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The role of along-strike variations in interface rheology on subduction dynamics: constraints from three-dimensional numerical models

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SUMMARY

The strength of the subduction interface plays a major role in controlling subduction dynamics on both local and global scales. While previous studies have primarily examined interface strength in 2-D models, natural subduction zones are inherently 3-D, with interface strength varying along-strike due to spatial differences in factors such as sediment input. Here, we use the geodynamic code ASPECT to conduct fully dynamic 3-D subduction models in which interface strength varies along-strike. We find that the interaction between strong and weak segments of the interface leads to a narrower range of convergence velocities while broadening the range of viable interface stresses compared to 2-D or homogeneous 3-D models. Stronger segments, when adjacent to weaker ones, exhibit increased convergence velocities. This promotes higher interface stresses and facilitates the subduction of otherwise stagnant strong segments. We find that the interface viscosity of the strong segment controls the baseline stress, whereas the viscosity contrast along-strike controls the magnitude of amplification of the stress due to velocity increases. The elevated interface stresses at strong segments also generate greater compressional forces in the overriding plate than expected from 2-D models. Combined with along-strike variations in convergence velocity, this results in trench migration, with stronger segments displaying more advanced trench positions relative to weaker segments. Possible natural analogues include the Bolivian Orocline in the central Andes and the Lesser Antilles, both of which show enhanced overriding plate compression and trench advance in areas of reduced sediment supply.

Key words: Dynamics of lithosphere and mantle; Rheology: crust and lithosphere; Subduction zone processes.

1 INTRODUCTION

The subduction of negatively buoyant lithosphere is fundamental to plate tectonics, driving plate motions, arc volcanism, crustal recycling and hazardous megathrust earthquakes. Much of our understanding of subduction dynamics comes from geological and geophysical observations and foundational 2-D modelling studies, which have provided fundamental insights into processes such as slab rollback, mantle flow and subduction zone deformation (e.g. S. Zhong & M. Gurnis 1995; J. Toth & M. Gurnis 1998; C.P. Conrad & B.H. Hager 1999; R.J. Stern 2002; C. Doglioni *et al.* 2007). However, 2-D models inherently simplify the system, omitting the along-strike variations now recognized as critical for understanding natural subduction systems. Such variations include differences in subducting plate age, thickness and slab dip (e.g. Nazca; T. Cahill & B.L. Isacks 1992; S.M. Kay & C. Mpodozis 2002; F. Capitanio *et al.* 2011), and laterally limited features such as oceanic plateaus (e.g. J. van Hunen *et al.* 2002; L. Liu *et al.* 2010), mid-ocean ridges (e.g. D.C. Bradley *et al.* 2003) or microcontinental fragments (e.g. J.M. van den Broek & C. Gaina 2020). These heterogeneities strongly influence subduction dynamics, affecting trench location, curvature, slab dip and the compressional forces driving mountain building across the convergent margin (e.g. F. Capitanio *et al.* 2011; J. Rodríguez-González *et al.* 2012, 2016; A. Balázs & T. Gerya 2024).

Another key factor that can vary both with depth and along strike is the mechanical nature of the subduction interface—the weak shear zone that forms between the subducting and overriding plates. This interface can be composed of a variety of materials, including lenses of mafic oceanic crust, ultramafic mantle slivers and sediments of diverse composition and thickness (P. Clift &

P. Vannucchi 2004; G.E. Bebout 2007; M.B. Underwood 2007; P. Agard *et al.* 2018). Its bulk frictional and viscous properties depend on the relative abundances and spatial distributions of different rock types (e.g. J. J. Zhang *et al.* 2006; W.M. Behr & J.P. Platt 2013; P.I. Ioannidi *et al.* 2021; L. Tokle *et al.* 2023; A.L. Abila *et al.* 2024; R. Stoner *et al.* 2025), and on weakening processes such as elevated pore fluid pressures (e.g. M.B. Underwood 2007; D.M. Saffer & H.J. Tobin 2011; T. Sun *et al.* 2020), shear heating (S.M. Peacock 1996) and/or grain size reduction (M.R. Riedel & S.I. Karato 1997; D.M. Fisher & G. Hirth 2024; J. Ruh *et al.* 2024). Interface rheology thus varies dramatically with depth (e.g. due to dehydration reactions and phase transitions with increasing pressure and temperature) and along strike (e.g. owing to variable sediment supply or the subduction of dominantly mafic volcanic features such as ridges or plateaus). Since interface rheology controls shear stress along the subduction interface, it strongly influences convergence velocity and deformation in the overriding plate. While 2-D modelling studies have investigated the effect of interface rheology (e.g. Z. Erdős *et al.* 2021; W.M. Behr *et al.* 2022), and numerous 3-D effects have been modelled (e.g. W.P. Schellart *et al.* 2007; F. Capitanio *et al.* 2011; N.G. Cerpa *et al.* 2021), no study has explicitly explored the influence of spatially variable interface strengths in dynamically evolving 3-D subduction models.

Sediments, in particular, strongly affect interface viscosity and shear stress due to their inherent rheologic weakness relative to mafic oceanic crust (S. Lamb & P. Davis 2003; J. Zhang & H.W. Green 2007; W.M. Behr & T.W. Becker 2018; L. Tokle *et al.* 2023). This weakness arises from both their mineralogical composition—dominated by relatively weak quartz and phyllosilicates compared to the stronger minerals that compose mafic rocks—and their ability to promote high-pore fluid pressures and the formation of metasomatic weak zones during compaction and dehydration (e.g. G.E. Bebout & M.D. Barton 2002; K.I. Hirauchi *et al.* 2016). Although the range of permissible shear stresses for subduction is thought to be relatively narrow (e.g. S. Lamb & P. Davis 2003; S. Lamb 2006; J.C. Duarte *et al.* 2015; A.L. Abila *et al.* 2024; W.P. Schellart 2024), variations within this range due to sediment effects can nonetheless have significant impacts on plate dynamics (S. Brizzi *et al.* 2021; W.M. Behr *et al.* 2022; A.E. Pusok *et al.* 2022; A.L. Abila *et al.* 2024). For example, changes in convergence velocity driven by sediment subduction have been proposed at both the regional (e.g. India–Eurasia collision; W.M. Behr & T.W. Becker 2018) and global scales (S.V. Sobolev & M. Brown 2019).

An additional consequence of sediment-induced weakening is its effect on stress transfer to the overriding plate from the slab. The degree of stress transmitted—and hence the extent and type of overriding plate deformation—is closely linked to the shear stress along the interface, with higher interface shear stresses typically correlating with greater compressional stresses (e.g. S. Lamb 2006; A. Dielforder *et al.* 2020). In systems where sediment input reduces the interface viscosity, the associated drop in shear stress is likely to lead to diminished stress transfer and in turn, less compressional deformation of the overriding plate. This effect can be large enough to promote back-arc extension (Z. Erdős *et al.* 2021). Conversely, when the interface is dominated by a mafic composition with a viscosity on the order of, for example, 10^{21} Pa·s, higher shear stresses can accumulate, leading to greater stress transfer into the overriding plate and hence elevating overriding plate compressional deformation (e.g. S. Lamb & P. Davis 2003) or potentially even stalling subduction (W.M. Behr *et al.* 2022). While a multitude of factors impact the overriding plate deformation (e.g. plateau subduction, L. Liu *et al.* 2010; trench curvature, N.G. Cerpa *et al.* 2021; or plate ages, F. Capitanio *et al.* 2011), natural examples such as the Andes and Lesser Antilles also highlight how along-strike variations in sediment input and effective interface properties may lead to localized differences in overriding plate topography and compressional stress state. For instance, in the Andes, in conjunction with an extensive trench length (e.g. W.P. Schellart 2024) and along-strike changes in plate thicknesses (F. Capitanio *et al.* 2011), decreases in sediment supply along the Bolivian Altiplano region relative to regions along-strike has been suggested to produce localized regions of high-interface shear stress that contribute to uplift (S. Lamb & P. Davis 2003; J. Hu *et al.* 2021). Along the Lesser Antilles subduction zone, the northern and southern segments have undergone different tectonic evolutions since the Eocene (e.g. M. Philippon *et al.* 2020; N.G. Cerpa *et al.* 2021; L. Montheil *et al.* 2023) that may also relate to variations in subducted sediment. During the Eocene, the development of a southward-thickening accretionary wedge (N.L. Bangs *et al.* 2003) coincided with a transition from compression in the north to extension in the southern Grenada Basin (e.g. N.G. Cerpa *et al.* 2021). This phase was subsequently followed by northern extension during the Oligocene (N.G. Cerpa *et al.* 2021). While these cases assume variation in sediment within the subduction channel along-strike, we acknowledge that the ratio of sediment accumulation within an accretionary wedge compared to the input into the interface is not straightforward (e.g. S. Lallemand *et al.* 2024). Regardless, such natural examples highlight the need for modelling approaches that can incorporate along-strike variations in interface rheology as a proxy for varying sediment input.

To systematically explore these effects, we employ fully dynamic 3-D subduction models to investigate how along-strike changes in interface rheology affect subduction zone dynamics. Our models use a constant viscosity interface that is varied from 10^{19} Pa·s, representing a weak sediment-rich interface, to 10^{20} or 10^{21} Pa·s, representing variably mafic interfaces. We examine how along-strike changes in interface viscosity, the spatial arrangement of weak versus strong regions and the length of the overriding plate influence convergence velocity, trench location, overriding plate topography and stress regime.

2 METHODS

2.1 Governing equations

We use the finite-element code Advanced Solver for Problems in Earth's ConvecTion (ASPECT, version 2.5.0-pre; M. Kronbichler *et al.* 2012; T. Heister *et al.* 2017; I. Rose *et al.* 2017; W. Bangerth *et al.* 2019) to simulate 3-D and time-dependent subduction in the presence of along-strike variations in interface strength. ASPECT solves the following conservation equations assuming the Boussinesq approximation with no inertial term or internal heating.

$$-\nabla\cdot(2\eta\cdot\dot{\varepsilon})+\nabla P=\rho g\ , \tag{1}$$

$$\nabla\cdot(u)=0\ , \tag{2}$$

$$\rho\cdot C_{\mathrm{p}}\left(\frac{\partial T}{\partial t}+u\cdot\nabla T\right)-\nabla\cdot k\nabla T=0\ , \tag{3}$$

$$\frac{\partial c_i}{\partial t}+u\cdot\nabla c_i=0\ . \tag{4}$$

Eq. (1) is the conservation of momentum, with the effective viscosity, η, the deviator of the strain rate tensor, $\dot{\varepsilon}$ (defined as $\frac{1}{2}(\nabla u+\nabla u^T)$, the velocity, u, the pressure, P, the density, ρ and gravity, g. Eq. (2) describes the conservation of mass in an incompressible system. Eq. (3) represents the conservation of energy with the specific heat capacity, C_p, the temperature, T and the thermal conductivity, k. Finally, we solve the advection eq. (4) for each compositional field c_i.

2.2 Rheology

The models have a viscoplastic rheology (A. Glerum *et al.* 2018) that includes viscous and plastic deformation without strain weakening. We use solely dislocation creep for the wet quartzite upper crust of the overriding, continental plate (L. Tokle *et al.* 2019). All other compositions including the background material, slab and oceanic crust deform based on a composite olivine flow-law with dislocation and diffusion creep (G. Hirth & D. Kohlstedt 2003). Both diffusion and dislocation creep flow laws are given as

$$\eta^{\mathrm{diff|disl}} = \frac{1}{2} A_{\mathrm{diff|disl}}^{-\frac{1}{n}} d^m \dot{\varepsilon}_e^{\frac{1-n}{n}} \exp\left(\frac{E_{\mathrm{diff|disl}} + PV_{\mathrm{diff|disl}}}{nRT}\right) \qquad (5)$$

with the scalar pre-factor, A, the grain size, d, the second invariant of the deviatoric strain rate, $\dot{\varepsilon}_e$, the activation energy, E, the pressure, P, the activation volume, V, the gas constant, R, the temperature, T and the stress exponent, n. For diffusion creep, $n = 1$ and there is no strain rate dependence. For dislocation creep, the grain size exponent, m, is 0 and so it is independent of grain size. Values for A, E, V and n used in our models are composition-dependent and based on experiments (Table S1, Supplementary Material). The scalar pre-factor for background material is set to give $\eta^{\mathrm{diff}} = \eta^{\mathrm{disl}} = 4 \times 10^{20}$ Pa·s at a depth of 330 km and a transition strain rate of $5 \times 10^{-15}\ \mathrm{s}^{-1}$ (cf. M.I. Billen & G. Hirth 2005). The effective viscosity is computed as a harmonic average of the dislocation and diffusion creep viscosity above the 660-km transition. We set the lower mantle to deform solely via diffusion creep (S.I. Karato & P. Wu 1993).

For plastic parameters we use the Drucker–Prager yield criterion (R.O. Davis & A. Selvadurai 2002) with a friction angle of 5° and cohesion of 10 MPa within all compositions. These values were chosen to approximate the yield stresses associated with the Byerlee Law implementation in W.M. Behr *et al.* (2022) for comparability. The Drucker–Prager yield stress is then computed as

$$\sigma_y = \frac{6C\cos\phi}{\sqrt{3}(3-\sin\phi)} + \frac{6P\sin\phi}{\sqrt{3}(3-\sin\phi)}, \qquad (6)$$

where C is the cohesion and ϕ the internal angle of friction. The effective viscosity is determined by whether viscous stresses exceed plastic stresses; where this occurs, viscosity is determined by the following equation.

$$\eta_{\mathrm{eff}} = \frac{\sigma_y}{2\dot{\varepsilon}_e} \qquad (7)$$

Alternatively, in regions where viscous stress are below the plastic yield stress, the effective viscosity is a harmonic average of the dislocation and diffusion creep viscosities:

$$\eta_{\mathrm{eff}} = \frac{\eta_{\mathrm{diff}} * \eta_{\mathrm{disl}}}{\eta_{\mathrm{diff}} + \eta_{\mathrm{disl}}} \qquad (8)$$

2.3 Model setup

We designed thermomechanical subduction models using ASPECT that are similar to recent 2-D (A.F. Holt & C.B. Condit 2021; W.M. Behr *et al.* 2022) and 3-D models (V. Turino & A.F. Holt 2024). Here, we adopt the 3-D approach. We use a box geometry with dimensions of 4480 (X), 3520 (Y) and 1280 km (Z; Fig. 1). The top and bottom boundaries are set to fixed temperatures of 273 and 1573 K, respectively. All the side boundaries are set as insulating. All mechanical boundaries are free-slip, except at the model surface, which is implemented as a free surface with an additional diffusion component (1.6 m yr^{-1}) for mesh stability (e.g. M. Pons *et al.* 2022; A.G. Grima & T.W. Becker 2024).

2.3.1 Model initial conditions

The model contains distinct overriding and subducting plates, as defined by half-space cooling thermal profiles corresponding to ages of 55 and 80 Myr, respectively (thermal diffusivity of $10^{-6}\ \mathrm{m^2\,s^{-1}}$). The uppermost 30 km of the overriding plate represents the upper crust and is compositional buoyant with a reduced reference density (3000 $\mathrm{kg\,m^{-3}}$ relative to 3300 $\mathrm{kg\,m^{-3}}$ for the background/mantle material). On the subducting plate, the top 12 km represents the interface, parametrized as a generic viscous crustal/metasedimentary layer that is slightly compositionally buoyant (3175 $\mathrm{kg\,m^{-3}}$) and set to a constant viscosity (10^{20} Pa·s) until a depth of ∼175 km, where it transitions into background material (e.g. W.M. Behr *et al.* 2022). The chosen interface density lies between typical values for compacted sediments (2700–2800 kg $\mathrm{m^{-3}}$) and high to ultra-high pressure metasedimentary rocks (3100–3400 $\mathrm{kg\,m^{-3}}$; e.g. H.J. Massonne *et al.* 2007), and thus implicitly averages over the moderate densification expected as subducted sediments undergo progressive metamorphism, which we do not model explicitly.

Initially, the edges of both the overriding and subducting plates are 1000 km from the model sidewall boundaries, allowing toroidal flow around the sides of the slab (Fig. 1; e.g. J. Dvorkin *et al.* 1993; F. Funiciello *et al.* 2003; C. Kincaid & R. Griffiths 2003; D.R. Stegman *et al.* 2010). At the surface, the plates are 1520 km wide along-strike and 1050 and 1420 km in across-strike length for the overriding plate (OP) and slab, respectively. At the start of the model, subduction is initiated by allowing an initial portion of the slab, or protoslab, to extend to a depth of 200 km with a 245 km radius of curvature. Between $x = 1500$ and 1000 km, the 80 Ma subducting plate thins to zero age to represent a mid-ocean ridge. On all sides, surrounding the plates is a 50 km (in depth and width) weak zone differentiated by a reduced friction angle and cohesion (4° and 5 MPa, respectively). This setup facilitates self-consistent subduction initiation due to the negative buoyancy of the protoslab and the decoupling of the subducting plate from the surround thermal boundary layer by the weak trailing edge (ridge) and sides (weak zones). At a depth of 660-km, we impose a 10× increase in the diffusion creep viscosity in order to represent the transition into the lower mantle; however, due to the lack of dislocation creep in the lower mantle, the effective viscosity increase can range from ∼10–50×, broadly inline with geoid constraints (e.g. B.H. Hager 1984). In models with along-strike variations in interface strength, the weak versus strong compositions linearly transition from one to another over 50 km along-strike. Larger transition lengths of 150 and 450 km were tested, however they did not significantly affect the results (Fig. S1, Supplementary Material).

2.3.2 Model resolution and walltime

Model resolution varies through time due to ASPECT's adaptive mesh refinement (AMR) functionality and is set to be spatially and compositionally dependent. Our AMR settings result in a maximum resolution of 5 km that is confined to the interface at depths less

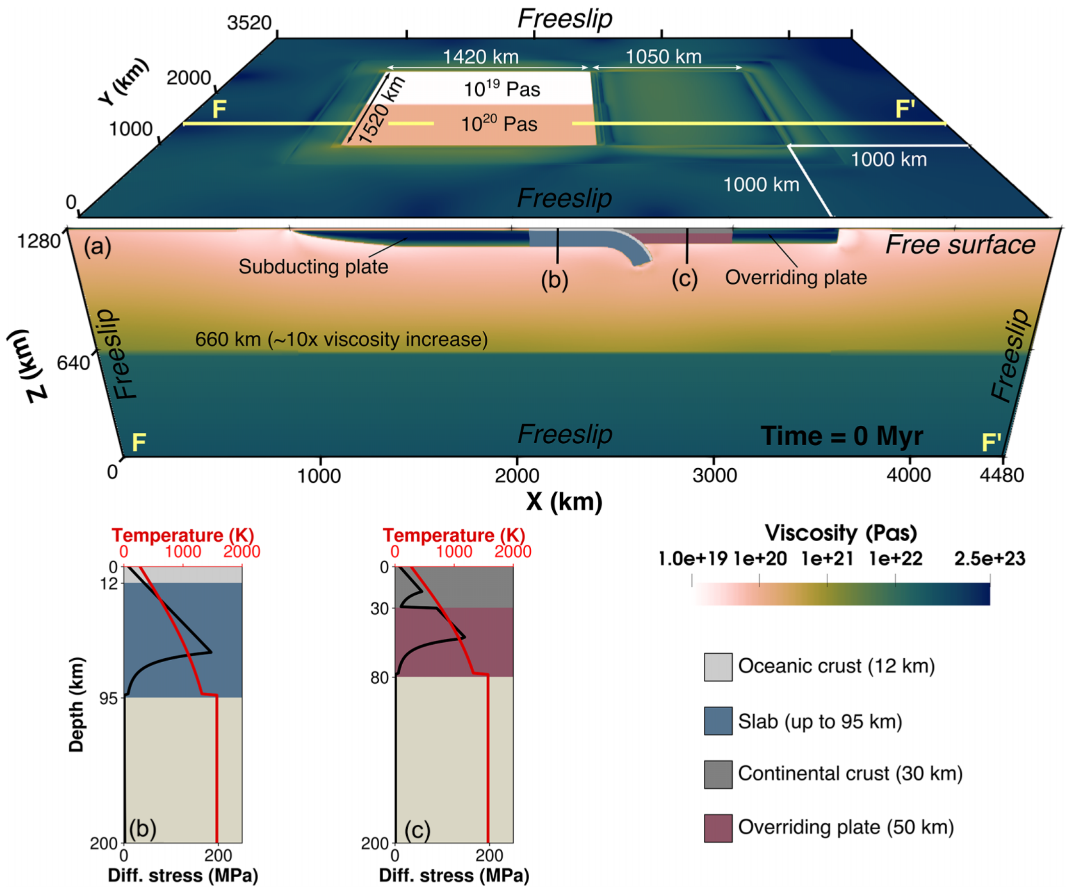

Figure 1. Model setup for case with along-strike variation in interface viscosity. Panel (A) shows the surface overlying a slice taken from F to F', coloured by viscosity and composition. Panels (B) and (C) show strength and temperature profiles of the subducting and overriding plate, respectively.

than 175 km. The overriding plate has a fixed resolution of 10 km, and the slab resolution varies from 10 to 20 km. To resolve plate edges, a temperature isosurface of 1225–1335 K is fixed to a 10 km resolution. The remainder of the model can vary in resolution from 80 to 20 km. Models had ∼ 150 million degrees of freedom (∼ 1.4 millions cells) and were run on the ETH Euler cluster; on 512 cores (AMD EPYC 9654) and with ASPECT's GMG solver (T.C. Clevenger & T. Heister 2021), they had walltimes between 4 and 70 hr, depending on the interface viscosity.

2.4 Model analysis

We ran 15 models to investigate the effects of along-strike changes in interface strength (Fig. 2). Subduction interfaces can consist of multiple different rheologies from weak sediment (e.g. P. Vrolijk 1990; H.J. Tobin & D.M. Saffer 2009) to strong mafic material, leading to a wide range of probable bulk viscosities (W.M. Behr & T.W. Becker 2018). To account for this variation, we use constant viscosity interfaces of either 10^{19}, 10^{20} or 10^{21} Pa·s. Our reference cases have homogeneous interfaces at these viscosities (Section 3.1). These models are compared to cases where there is an along-strike split of the interface into a weak and a strong half (e.g. Fig. 1), which are similarly varied between viscosities of 10^{19}–10^{21} Pa·s (Section 3.2). Section 3.3 investigates the along-strike location of the interface change, with a stripe of weak or strong interface instead placed in the centre of the subducting plate. Finally, in Section 3.4, we change the trench-perpendicular length of the overriding plate to explore the effect of the overriding plate width-to-length aspect ratio.

In analysing the models, we focus on 6 model output parameters: (1) convergence velocity, (2) average interface stress, (3) overriding plate compressional forces, (4) area of overriding plate with a thrust faulting regime, (5) overriding plate rotation and (6) along-strike differences in trench location. To analyse these parameters we extract and analyse 2-D trench-perpendicular model slices up to 20 Myr of model time. As the location of the along-strike change in interface rheology may shift both between the model runs, slices are taken an along-strike distance of ∼250 km from the change to ensure consistent comparisons. Velocities are averaged over constant 15-km depth profiles. Overriding plate forces, thrust faulting area and interface stress are determined within the OP or interface, respectively, with the former two being integrated over the plate thickness and from the start of the overriding plate up to 500 km into it; this limit is set to avoid including values at the trailing edge of the plate that are not related to trench dynamics. In calculating the OP forces, we only include trench-perpendicular compressional forces greater than 1 MPa; overall, this analysis method gives us a general idea of the net OP compressional force without constraining where deformation would be focused or whether other areas of the plate may be extensional (see Text S1 in the Supplementary Material for more details). The faulting regime is determined using a discontinuous stress state based on the compressive principal stresses (M.L. Zoback 1992), although we do not consider transitional or strike-slip regimes.

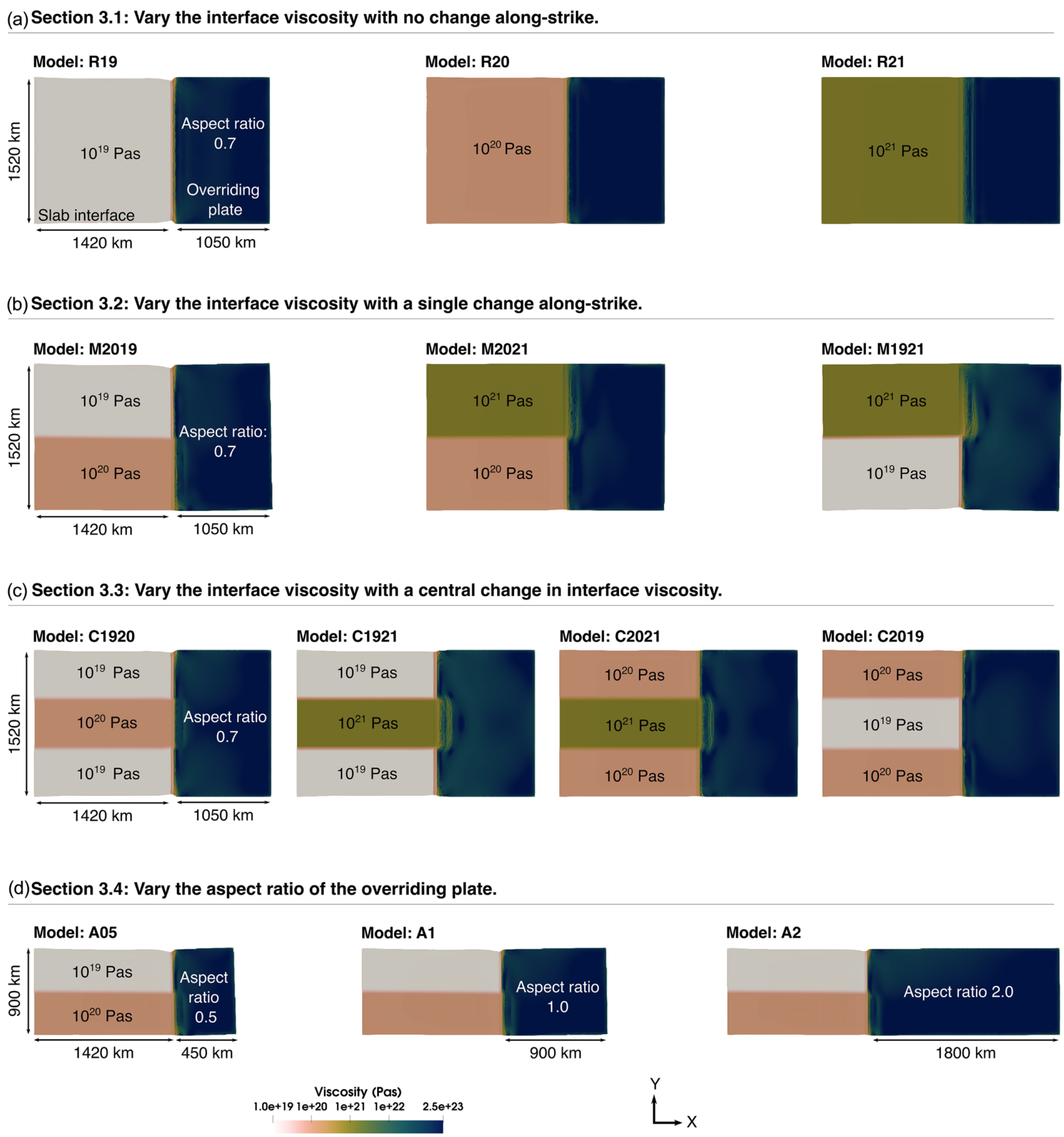

Figure 2. List of models by section to show the initial setup and what is varied, coloured by the viscosity. Models are visualized through a composition contour, showing only the slab interface and overriding plate at the surface at 0.5 Myr. Models are given a letter by section. For panels (A)–(C), model name numbers are assigned by the interface viscosity. For panel (D), model numbers are assigned according to the overriding plate aspect ratio.

3 RESULTS

3.1 Homogeneous models

3.1.1 Reference homogeneous model evolution

To investigate effects of along–strike changes in interface strength, we first ran three reference models with homogeneous interface viscosities of 10^{19}, 10^{20}, and 10^{21} Pa·s (models R19, R20, and R21). Here, we describe the evolution of model R20 using three trench-perpendicular slices (Fig. 3; Video S1, Supplementary Material). The model evolution follows three distinct phases: (1) build up to peak convergence velocity (free-sinking phase), (2) decrease in velocity as the slab interacts with the 660-km mantle viscosity increase and (3) post-660 km slab interaction (cf. F. Funiciello *et al.* 2003, 2006).

During the first phase, the model has a low-convergence velocity (~3 cm yr^{-1}) that increases through time as more plate subducts and the slab lengthens (Fig. 3a). The overriding plate experiences its highest trenchward velocity at the start of the model run before gradually slowing (Fig. S2, Supplementary Material). Thrust faulting within the overriding plate is widespread during this stage, with an affected area of ~9250 km^2 along a 2-D slice (Fig. 3d),

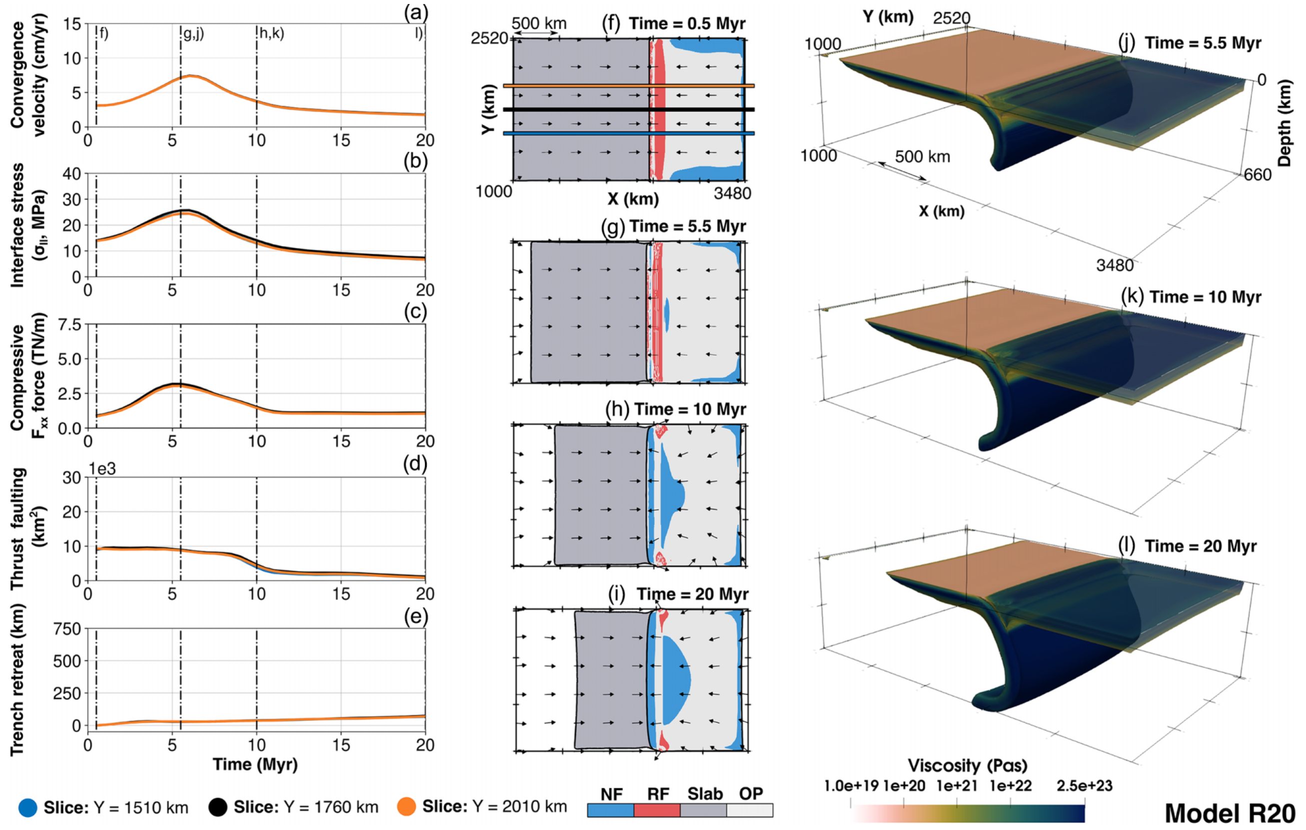

Figure 3. Evolution of the reference 10^{20} Pa·s model without along-strike variations. Panels (a)–(e) show the time evolution of multiple parameters at different slice locations. Panels (f)–(i) show a 15-km depth contour coloured by composition (dark grey for slab and light grey for overriding plate) and the stress regime based on M.L. Zoback (1992) (blue indicates normal faulting and red reverse faulting). Arrows represent velocity direction. Panels (j)–(l) show a 3-D view of the slab and overriding plate coloured by viscosity.

predominantly focused near the plate boundary. As subduction continues, the convergence velocity reaches a maximum of 7.2 cm yr^{-1} at ∼6 Myr. The average interface stress (Fig. 3b) and compressional force (Fig. 3c) within the overriding plate increase in tandem but reach their peak values (26 MPa and 3.2 TN m^{-1}, respectively) slightly earlier, at ∼5.5 Myr (Figs 3g and j). Near this time, discrete thrust faults develop in the forearc basin of the overriding plate. By the end of this phase, the slab has reached ∼450 km depth and exhibits a convex shape towards the mantle wedge side.

At ∼6.5 Myr, the model enters its second phase, where convergence velocity, interface stress and OP compressional forces begin to decline. While the area of thrust faulting remains large, it starts to decline at a faster rate. As the slab reaches the 660-km transition at ∼10 Myr (Figs 3h and k), it undergoes further bending and begins to buckle near the transition to the lower mantle. Around this time, the overriding plate reaches a minimum velocity of ∼0.1 cm yr^{-1}, coinciding with a pronounced decline in the area of thrust faulting.

In the final phase, following the interaction with the 660-km viscosity transition, slab rollback occurs leading to renewed overriding plate motion and a largely extensional OP faulting regime. During this phase, all of the extracted parameters continue to decline, excluding the OP horizontal compressional forces, which remain relatively constant. We note that trench retreat occurs throughout the model run, and that the slab maintains a similar shape along-strike (Fig. 3l). Within the model, the radius of trench curvature reduces with trench retreat (cf. W. Schellart 2010), reaching a minimum of ∼8400 km by 20 Myr.

3.1.2 Homogeneous model comparisons

Before investigating the impact of along-strike variations, we ran two models with constant but distinct interface viscosities of 10^{19} and 10^{21} Pa·s. These models demonstrate a systematic relationship between interface viscosity and convergence velocity similar to results from previous 2-D models (cf. W.M. Behr *et al.* 2022). In the low viscosity 10^{19} Pa·s $^{\text{model}}$ (R19), the peak convergence velocity reaches ∼23.6 cm yr^{-1}, much greater than the ∼7.5 cm yr^{-1} within the 10^{20} Pa·s model (R20) previously described. The greater trench retreat in model R19 also leads to a lower radius of trench curvature (∼3300 km; e.g. W. Schellart 2010). In the strongest interface viscosity model of 10^{21} Pa·s (R21), subduction nearly stalls, with a maximum velocity of only ∼0.5 cm yr^{-1}. Differences can also be seen in the total trench retreat after 20 Myr of evolution, which range from 620 km in R19 to 75 and 30 km in R20 and R21, respectively.

The maximum of the average OP compressional force is reduced in models R19 and R21 (1.1 and 1.6 TN m^{-1}, respectively) relative to the that of R20 (∼3.2 TN m^{-1}). Similar to R20, model R19 shows a local peak in the compressional force during the free-sinking phase when convergence velocities reach the maximum (∼1.1 TN m^{-1}). However, despite the overall greater convergence velocities in R19 compared to R20 and R21, compressional forces are generally low during the free-sinking phase when convergence velocities are high, and primarily increases late in the run when most of the initial slab has subducted and convergence velocities are low (Fig. S3, Supplementary Material). Model R21, likely due

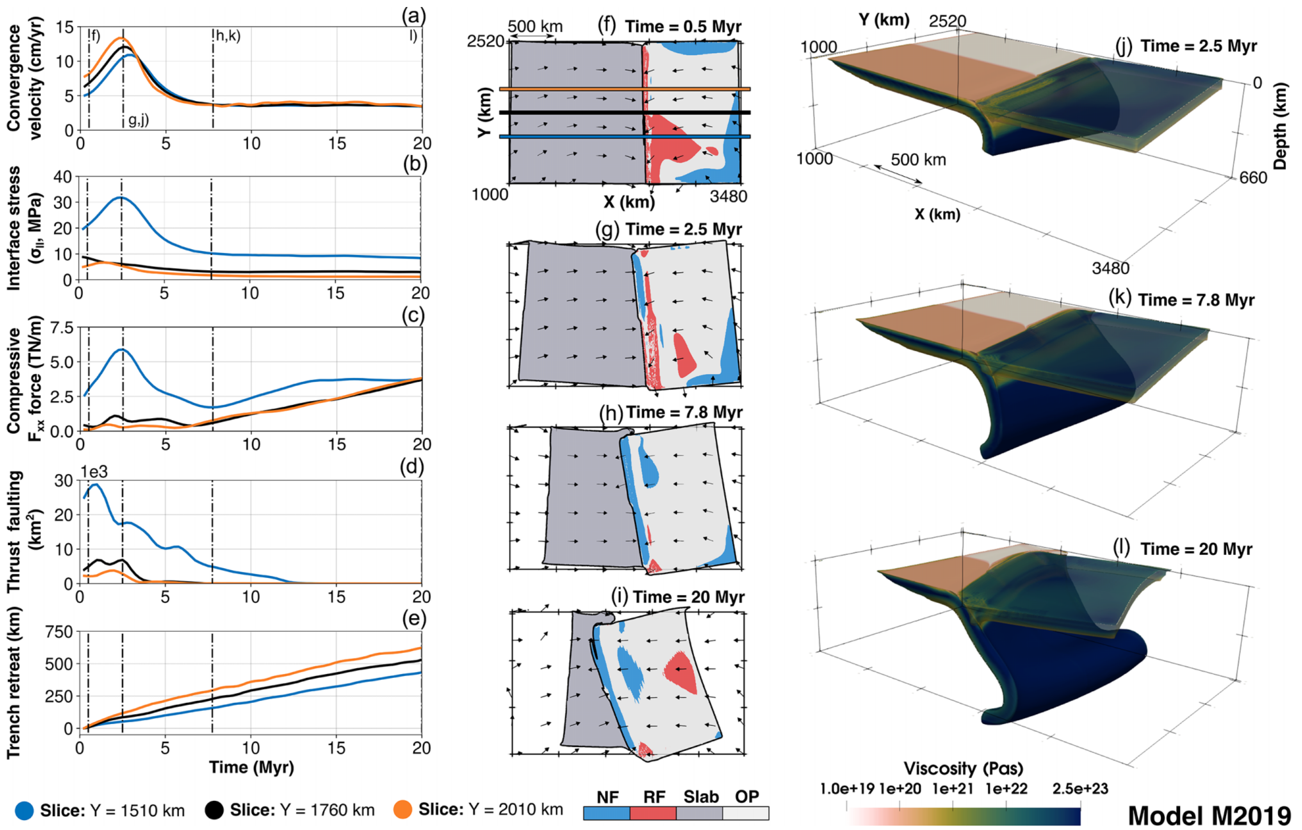

Figure 4. Evolution of the 10^{20} (bottom half) to 10^{19} (top half) Pa·s model without along-strike variations. Panels (a)–(e) show the time evolution of multiple parameters at different slice locations. Panels (f)–(i) show a 15-km depth contour coloured by composition (dark grey for slab and light grey for overriding plate) and the stress regime based on M.L. Zoback (1992) (blue indicates normal faulting and red reverse faulting). Arrows represent velocity direction. Panels (j)–(l) show a 3-D view of the slab and overriding plate coloured by viscosity.

to the stalled subduction, has a consistently high value throughout the model run (∼1.5 TN m^{-1}).

3.2 Varied interface strength along-strike

To examine the impact of along-strike variations in interface strength, here we introduce models in which the interface viscosity is spatially heterogeneous. These models build upon the homogeneous reference cases by incorporating lateral changes in interface viscosity, allowing us to assess how such variations modify convergence velocity, slab morphology and overriding plate stress distribution. We begin by describing a reference heterogeneous interface model where the interface viscosity is set to 10^{20} Pa·s from y of 1000–1735 km, and 10^{19} Pa·s from 1785 to 2520 km (model M2019; Fig. 4). The transition between these two interface properties occurs over a 50 km region in the centre of the model, where the composition linearly shifts from that of the strong to that of the weak material.

3.2.1 Reference varied model evolution

Similar to the reference model without interface strength variations (R20), model M2019 follows three phases: (1) build up to peak velocity, (2) decrease in velocity as the slab reaches the 660-km transition and (3) post-660 km slab interaction (Video S2, Supplementary Material). To illustrate the effects of along-strike variations, we analyse slices taken at 250 km to either side of the viscosity transition, allowing for a direct comparison between strong and weak interface segments.

Phase 1: The initial increase in convergence velocity follows that of the homogeneous reference models but exhibits pronounced along-strike differences. On the weak interface side, convergence accelerates rapidly, peaking at 13.4 cm yr^{-1} by 2.3 Myr (Fig. 4a). In contrast, the strong interface segment reaches a lower peak velocity of 10.9 cm yr^{-1} at 3 Myr. These changes in velocity can partially be attributed to differences in OP velocity (Fig. S4, Supplementary Material), which in these models is broadly equivalent to the trench retreat rate (due to relatively limited OP deformation). This differential trench retreat rate induces a counter-clockwise rotation of the overriding plate and results in a lateral variation in trench position (Fig. 4e). The stress distribution reflects this contrast, with the weak interface section associated with both lower interface stress and lower overriding plate compressional force (Figs 4b and c), while the strong side shows an increase in both properties. By 2.5 Myr, the strong interface segment reaches a peak interface stress of ∼32 MPa and an OP force of ∼5.9 TN m^{-1}. This coincides with high variability in the distribution of thrust faulting along-strike, with the strong interface slice showing a thrust faulting regime that extends much farther into the OP (∼17 400 km^2, Figs 4d and g). During this period of elevated compression, discrete plastic shear zones, mimicking faults, develop near the front of the plate (Fig. 4j). During Phase 1, the radius of trench curvature decreases in both segments, with the strong segment developing a smaller radius. By the end of this phase, there is an ∼65 km along-strike difference in trench position between the two slices, which can be attributed to the OP rotation associated with differential trench retreat rates and overriding plate shortening (Fig. 4g).

Phase 2: At ∼2.5 Myr, the slab segment with the weak interface reaches ∼460 km depth, and the convergence velocity begins to decline. However, the velocity along the strong interface segment continues to increase, peaking slightly later before also beginning to decrease. As velocity decreases, interface stress, the OP force and the area of thrust faulting begin to subside. By ∼5 Myr, the slab on the weak interface side reaches the 660 km transition and anchors into the stronger lower mantle. This initiates elevated slab rollback on the weak interface side and an associated shift towards an extensional OP. Along the strong interface, the slab has not yet reached 660 km, overriding plate speeds are low, and there is a small peak of thrust faulting in the OP (Fig. 4d). At ∼7.8 Myr (Figs 4h and k), the rest of the slab, in the strong segment, anchors into the lower mantle, coinciding with the minimum OP force on that side of the subduction zone. By this time, because of the difference in trench retreat rate along-strike and slab anchoring on the weak interface side, there is an along-strike variation in slab dip, with the strong interface side exhibiting a steeper dip (∼ 53°) that reduces towards the weak interface (∼ 41°). Throughout Phase 2, the radius of trench curvature continues to decline for both segments.

Phase 3: By 8 Myr, extensional stresses develop along the entirety of the OP, and although the compressional forces in the X direction rise, due to stress orientations the area of thrust faulting reduces to less than ∼100 km^2 by 13 Myr (Fig. 4d). The convergence velocities and average interface stress are near-constant during this phase, and the along-strike change in trench location continues to increase due to continued differential trench retreat rates. By the end of the model run, the weak interface side has retreated 190 km more than at the strong interface and exhibits a 6° shallower dip, leading to a highly variable slab shape along-strike (Fig. 4l). The radius of trench curvature continues to decrease in both segments until 13.5 Myr, at which point the strong segment reaches a minimum of ∼1550 km, compared to ∼1950 km along the weak segment. During subsequent slab rollback, the radius in the strong segment begins to increase as the curvature evolves under the influence of trench retreat rather than OP compression (cf. W. Schellart 2010). By the end of the model, the strong and weak segments reach radii of ∼1650 and ∼1250 km, respectively (radius of trench curvature values for all remaining models are provided in Table S2 in the Supplementary Material).

3.2.2 Comparison of models with variable segment viscosities

To further investigate the influence of along-strike viscosity changes, we ran two additional models with weak segment viscosities of 10^{19} and 10^{20} Pa·s, and a strong segment viscosity of 10^{21} Pa·s (models M1921 and M2021, respectively). We discuss the models by comparing each segment to its respective homogeneous reference case, as well as assessing the overall along-strike variability.

The first-order model evolution of M1921 and M2021 are similar to model M2019. On the weak interface side, both convergence velocities and overriding plate compressional forces are lower than in their corresponding reference homogeneous model (Figs 5 and 6). Compared to the stronger side, the weak subduction segments have both higher subduction velocities and greater trench retreat rates, which results in a change in the trench location along-strike and hence rotation of the overriding plate. On the strong interface side, all models show increased OP compressional forces and convergence velocities relative to their reference homogeneous models (Figs 5d and e; Fig. 6). Notably, while the reference 10^{21} Pa·s model (homogeneous plate interface) experiences stalled subduction (cf. W.M. Behr *et al.* 2022), pairing it with a lower-viscosity segment allows subduction to proceed at low velocities while maintaining elevated overriding plate compressional forces.

The differences between models primarily depend on the magnitude of the viscosity contrast along-strike. While lower along-strike averages of viscosity result in higher convergence velocities overall, the OP compressional forces are sensitive to both the viscosity of the strong segment and the viscosity ratio between segments. For example, in two models with a 10× viscosity increase along-srike (M2019 and M2021), the strong segment of model M2019 (10^{20} Pas) shows higher convergence rates but lower average compressional forces than the strong segment of M2021 (10^{21} Pas, 4.6 versus 3.3, Fig. 5e), suggesting the importance of the viscosity magnitude. Although both models reach a similar peak compressional force (∼5.6 TN m^{-1}), the slab within M2021 takes longer to reach the 660-km viscosity jump (7.8 Myr in M2019 versus >20 Myr in M2021), leading to an extended phase of high OP compression. This effect is even more pronounced in model M1921 where the increased viscosity contrast (20×) relative to M2021 (10×) results in the highest compressional stresses. In this case, the strong interface side in M1921 exhibits both higher average convergence velocities (∼1.5 cm yr^{-1}) and greater OP force (∼9.1 TN m^{-1}) compared to M2021 (∼1 cm yr^{-1} and 4.6 TN m^{-1}), indicating that convergence velocity also controls the OP force even if the strong interface viscosity remains constant. These factors collectively contribute to along-strike changes in trench location with strong interfaces associated with trench advance and weak trenches with trench retreat, either through increased overriding plate rotation due to differential trench velocity, or through variations in compressional deformation. We also examined the influence of the amount of weak interface along-strike, which showed that a higher proportion of weak interface further elevated the velocities in the strong segment, but the effect was less pronounced than those explored here (Fig. S5 and Text S2, Supplementary Material).

3.3 Central interface models

While in our previous models we explored a single viscosity contrast along-strike, the relative location of strong versus weak interface segments may also influence model evolution. To investigate this, we ran four additional models where we split the downgoing plate into thirds, varying the viscosity of the central segment of the interface relative to the outer two segments (model C1920, C1921, C2021, C2019). These models (Fig. 7) are compared to the models from Section 3.2 (Fig. 5).

Similar to the models in Section 3.2, the central interface models show increased OP forces at strong interface regions and variations in trench position along-strike. The margin geometry appears to take one of two characteristic shapes: (1) a concave profile with an advanced central region when the central interface is strong (Fig. 7e), or (2) a convex profile with a retreated central trench when the central interface is weak. A notable difference between these models and those in Section 3.2 is less variation in the convergence velocity along-strike. For example, model M2019 exhibits an average along-strike convergence velocity difference between the strong and weak slices of ∼1.2 cm yr^{-1}, whereas C2019 exhibits smaller variation of only 0.2 cm yr^{-1} (Fig. 7a). Compared to the

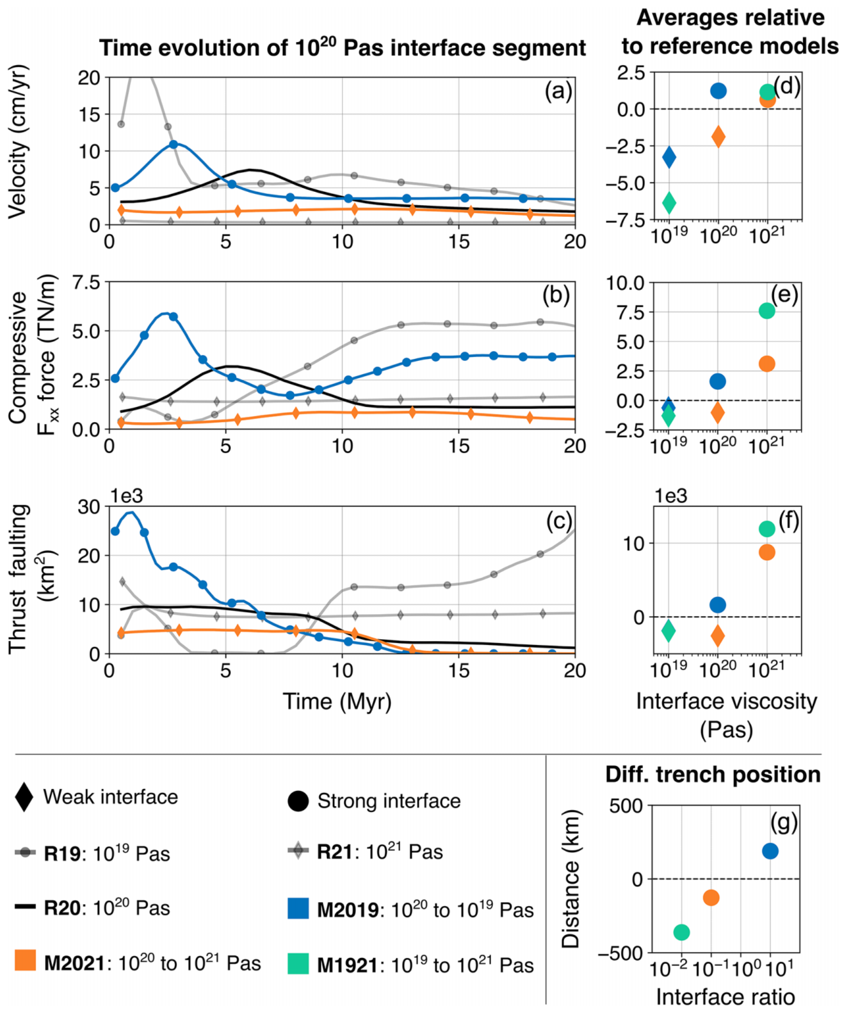

Figure 5. Comparison between models with different interface viscosities along-strike. Symbols indicate whether the model is a slice from the strong or weak interface. Colour indicates the model. Panels (a)–(c) show the time evolution of two models compared to the reference model (black), with both lines indicating the 10^{20} Pa·s model side. Panels (d)–(e) show the same values time-averaged and scaled to the reference model, and additionally show values corresponding to the 10^{19} and 10^{21} Pa·s sides of the models. Panel (g) indicates the difference in trench location for each model along-strike, where the interface ratio refers to the difference in viscosity from the interface at low Y-values to that at high Y-values (e.g. model M2019 from Fig. 4 has an interface ratio of 10^1 because it has a 10^{20} Pa·s interface at lower Y-values and a 10^{19} Pa·s interface at high Y-values) and where a positive value indicates the trench at the front model side is advanced relative to the back.

models in Section 3.2, these central models are pinned on either side of the central segment resulting in less OP rotation (Fig. 7b; Video S3, Supplementary Material) and a greater increase in OP compressional force (Fig. 7c) within the strong interface section. Overall, this results in less pronounced changes in trench location (e.g. 365 and 270 km for M1921 and C1921, respectively).

3.4 Overriding plate aspect ratio

In the previous models, we used a 0.7 aspect ratio of the overriding plate (strike-perpendicular length to along-strike width); however, this ratio is likely to also play a role in overriding plate dynamics due to its impact on the net mantle shear forces, which scale with plate size, acting at the base of the overriding plate. To explore this, we ran three models with an altered geometrical setup, where the model is 4480 × 2400 km in X and Y, with 750 km gaps around the edges of 900 km wide subducting and overriding plates. The trench-perpendicular length of the overriding plate was varied between 450, 900 and 1800 km, corresponding to aspect ratios of 0.5, 1 and 2 (models A05, A1 and A2). These models have an interface viscosity change in the model centre from 10^{20} to 10^{19} Pa·s, as in model M2019.

Increasing the overriding plate length and aspect ratio affects the models in several ways. A larger aspect ratio reduces overriding plate rotation (Fig. 8a), approaching no rotation at an aspect ratio of 2. Unlike the models in Section 3.3, the reduction in rotation is accompanied by a reduction in overriding plate compressional force within the strong interface section, from 1.6 TN m^{-1} with an aspect ratio of 0.5 to 1.3 TN m^{-1} at a ratio of 2 (Fig. 8b). Interestingly, the average convergence velocities on the strong interface side remain relatively unchanged across aspect ratios, with values of 3.6, 3.5 and 3.7 cm yr^{-1} for A05, A1 and A2, respectively. However, due to variation in the net basal shear force on the OP, the partitioning between subduction velocity and OP velocity changes (cf. A.F. Holt & T.W. Becker 2016): as the aspect ratio increases, subduction velocity increases while OP velocity decreases. This occurs because, for larger OPs, subduction via trench retreat becomes more difficult (due to increased basal shear resistance associated with OP

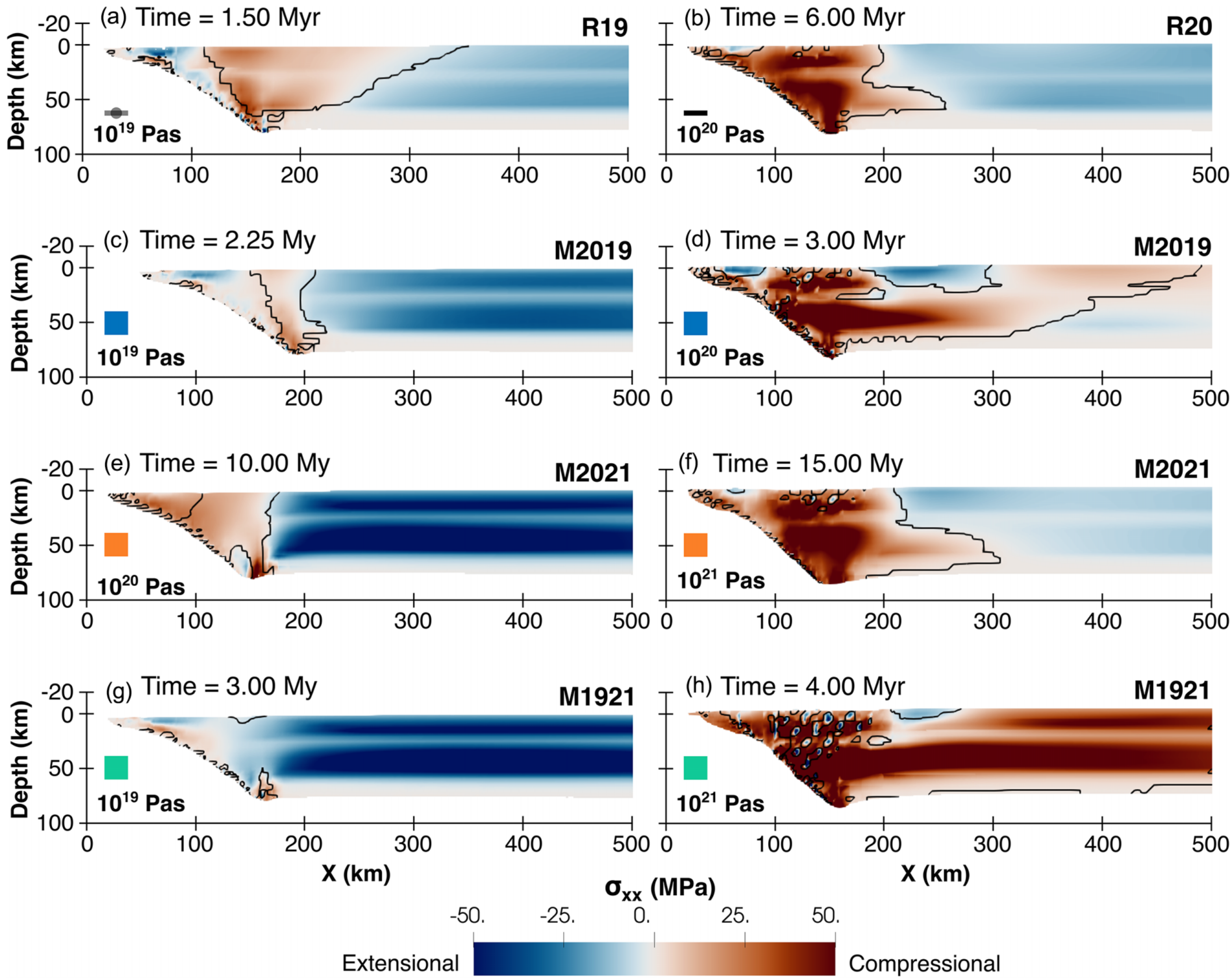

Figure 6. Model slice snapshots of the overriding plates deviatoric stress in X showing the 500 km used in determining the integrated compressive force and area of thrust faulting. Coloured boxes indicate the model with respect to Fig. 4. Positive Sxx represents compressive stresses and negative extensive stresses. The black contour shows the region of thrust faulting. Snapshots are taken at the time of max convergence velocity for each slice.

motion). Compared to model M2019, all the models here exhibit lower OP compressional forces, and reduced convergence velocity on the strong interface (4.8 cm yr^{-1} for M2019). As OP rotation and compressional forces decrease with increasing aspect ratio, along-strike trench position variations also become less pronounced. The difference in the trench X-location along-strike between the two slices reduces from 188 to 58 km at aspect ratios of 0.5 and 2, respectively (Fig. 8c). This suggests that multiple factors influence trench location, including shortening and the relative balance between overriding and subducting plate basal tractions (and hence velocities). The latter of which affects OP rotation and depends on not only interface strength but also relative plate sizes.

4 DISCUSSION

4.1 Effects of along-strike variations and extrapolation from two-dimensional to three-dimensional

The viscosity of the interface can span 2–3 orders of magnitude depending on the distribution of rock types (e.g. W.M. Behr & T.W. Becker 2018; A.L. Abila *et al.* 2024), which itself can vary substantially along-strike (e.g. M.E. MacKay & G.F. Moore 1990; P.B. Kelemen *et al.* 2003; S. Lamb & P. Davis 2003). Several 2-D numerical modelling studies have shown that interface strength strongly influences subduction dynamics: weak interfaces promote slab rollback, increase convergence velocity and reduce slab dip which can inhibit slab penetration through the 660-km phase transition (e.g. C.P. Conrad & B.H. Hager 1999; H. Čížková & C.R. Bina 2013, 2019; W.M. Behr *et al.* 2022).

To compare to these results from 2-D models, we first examine our homogeneous interface strength (R19, R20, R21) models. Our 3-D model results are consistent with the 2-D studies: at low-interface viscosity (10^{19} Pa·s), convergence rates approach ~24 cm yr^{-1} (R19), while at high viscosity (10^{21} Pa·s), subduction almost stalls with velocities below ~0.5 cm yr^{-1} (R21), in line with the 2-D values reported by W.M. Behr *et al.* (2022). We also observe trends in slab geometry that are consistent with earlier work: higher interface strength produces steeper slabs (62° versus 71° in R19 and R20, respectively; H. Čížková & C.R. Bina 2013; W.M. Behr *et al.* 2022) and reduced trench retreat/rollback distances within a 20 Myr time interval (620 versus 70 km in R19 and R20).

When along-strike variations in interface strength are introduced, their effects on convergence rate, slab dip and rollback interact spatially, leading to more complex subduction dynamics. Our models

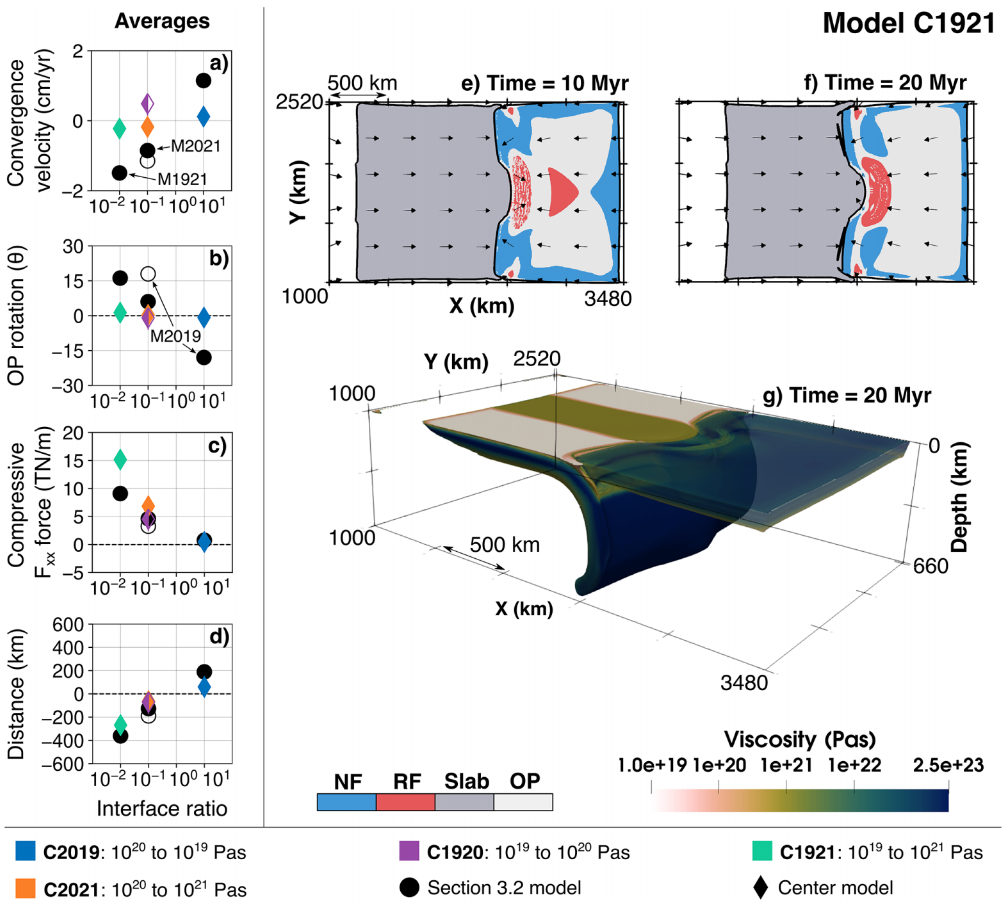

Figure 7. Comparison between models where the change in interface is placed in the centre of the model. Symbols indicate whether it is a centre model (coloured diamond), or a varied model from Section 3.2 (black circles). Shape fill is used to better see overlapping symbols. (a)–(d) Parameters against the along-strike interface viscosity ratio, shown as front to back slice. (a) Differential convergence velocity along-strike. (b) Total overriding plate rotation by model end. (c) Average compressive force in overriding plate. (d) Difference in trench location along-strike. Model M2019 is shown twice, with the hollow circle showing the back to front slice. Panel (d) indicates the difference in trench location for each model along-strike where a positive value indicates the trench at the front model side is advanced relative to the back. Panels (e)–(g) show snapshots of a single model evolution, see Fig. 2 for explanation.

show that these variations affect both local and plate-scale convergence rates. For example, the homogeneous reference model R20 exhibits a peak convergence rate of 7.5 cm yr^{-1}. When paired with a weak interface (M2019), the 10^{20} Pa·s segment velocities increase to 10.9 cm yr^{-1}. Conversely, coupling the same 10^{20} Pa·s interface with a stronger rheology (M2021) reduces convergence velocities to 2.1 cm yr^{-1} (Fig. 5a). This behaviour is particularly interesting in configurations where a very strong interface—such as model R21, which stalls at 0.5 cm yr^{-1} when homogeneous—is instead partially accelerated by an adjacent weak interface segment. In the axial-symmetric model shown in Fig. 7, the weak interface segments cause the entire subduction zone to subduct faster relative to the homogeneous R21 model (up to 2.9 cm yr^{-1} in model C1921), with only minor variations in convergence velocity along-strike (approximately 0.3 cm yr^{-1}). In non-axial symmetric models (e.g. Fig. 4), the slab on the weak interface side subducts faster and penetrates deeper into the mantle. This forms a slab step that increases slab pull along-strike (e.g. M. Zuhair *et al.* 2022), and causes the strong interface to subduct faster (up to 2.6 cm yr^{-1} in model M1921) than the 0.5 cm yr^{-1} convergence rates in the homogeneous R21 case. Faster subduction along the weak interface segment leads to earlier anchoring in the 660-km transition zone, triggering rollback within that section and altering the trench orientation. This along-strike variation in slab behaviour driven by differences in dip, rollback rate, convergence velocity and compressional forces produces complex slab and trench geometries (e.g. Fig. 4l), with strong interface segments exhibiting trench advance relative to weaker ones. These results highlight how along-strike variations in interface rheology can influence subduction dynamics well beyond their local region.

4.2 Interface stress and overriding plate deformation

Overriding plate deformation along subduction margins is primarily driven by (1) normal stresses related to trench motion (e.g. T. Nakakuki & E. Mura 2013; A.F. Holt *et al.* 2015), (2) basal drag (F.A. Capitanio *et al.* 2010; T. Nakakuki & E. Mura 2013; W.P. Schellart & L. Moresi 2013) and (3) interface shear stress (S. Lamb & P. Davis 2003; J. Hu *et al.* 2021; W.M. Behr *et al.* 2022). While all are present in our models, we focus here on the interface shear stress, and specifically the second invariant of deviatoric stress along the interface (as a proxy for interface shear stress which is the main stress component within these models, as shown in Fig. S6 in the Supplementary Material), and how it scales with convergence velocity, interface viscosity and overriding plate deformation regime. In nature, the range of interface shear stresses is not well known and appears to vary over an order of magnitude, depending on both

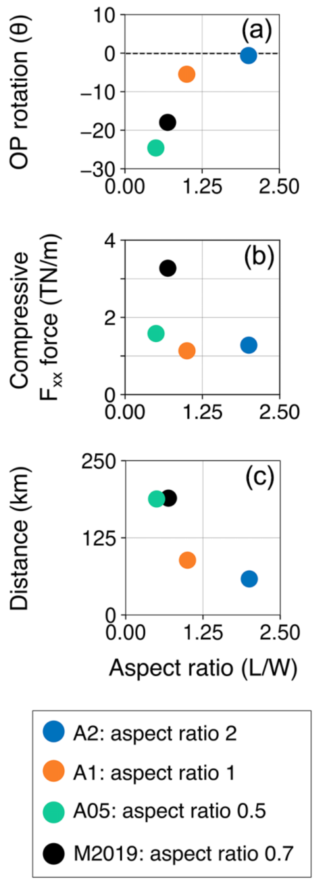

Figure 8. Comparison between models where the length to width aspect ratio of the overriding plate is varied. Colour indicates the model, with the black circle representing M2019 from Section 3.2. Panels (a)–(c) show the (a) total overriding plate rotation, (b) time-averaged compressive OP force and (c) total trench difference between the front and back slice against the aspect overriding plate ratio.

geologic setting and the method or timescale associated with the estimation (e.g. force-balance, S. Zhong & M. Gurnis 1994; paleopiezometry, W.M. Behr & J.P. Platt 2013; heat flow constraints, P. England 2018). While some studies suggest that low long-term viscous interface shear stresses (e.g. less than 35 MPa) are most consistent with global subduction behaviour (e.g. J.C. Duarte *et al.* 2015), others propose that higher values are viable, especially in systems with significant topographic loads (e.g. S. Lamb & P. Davis 2003; S. Lamb 2006).

Previous geodynamic subduction modelling has shown that strong interface viscosities can lead to high-shear stresses that stall subduction (e.g. W.M. Behr *et al.* 2022). This is consistent with our homogeneous interface models; for example, model R21 exhibits an average interface shear stress of 14.6 MPa and undergoes subduction stalling. However, in models that incorporate along-strike variations in interface strength, we observe significantly higher interface stresses than in homogeneous models for equivalent interface strengths, yet subduction still proceeds. This is because, as discussed in Section 4.1, pairing a strong interface segment with a weak segment increases the velocity of the strong side relative to the homogeneous reference case, due to increased slab pull sourced from the deeper, adjacent slab with a weaker interface. This velocity increase results in elevated interface stresses (Fig. S7, Supplementary Material), with the magnitude of stress controlled by both the viscosity of the strong segment and the viscosity contrast along strike. For instance, interfaces with 10^{21} Pa·s viscosity on the strong side exhibit the highest average interface stresses (e.g. C2021; up to 43 MPa compared to a maximum of 26 MPa seen in homogeneous models), and larger viscosity contrasts along strike further amplify these stresses (e.g. C1921; up to 56 MPa). Because 2-D and homogeneous 3-D models cannot capture this interaction, they likely underestimate the range of viable shear stress conditions under natural subduction scenarios. These results suggest that the viscosity of the strong segment controls the baseline stress, while the along-strike viscosity contrast controls how much that stress is amplified, with larger contrasts causing greater velocity increases in the strong segment (e.g. Fig. S7, Supplementary Material).

To assess the realism of our modelled stresses and velocities (Fig. 9), we compare them with natural subduction zone observations. A global compilation by W.P. Schellart & N. Rawlinson (2013) shows that 86 per cent of subduction segments converge between 2 and 12 cm yr^{-1}, a range J.C. Duarte *et al.* (2015) linked to a maximum interface shear stress of ∼35 MPa based on analogue modelling. This suggests sustained subduction requires low-interface strength, which these authors suggested could be due to fluids and consistently weak materials at the interface. Our models align fairly well with the convergence velocity constraints. Prior to slab anchoring (first 700 km of subduction), 70 per cent of our 23 modelled segments fall within the 2–12 cm yr^{-1} range (Fig. 9a). While slightly lower than natural estimates, this includes 22 per cent of segments with a high-interface viscosity (10^{21} Pas) that would cause stalled subduction in homogeneous models. Across our suite, homogeneous models show a broad velocity range (0.3–16 cm yr^{-1} time-averaged, 0.5–23.6 cm yr^{-1} maximums), but the models with along-strike variations more closely mirror nature (1–9.6 cm yr^{-1} time-averaged, 1.1–14.4 cm yr^{-1} maximums). These variable models simultaneously yield a wider range of interface stresses, with time-averaged values that span 2–44 MPa (58 MPa maximum) compared to 7–19 MPa (23 MPa maximum) in homogeneous models (Fig 9b). While only 17 per cent (4 segments) of segments exceed the empirical 35 MPa stress threshold at some point during the model run, three of these maintain convergence rates above 2 cm yr^{-1}, indicating that higher interface stresses are viable under certain geometries. Notably, 43 per cent of segments exceed 30 MPa, even with moderate viscosities (10^{20} Pas). These findings show that along-strike variability not only narrows velocity distributions to a more realistic range but also allows a broader range of interface stresses, challenging the assumption that velocity alone can be used to constrain interface strength.

4.2.1 Deformation and topography

High-interface shear stresses play a key role in transmitting stress from the negatively buoyant slab and into the OP (e.g. S. Lamb 2006), promoting subduction margin deformation. Fig. 10 illustrates a clear relationship between OP compressional force, extracted before slab anchoring and interface stress, showing an exponential increase in OP force with rising interface stress. As discussed in Section 4.2, homogeneous models yield a relatively narrow range of interface stress, whereas introducing along-strike strength variations significantly expands the range of interface stresses leading to a larger range of compressional forces (0.1 to 15.2 TN m^{-1}; Fig. 10).

In addition to interface shear stresses, the exponential relationship observed in Fig. 10 may also result from variations in the

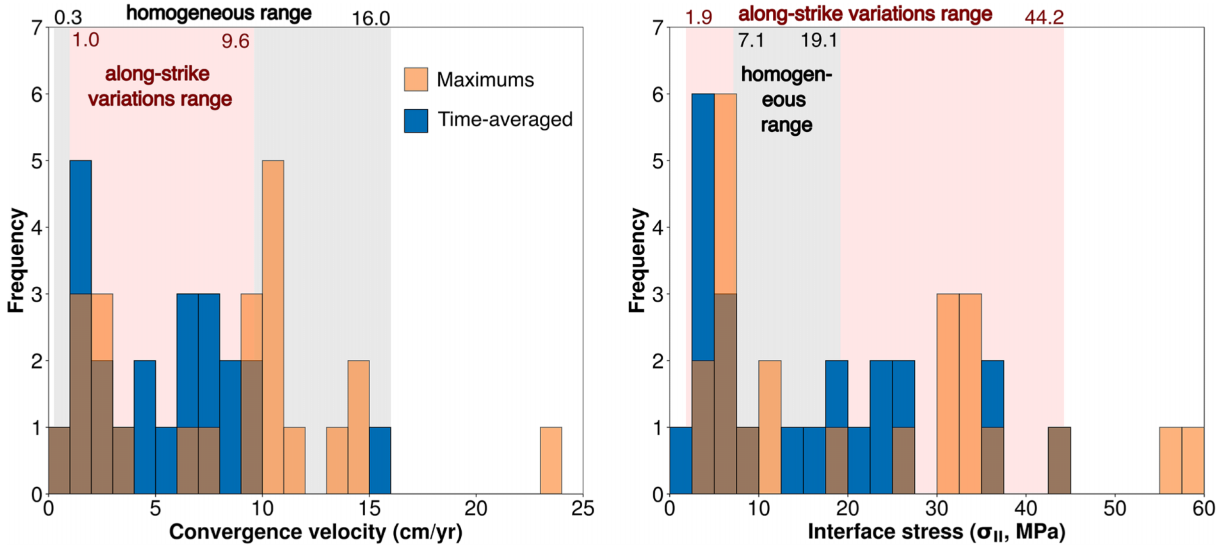

Figure 9. Histograms showing values for the (a) convergence velocity and (b) interface stresses. Shaded red regions show the range for the time-averaged models with along-strike variations, and shaded grey shows the range for homogeneous models.

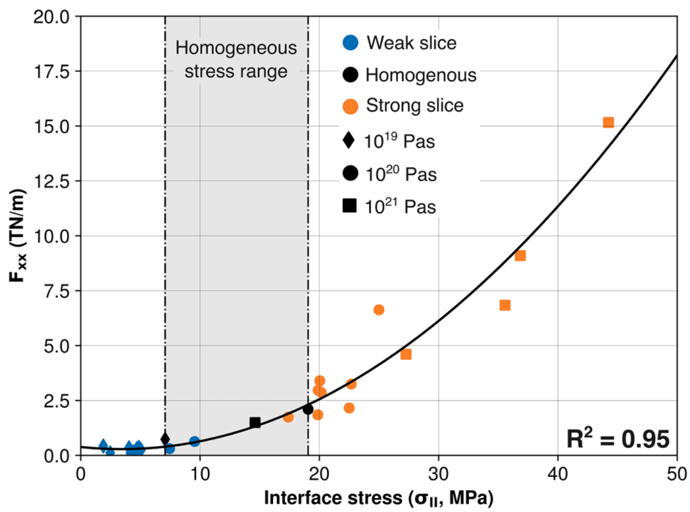

Figure 10. Time-averaged compressive overriding plate force versus interface stress, from initiation (200 km) to 700 km of subduction.

normal stresses, as potentially influenced by geometric factors such as slab dip. Slab dip affects the connection between interface and OP stress states: a steeper dip can increase interface normal stresses while reducing shear stresses. However, we note that in our models, high-shear stresses persist even at the largest slab dips (e.g. in model C1921, shear and normal interface stresses reach up to 95 and 28 MPa in the strong central segment, respectively, at the time of maximum OP compression). Force balance analysis further suggests that larger slab dips at strong interfaces enhance both the normal component of interface stresses and the horizontal normal stresses transmitted to the overriding plate (S. Lamb 2006). When OP forces are normalized by slab dip, the relationship becomes approximately linear (Text S3 and Fig. S8, Supplementary Material), implying that slab dip variations contribute to the elevated OP compression observed at high-interface strengths during the free-sinking phase. Strong interfaces also reduce trench retreat (W.M. Behr *et al.* 2022), and in some cases, promote a transition to trench advance, further enhancing compressional deformation in the overriding plate (F.A. Capitanio *et al.* 2010; A.F. Holt *et al.* 2015; M. Pons *et al.* 2022; W.P. Schellart 2024). We note that basal drag is another important component to consider for subduction dynamics (e.g. L. Suchoy *et al.* 2021; N.G. Cerpa *et al.* 2022) and the OP force balance (e.g. W.P. Schellart & L. Moresi 2013), with trench-perpendicular gradients in basal traction shown to correlate with OP stress regime (e.g. F.A. Capitanio *et al.* 2010; W.P. Schellart & L. Moresi 2013), particularly in the far-field. However, in this study we primarily focus on OP deformation in the near-forearc region and on the impact of interface-stress transmission, which we expect to dominate adjacent to the plate interface. Analyses of subduction transects suggests back-arc compression occurs primarily above relatively low-slab dips (S. Lallemand *et al.* 2005), and that the stress regime of the overriding plate following slab anchoring depends on temporal variations in slab dip (N.G. Cerpa *et al.* 2018). More research is required to better understand the interplay between slab dip and the normal and shear stresses along the interface, how this varies along-strike within 3-D subduction zones containing variable interface lithologies, and how this relates overriding plate stress state and deformation

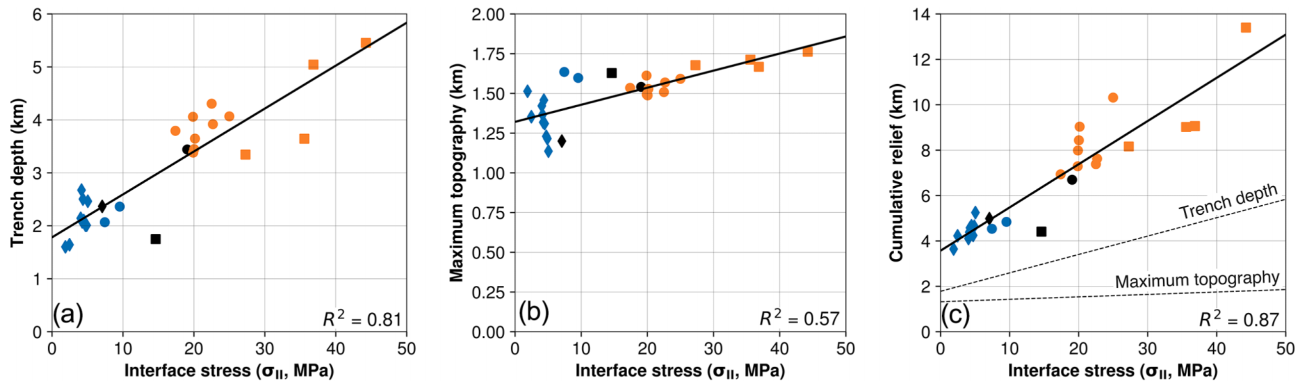

Figure 11. Time-averaged interface stress versus time-averaged (a) trench depth, (b) maximum topography and (c) cumulative relief. Cumulative relief is the integration of topographic changes from the trench to 500 km towards the direction of subduction. For information on symbols and colours, see Fig. 10.

Our results indicate that overriding plate deformation is significantly influenced by along-strike variations in slab pull forces, which arise from interface strength heterogeneities. Homogeneous models, by neglecting these variations, tend to underestimate both interface shear stresses and overriding plate forces. In contrast, incorporating along-strike heterogeneity reveals forces up to ∼7 times higher, suggesting a more prominent role for these variations in controlling overriding plate dynamics than previously recognized.

4.2.2 Topography

Subduction zone and overriding plate topography varies based on a variety of factors such as the slab dip, buoyancy and plate strengths (F. Crameri *et al.* 2017), as well as overriding plate heterogeneity (A.G. Grima & T.W. Becker 2024). Here, we focus on the interface stress magnitude which is also thought to impact topography, affecting trench depth and maximum elevation of the overriding plate (S. Lamb 2006; A. Dielforder *et al.* 2020). In our models, we see that trench depths correlate positively with interface stresses, ranging from 1.6 to 5.5 km (Fig. 11a). Maximum elevations similarly increase with higher stresses (Fig. 11b) but less dramatically than anticipated by analytical studies (e.g. up to ∼4 km for a ∼50 MPa 80-km thick constant stress interface A. Dielforder *et al.* 2020). This lower-than-expected topography may result from limited model resolution (10 km in the OP) and an absence of plastic strain weakening, both of which restrict the development of faults that would drive uplift. Nonetheless, strong interfaces clearly produce greater cumulative relief variations (3.6–13.4 km, Fig. 11c), correlating with deeper forearc basins and highs. Although compressional forces typically promote surface uplift, the slab's downwards pull in strongly coupled, strong interface regions can counteract this effect and suppress topography (e.g. W.M. Behr *et al.* 2022). Our strong interface models show increased relief and deeper forearc basins, consistent with features like the Atacama Bench in the Andes (R. Armijo *et al.* 2015). Our results also suggest that after slab anchoring, when the slab becomes supported by the lower mantle, uplift occurs in the forearc basin. These findings highlight the need for further study on how along-strike variations in interface strength influence sediment transport and deposition from the forearc to the trench.

4.3 Trench location and natural comparisons

A key observation from our models is that along-strike variations in interface strength significantly influence trench location and slab shape. Such variability in subduction geometry is commonly observed in nature, for example, within the Andes and Lesser Antilles subduction zones. In our simulations, strong interfaces limit slab roll back and increase OP compression, resulting in steeper slab dips and advanced trench positions. When a strong interface region is centrally located, it produces concave margin geometries (e.g. model C1921 or C1920, Fig. 7 and Fig. S9, Supplementary Material). A natural analogue to this configuration is potentially the Bolivian Orocline in the Andes, characterized by an advanced trench with substantial overriding plate deformation. Previous research has attributed this geometry partially to elevated shear stresses resulting from reduced sediment supply (S. Lamb & P. Davis 2003; J. Hu *et al.* 2021), although trench length, plate thickness and slab dynamics may also significantly contribute to such configurations. Shorter subduction zones typically exhibit convex trench geometries towards the subducting plate (e.g. Fig. 3), while longer subduction margins, exceeding several thousand kilometers, often develop concave shapes (W.P. Schellart 2024). Additionally, along-strike variations in the thickness of the subducting and overriding plates influence trench migration and curvature: thicker subducting plates promote rollback and enhance OP compressional stresses, yielding concave shapes when centrally positioned (F. Capitanio *et al.* 2011). Conversely, thicker overriding plates tend to reduce trench migration, causing convex trench geometries (F. Capitanio *et al.* 2011). The Andean margin, extending approximately 7400 km, exhibits notable along-strike variations in subducting plate age (S. Lamb & P. Davis 2003; F. Capitanio *et al.* 2011), and sediment availability possibly linked to climatic variations (S. Lamb & P. Davis 2003; J. Hu *et al.* 2021). The distinctive concave shape and pronounced deformation along the Bolivian Orocline may arise from a combination of its extensive length, thicker subducting plate (F. Capitanio *et al.* 2011) and amplified local stresses due to a stronger interface set up by reduced sedimentation (S. Lamb & P. Davis 2003; J. Hu *et al.* 2021).

When strong interfaces are not axially symmetric but instead paired with weaker interfaces along-strike, an S-shaped trench can form as the weak-interface slab retreats (e.g. M2019, M2021, or M1921, Fig. 4 and Fig. S9, Supplementary Material). A similar trench morphology is observed in the Lesser Antilles subduction zone in the Caribbean, which during the Eocene experienced shortening, partially due to trench curvature (N.G. Cerpa *et al.* 2021), in the sediment-starved north and extension in the sediment-rich south (N.G. Cerpa *et al.* 2021). This northern compressional phase was followed by forearc extension in the Oligocene (M. Philippon *et al.* 2020; N.G. Cerpa *et al.* 2021). In the southern segment, sediment influx from the South American continent that began in the Eocene (N.L. Bangs *et al.* 2003) may have lubricated the subduction interface (e.g. W.M. Behr & T.W. Becker 2018), potentially facilitating faster convergence velocities and promoting slab retreat. Following initial lubrication, sustained sediment supply could have supported growth of the accretionary wedge, increasing the interface length. If the interface viscosity remained constant, this increased interface length coupled to reduced slab pull due to sediment buoyancy might have subsequently slowed convergence velocities and slab retreat (S. Brizzi *et al.* 2021; J.Y. Keum & B.D. So 2021; J. Munch *et al.* 2022; J.Y. Keum & B.D. So 2023). Thus, the overall observed velocities and compressive forces along the margin could reflect a dynamic balance among the rates of sediment input to the interface, accretionary wedge accumulation, and northward sediment propagation.

4.4 Model limitations

The models presented here are simplified to remain interpretable and computationally feasible, while still capturing first-order subduction dynamics. However, because of this we neglect multiple factors that can have an impact on evolving subduction systems. For example, in using a constant interface viscosity, we neglect temperature-dependent, power-law rheology. Importantly, this means the entire interface exhibits an average, viscous shear stress, whereas in nature one would expect an increase in shear stress with depth until the brittle–ductile transition, producing a shallow region of very high shear stresses to drive fault localization and overriding plate deformation. Similarly, we neglect surface processes, which could introduce weak sediment rheology to the trench and alter interface strength through time (e.g. W.M. Behr & T.W. Becker 2018; S.V. Sobolev & M. Brown 2019). Additionally, while in our models a weak interface broadly represents a sedimented interface, to reduce model complexity and focus solely on strength we omit the associated interface density variations. Such density changes have been suggested to influence dynamics, trench location and convergence velocities (S. Brizzi *et al.* 2021; J.Y. Keum & B.D. So 2021; J. Munch *et al.* 2022; J.Y. Keum & B.D. So 2023). In large-scale 3-D models we are constrained by the model resolution, reaching a maximum of 5 km in the interface or 10 km in the overriding plate. These resolutions can effect minimum shear zone widths and stress transfer rate across the interface (C.P. Conrad & B.H. Hager 1999), as such we may underestimate compressional forces within the overriding plate. Our models capture only a single free-sinking phase. However, in nature, multiple episodes of slab sinking may occur due to processes like slab buckling, which is associated with elevated convergence rates and compressional forces (e.g. N.M. Ribe 2003; C. Lee & S.D. King 2011; G. Gibert *et al.* 2012; N.G. Cerpa *et al.* 2014; M. Pons *et al.* 2022; E. van der Wiel *et al.* 2024). This means that compressional events linked to interface strength variations could occur repeatedly over time, a behaviour not captured in our current modelling framework.

5 CONCLUSIONS

In this study, we explored how along-strike changes in plate interface strength affects a time-evolving 3-D subduction zone. Our results show that many aspects of subduction zone behaviour in our models, such as slab dip, trench location and overriding plate deformation, are broadly consistent with findings from previous 2-D studies. For example, weaker interface segments promote trench retreat, lower slab dips and extension in the overriding plate. However, our 3-D models reveal an important dynamic not captured in 2-D or homogeneous setups: slab segments interact laterally, with weak segments driving faster convergence that enhances velocities even in adjacent, stronger segments. This coupling leads to elevated interface shear stresses and increased transfer of compressional stress to the overriding plate within the strong interface regions. Critically, the magnitude of the along-strike variation in interface strength governs the velocity change within a strong segment, while the local interface strength determines the magnitude of stress amplification.

On a broader scale, along-strike variations in interface strength influence subduction dynamics beyond their immediate location. These variations narrow the range of convergence velocities compared to homogeneous models with similar strength end-members, bringing model behaviour closer to natural observations. At the same time, they expand the range of viable interface stresses. Natural examples such as the Andean and Lesser Antilles subduction zones may reflect this behaviour: both exhibit elevated overriding plate deformation in sediment-starved regions, with trench geometries characterized by locally advanced trenches adjacent to more retreated, sediment-rich segments. In the Lesser Antilles, increased compressional stresses since the Eocene also coincide with accretionary wedge formation. These observations are consistent with our results, which suggest that along-strike strength variations can amplify interface shear stress and overriding plate deformation. This mechanism may be underestimated in earlier models and operates across spatial scales relevant to evolving subduction zones.

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SUPPORTING INFORMATION

Supplementary data are available at *GJIRAS* online.

suppl_data

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DATA AVAILABILITY

Files and the specific ASPECT version used to run the models here can be found at https://doi.org/10.5281/zenodo.16779892 (Neuharth *et al.* 2025).

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