90 cm yr\(^{-1}\)), even a small slab–parallel component of flow (10–20%) can result in apparently rapid (9–18 cm yr\(^{-1}\)) transport of geochemical signatures along the arc\(^{25,26}\). High strain rates around the slab can lead to more rapid development of LPO\(^1\), allowing changes in mantle flow on shorter length scales to be tracked\(^2\). Finally, partial decoupling of the mantle flow and slab-pull force from surface plates as a result of non-Newtonian viscosity and localized internal slab deformation may explain why 50% of present-day global plate motions can be accounted for without a direct contribution to slab pull from upper-mantle slabs\(^1\).
METHODS SUMMARY

Surface plate motions and solid-state creep within the models are driven by thermally induced density anomalies prescribed by the initial thermal structure. The geometry and thermal structure of the Pacific plate, the Pacific slab and the North American plate are constrained by geological and geophysical observations, thereby reflecting the observed variability in plate age and slab dip (Supplementary Information). We use the finite-element code CitcomCU\textsuperscript{26,27} to solve for the instantaneous solid-state flow of the lithosphere and mantle. The flow is governed by the conservation of mass and momentum for an incompressible, infinite Prandtl fluid, and a viscoplastic constitutive equation, which links the density and viscosity through the thermal field. We use a strain-rate-dependent viscosity assuming the experimentally determined composite rheology for olivine (Supplementary Information). Diffusion creep is assumed for the lower mantle. To resolve the variable slab dip and overriding plate as well as solve accurately for the large gradients in viscosity and velocity, the models contain 100,413,929 mesh nodes with mesh resolution of up to ~2.35 km. The highest resolution coincides with the corner of the plate boundary, which is located far from the model boundaries. We calculate the ISA orientations\textsuperscript{23,24}, which provide a good approximation to the olivine LPO throughout much of the subduction-zone region of the models, as indicated by lag parameter magnitudes less than one. See Supplementary Information for further details.

Full Methods and any associated references are available in the online version of the paper at www.nature.com/nature.

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Author Contributions Both authors contributed equally to the overall development of the project, model design considerations, analysis and interpretations. M.A.J. performed all of the numerical modelling, except for the ISA calculations, which were done by M.I.B.

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METHODS
Model set-up. We use the finite-element code CitcomCU28,29 to solve for the instantaneous solid-state flow of the lithosphere and mantle. Models of the subduction–transform plate boundary in southern Alaska contain an overriding plate (the North American plate), a subducting plate (the Pacific plate) and the underlying mantle. A 3D low-viscosity shear zone separates the two plates (see below). We assume that the Wadati–Benioff zone represents the shape of the subducting lithosphere. The thermal structure of the subducting plate is defined using a half-space cooling model and lithosphere age26,30. For the subducted portion of the Pacific plate, the age of the plate at the trench is projected down-dip. To account for warming of the slab with depth due to subduction into the mantle, the half-space cooling thermal field is adjusted, using a length-scale diffusion analysis. For the overriding plate, effective plate ages are used. For simplicity, we do not include an additional density anomaly for the subducting Yakutat terrane. Perpendicular distances between the model grid points and the slab surface are calculated for each node in the half-space mesh, allowing the initial 3D thermal and weak-zone fields to be smoothly mapped onto the finite-element grid. The mesh spans 961 × 649 × 161 nodes in the longitudinal (185–240°), latitudinal (45–72° N) and radial (0–1,500 km) directions, respectively; mesh resolution ranges from 0.04° to 0.255°. The thickness of the viscous layer at least 40 km thick and extending to a depth of approximately 25 km in the respective directions. The highest resolution coincides with the corner of the plate boundary, which is located far from the model boundaries. All regions of all model boundaries are free-slip. The models were run on 360 processors. See Supplementary Information.

Rheology. The composite rheology includes the diffusion (d) and dislocation (ds) creep deformation mechanisms for olivine

\[
\eta_{\text{com}} = \eta_d/\eta_{\text{ds}}
\]

where we assume that both deformation mechanisms accommodate deformation at a constant stress and that the total strain rate is a linear combination of the contributions from the two mechanisms31,32. The component viscosities are defined by

\[
\eta_{d,ds} = \left( \frac{d^P}{AC_{\text{eff}}} \right)^{1/\nu} \~\exp \left( \frac{E + PV}{nRT} \right)
\]

where \(P\) is the lithostatic pressure, \(R\) is the universal gas constant, \(T\) is the temperature, \(d\) is the strain rate, \(A\) is the grain size, \(C_{\text{eff}}\) is the water content, and \(\lambda, \eta, p, v, E, V\) are experimentally constrained flow-law parameters6. The grain size in the upper mantle is set to 10 mm, such that the respective viscosities predicted by the two mechanisms are equal for a strain rate of \(10^{-15}\text{ s}^{-1}\), with a value of \(10^{20}\text{ Pa s}\) at 250 km for a water content of 1,000 p.p.m. H2O. In the lower mantle, the grain size is set to 40 mm, giving a tenfold increase in viscosity between the upper and lower mantle. At low temperatures and pressures, plastic deformation occurs32; the effective viscosity, \(\eta_{\text{eff}}\), is therefore limited by a depth-dependent yield stress, \(\sigma_r\). If \(\sigma > \sigma_r\), then \(\eta_{\text{eff}} = \eta_d/\eta_{\text{ds}}\), and if \(\sigma < \sigma_r\), then \(\eta_{\text{eff}} = \eta_{\text{com}}\), where \(\sigma\) is the stress and \(\eta_{\text{ds}}\) is the second invariant of the strain rate tensor. For a gradient of 15 MPa km⁻¹, \(\sigma_r\) increases linearly from 0.1 MPa at the surface to a maximum value of either 500 MPa or 1,000 MPa. The models allow for viscosity variations of up to seven orders of magnitude (\(10^{12}–10^{14}\text{ Pa s}\)).

Plate boundary shear zone. We model the boundary between the plates as a thin viscous layer at least 40 km thick and extending to a depth of approximately 90 km (Supplementary Information). The viscosity in the thin viscous layer, \(\eta_{\text{wkb}}\), is smoothly blended into the background viscosity, \(\eta_{\text{eff}}\), using

\[
\eta_{\text{wkb}} = \left( 1 - A_2 \log_{10}(d/\text{slab}) \right) \eta_d
\]

where \(A_2\) is a scalar weak-zone field defined a priori with values ranging from 0 to 1 (corresponding to unweakened and fully weakened regions, respectively) and \(\eta_d\) is the reference viscosity, equal to \(10^{20}\text{ Pa s}\). The \(\eta_{\text{wkb}}\) value is overwritten if the viscosity calculated for \(\eta_{\text{eff}}\) is lower, such that the final form of the viscosity, \(\eta_f\), becomes

\[
\eta_f = \min(\eta_{\text{ds}}, \eta_{\text{wkb}})
\]

Within the multi-resolution finite-element mesh, the thin viscous layer spans a specified minimum number of elements rather than a specific width. This allows us to constrain the viscosity jump across the elements, which has been shown to be important for convergence in models with large viscosity variations33,34. We found good convergence using eight elements to span four orders of magnitude change in viscosity. This implementation was also tested on 3D models with a simplified subduction zone geometry35.

ISA orientations. Although mantle velocity fields from geodynamic models have been compared with the azimuth of the fast seismic velocity36, this is strictly valid only for simple shear deformation and after a sufficient amount of strain has accumulated to align the axis of maximum finite strain with the shear plane37. However, in regions with rapidly changing flow, stretching axis orientations38 or full tracking of LPO development is required27. We calculate the lag parameter, \(\Pi\), which is the ratio of \(\Omega_{\text{iso}}\) to the rate at which the LPO forms along the ISA direction, to \(\Omega_{\text{flow}}\), the rate at which the ISA rotates in response to flow pattern39 (Supplementary Information). The parameter \(\Omega_{\text{flow}}\) is the sum of a spatial contribution related to the instantaneous flow and a time-dependent contribution related to non-steady-state flow. Because the models are instantaneous, we account for the time-dependent contribution to \(\Omega_{\text{flow}}\) by assuming that it is equal in magnitude to the spatial contribution, as has been shown to be true for density-driven mantle flow models9 (Supplementary Information). For lag parameter values of less than one, the ISA gives a robust indication of the LPO37. We expect that for lag parameter values of less than two, the calculated ISA still provides a better indication of the fabric orientation than does mantle flow direction. In comparing the ISAs with SKS fast-axis directions, we assume that A-type fabric develops in the mantle below and near the slab and in most of the mantle wedge. However, within the cold, innermost corner of the mantle wedge, trench-parallel fast seismic directions may be due to B-type fabric20.