

# Decoupling of Plate-Asthenosphere Motion Caused by Non-linear Viscosity During Slab Folding in the Transition Zone

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## Abstract

Although most present-day subduction zones are in trench retreat, plate reconstructions and geological observations show that individual margins experience episodes of advancing, retreating or stationary trench motion with time-variable subduction rates. However, most laboratory and numerical simulations predict steady plate velocities and sustained trench retreat unless the slab experiences folding in the transition zone. Using 2D dynamical models of subduction with a mobile trench and overriding plate, we find that rapid sinking of the slab during folding causes a reduction in asthenosphere viscosity through the non-linear rheology, which allows the overriding plate to move in the opposite direction of the asthenosphere. This decoupling of the direction of plate and asthenosphere flow allows for episodes of rapid trench advance after each slab folding event. By analyzing the interaction between slab deformation (sinking direction and speed), stress-induced changes in asthenosphere viscosity, asthenosphere flow and plate motions, we show that there are three modes of slab-flow-plate interaction: 1) coupled trench retreat during rapid vertical sinking, 2) coupled trench advance during prograde sinking of the slab, and 3) decoupled, rapid trench advance during folding with prograde motion of the shallow slab and retrograde motion of the deep slab. These results show that non-linear viscosity plays an important role in determining the force balance controlling trench motion and conversely that trench motion can be used as a constraint on the asthenosphere viscosity underlying the overriding plate. In addition, cooling by several hundreds of degrees during episodes of fast subduction could lead to a reduction in slab dehydration and fluid-induced melting in the mantle wedge. Such cold episodes would also likely lead to time-variability in the water content and related geochemical tracers in erupted lavas, as well as the amount of water being transported by slabs into the deep mantle.

*Keywords:* slab deformation, phase transitions, plate-mantle coupling, rheology, slab thermal structure, mantle wedge thermal structure

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## 1. Introduction

2 The motion of tectonic plates at the Earth's surface is the most direct observation of large-scale mantle flow  
3 reflecting the time-dependent balance of driving and resisting forces acting on the base of the plates and through  
4 slab-pull and ridge push (Forsyth and Uyeda, 1975; Lithgow-Bertelloni and Richards, 1998; Conrad and Lithgow-  
5 Bertelloni, 2002; Gérard et al., 2012). A recent analysis of present-day trench motion using different reference

6 frames concludes that 62–78% of trenches are in retreat, with median trench velocity of 0.9–1.3 cm/yr (retreating),  
7 and faster retreat only observed near slab edges (Schellart et al., 2008). In about 80% of subduction zones, subducting  
8 plate motion accounts for more than 50% of net convergence (Schellart et al., 2011). Plate tectonic reconstructions of  
9 past plate motion also show that individual trenches experience both advancing and retreating motion, although this  
10 is hard to constrain further back in time (Sdrolias and Müller, 2006).

11 Subduction zone force-balance analysis attempts to account for how each of the forces acting on the subducting  
12 plate, overriding plate, slab, and across the plate boundary interface, result in the observed plate motions (Forsyth  
13 and Uyeda, 1975). There have been various iterations of force-balance models, each making different assumptions  
14 or simplifications in an attempt to elucidate the first-order balance of forces controlling trench motion (Schellart,  
15 2004; Heuret and Lallemand, 2005; Stegman et al., 2006; Lallemand et al., 2008; Billen, 2008; Capitanio, 2013; Holt  
16 and Becker, 2017). In a more sophisticated approach, semi-analytic flow solutions accounting for the forces due to  
17 poloidal (corner-flow) and torroidal flow induced by the sinking slab are combined to predict trench motions (Royden  
18 and Husson, 2006; Capitanio et al., 2007; van Dinther et al., 2010). In all of these analysis, when the overriding plate  
19 is included, it is assumed that the plate and underlying asthenosphere move in the same direction. If the asthenosphere  
20 moves faster than the plate, then it exerts a driving force on the plate helping to drag it forward, whereas if the  
21 asthenosphere moves slower it exerts a resisting force.

22 While most force-balance analysis assume a steady-state slab geometry, the observed shapes of subducting slabs  
23 (e.g., van der Hilst et al., 1997; Ritsema et al., 2004) are due to the time-dependent variation in both driving and re-  
24 sisting forces and resulting changes in plate and trench motions. For example, trench retreat combined with increased  
25 resistance to sinking into the lower mantle is thought to form sub-horizontal slabs just above, on, or just below the  
26 upper-lower mantle boundary at 660 km (Fukao et al., 2009). Whereas, broadening of slabs in the lower mantle is  
27 thought to be caused by folding or buckling of the slab as it encounters increased resistance to sinking into the lower  
28 mantle (Ribe et al., 2007). Such buckling has been shown to depend on the strength contrast between the slab and  
29 the surrounding mantle and the viscosity contrast (and/or resistance due to phase transitions) across the upper-lower  
30 mantle boundary (Ribe, 2010; Stegman et al., 2010; Lee and King, 2011).

31 Geodynamic models of subduction show that strong trench retreat is promoted by stiff and less dense slabs (Fu-  
32 niciello et al., 2008; Garel et al., 2014; Agrusta et al., 2017), shorter along-strike trench length (Stegman et al., 2006;  
33 Schellart et al., 2007; Stegman et al., 2010), proximity to a slab edge (Schellart et al., 2011) and thinner and/or more  
34 buoyant overriding plates (Holt et al., 2015). A synthesis of results from analogue models found that both weak  
35 ( $\eta_{slab}/\eta_{mantle} < 10^2 - 10^3$ ) or stiff ( $\eta_{slab}/\eta_{mantle} > 10^4$ ) and less-dense slabs exhibit trench retreat, with intermediate  
36 slab stiffness generating either slab folding or trench advance (Schellart, 2008b). Similarly, a comprehensive study us-  
37 ing 3D numerical models (but not including the lower mantle or an overriding plate) and linear visco-plastic rheology  
38 showed that dense and weak or less dense and stiff slabs exhibited trench retreat with or without folding, respectively,  
39 while dense and stiffer slabs exhibited slab folding and trench advance (Stegman et al., 2010). However, it has also  
40 been shown that non-linear viscosity, which increases  $\eta_{slab}/\eta_{mantle}$  by weakening the mantle, promotes trench advance

41 (Holt and Becker, 2017) consistent with the results of simulations exhibiting slower rates of trench retreat (Quinteros  
42 et al., 2010; Garel et al., 2014).

43 In addition to subduction geometry and rheology, phase transitions also affect slab dynamics and surface plate  
44 motion (Christensen, 1996; Cížková et al., 2007; Cížková and Bina, 2013; Agrusta et al., 2017; Yoshida, 2017). In  
45 simulations similar to Quinteros et al. (2010) and (Garel et al., 2014), but including a strong negative Clapeyron slope  
46 for the phase transition at 660 km, strong trench retreat occurs, but with time-variable subduction rates due to localized  
47 slab buckling in the transition zone (Cížková and Bina, 2013). In contrast to these models, a study by Arredondo and  
48 Billen (2017) found that when using a more complete compositionally-dependent phase transition (CDPT) model,  
49 non-linear rheology and a deep model domain (bottom at the core-mantle boundary) trench motion was primarily in  
50 advance, even with episodic slab folding in the transition zone. However, the study by Arredondo and Billen (2017)  
51 was focused on the effect of the new CDPT model and the role of the shear zone viscosity and plastic yielding on slab  
52 behavior and did not consider how other aspects of the model affected slab deformation and trench motion. These  
53 divergent model behaviors found in similar studies illustrates the complex feedback that occurs between different  
54 choices of model parameters (e.g., phase transitions, type of rheology) and model design (e.g., 2D vs. 3D, box depth),  
55 making it challenging to constrain the balance of forces controlling plate and trench motion.

56 The thermal structure of the shallow slab and mantle wedge can also provide a further constraint on geodynamic  
57 models of subduction because the mantle wedge must remain hot enough to generate magmas that feed the volcanic arc  
58 (Elkins-Tanton et al., 2001; Kelemen et al., 2003; Cagnioncle et al., 2007; England and Katz, 2010). This constraint is  
59 violated in most global models of subduction simply because the resolution necessary to accurately model the mantle  
60 wedge is not usually computationally feasible although this is changing (see Stadler et al., 2010; Alisic et al., 2010).  
61 However, most subduction modeling studies also do not explicitly consider this constraint (however, see Arcay, 2017)  
62 and sometimes end up with a broad, deep, cold mantle wedge corner (Cížková and Bina, 2013; Garel et al., 2014;  
63 Agrusta et al., 2017). The broad, thick and higher viscosity region that develops above the slab in these models is also  
64 likely to affect how the slab deforms by preventing the shallow slab dip from changing and creating a large positive  
65 pressure above the slab (pushing it backwards).

66 Here we build on the growing understanding of the link between slab deformation and plate motions from previous  
67 studies in an effort to disentangle the various model design and parameter choices that determine slab, flow and plate  
68 interaction. Here, we first build on the Arredondo and Billen (2017) study by varying the overriding plate structure  
69 (age, compositional density, and a spreading ridge) and show that while these changes do produce more trench retreat,  
70 the model parameters still do not produce steady trench retreat as found in many other models. Instead, we present  
71 new analysis showing that non-linear viscosity controls the time-dependent balance of forces on the overriding plate  
72 and trench through stress-induced reduction of the asthenosphere viscosity (as low as  $10^{18}$  Pa s). The low viscosity  
73 allows for decoupling the direction of flow in the asthenosphere from that in the overriding plate. More specifically,  
74 by analyzing the interaction between slab deformation (sinking direction and speed), stress-induced changes in as-  
75 thenosphere viscosity, and asthenosphere flow and plate motions, we show that there are both periods of coupled

76 motion of the overriding plate and underlying asthenosphere in retreat (mode 1) or advance (mode 2), and there are  
77 periods of decoupled motion during which the overriding plate advances while the underlying asthenosphere is pulled  
78 toward the slab (mode 3). In addition, we show how the resulting time-variable subduction leads to changes in the  
79 thermal structure of the slab and mantle wedge, the resulting effect on slab dynamics, and how this compares to the  
80 steady-state thermal structure resulting from kinematic-slab wedge-flow thermal models (Peacock and Wang, 1999;  
81 van Keken et al., 2002; Currie et al., 2004; Wada et al., 2008; Syracuse et al., 2010).

## 82 **2. Methods**

83 We model the time-evolution of slab deformation and thermal structure, in a two-dimensional equatorial slice  
84 of a sphere, using the finite element code CitcomS (Zhong et al., 2000; McNamara and Zhong, 2004; Tan et al.,  
85 2006). CitcomS solves the conservation equations for mass, momentum and energy using the extended Boussinesq  
86 approximation, which assumes incompressibility but includes an initial adiabatic gradient, shear heating and latent  
87 heat from phase transitions (Christensen and Yuen, 1985; Ita and King, 1994). The full model domain is  $61^\circ$  in  
88 longitude and 2890 km in depth with a minimum element size of 1.5 km x 2.5 km in a 1000 km wide by 170 km  
89 deep region centered at  $35^\circ$  (Fig. 1a). The element size then gradually increases with depth and longitudinal distance  
90 reaching a maximum size of 8.3 km x 10 km below 1100 km. There are free-slip boundary conditions on all the  
91 domain boundaries (Fig. 1a).

92 *Initial Temperature and Composition.* The top and bottom surface are isothermal ( $0^\circ\text{C}$ ;  $2075^\circ\text{C}$ ) and there is an initial  
93 adiabatic gradient of 0.25 K/km starting at 190 km depth. The sidewalls are insulating. The subducting and overriding  
94 plates are defined thermally using a half-space cooling model and the age of the plate: for the subducting plate the  
95 initial age increases from a ridge at  $0^\circ$  longitude to either 40 my or 80 my using a constant spreading rate. The age  
96 of the overriding plate is constant (either 20 or 40 my), except in some models in which a spreading ridge is added at  
97 the model boundary at  $61^\circ$  (see Table 1). The proto-slab is created by first running the model with kinematic surface  
98 boundary conditions (5 cm/yr on the subducting plate) until the tip of the slab reaches 200 km depth (Fig. 1a, b).  
99 The density anomaly of the slab depends on the minimum temperature of the slab, which varies with a maximum  
100 difference in the range of 33 – 46  $\text{kg/m}^3$  for a temperature anomaly of  $500\text{--}700^\circ\text{C}$ .

101 In addition to the temperature, the plates are also defined by compositional layers that are tracked using tracer  
102 particles (Fig. 1b): a 7.5 km basaltic crust ( $3000 \text{ kg/m}^3$ ) overlying a 27.5 km harzburgite ( $3235 \text{ kg/m}^3$ ) layer, with  
103 the remaining mantle having the composition of pyrolite ( $3300 \text{ kg/m}^3$ ). In some models these layers are omitted on  
104 the overriding plate to test the effect of compositional buoyancy on plate motion. Basalt is modeled as transitioning  
105 to eclogite ( $3540 \text{ kg/m}^3$ ) starting at  $700^\circ\text{C}$  for pressures above 15 kbar and ending at  $850^\circ\text{C}$  (Arrial and Billen, 2013;  
106 Arredondo and Billen, 2016).

107 *Rheology.* The rheology is the same as that used in previous studies (Billen and Hirth, 2007; Arredondo and Billen,  
108 2017). The upper mantle (pyrolite, harzburgite, eclogite) are modeled with a composite viscosity using the flow-laws

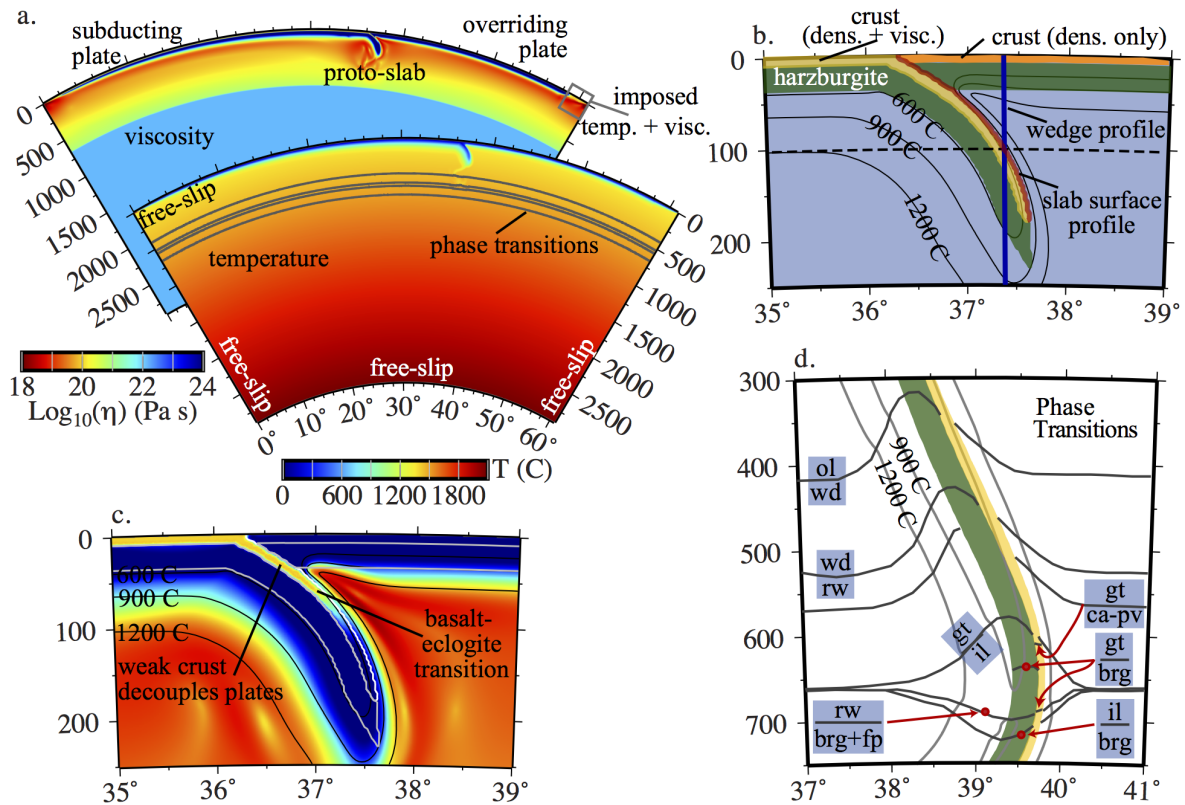


Figure 1: Model Design. a) Full Model domain showing initial viscosity and temperature structure, phase transitions and boundary conditions. Note 500 km wide box on trailing edge of overriding plate: the temperature and viscosity are reset in the box to allow for a mobile overriding plate and preventing formation of a subduction zone or maintaining the imposed mid-ocean ridge. b) Compositional layers and profile locations used to analyze the evolution of the temperature structure. Yellow layer: oceanic crust (basalt or eclogite density, fixed viscosity). Green layer: harzburgite with an olivine flow-law. Orange layer: oceanic crust (basalt density only). Light blue background: pyrolite composition with olivine flow law. Red contour: the slab surface profile follows the top of the crustal layer. Blue profile: the vertical wedge profile is located at the longitude where the slab-surface crosses a depth of 100 km. Note that because the slab and trench are free to move, the absolute location of the profiles change with time. c) Zoom-in on subduction plate boundary showing the viscosity structure for the proto-slab. The subducting and overriding plate are decoupled by a weak crustal layer with basalt-eclogite composition. Black contours are temperature. White contours outline the crust and harzburgite layers (shown in yellow and green in parts C and D). d) Zoom-in on the transition zone showing the compositionally-dependent phase transition boundaries across a sinking slab (snap-shot is from model 4 at 46.0 my). Yellow/green layers are crust/harzburgite composition.

109 for diffusion and dislocation creep in olivine (Hirth and Kohlstedt, 2003) and a depth dependent yield stress reaching  
 110 a maximum yield stress of 1000 MPa at 80 km. The background, upper mantle viscosity is  $10^{20}$  Pa s at 250 km  
 111 depth for a strain-rate of  $10^{-15}$   $\text{s}^{-1}$  and increases with depth. The maximum viscosity for the basalt layer is  $10^{20}$  Pa s,  
 112 which allows the subducting plate to slide past the overriding plate (Fig. 1c). The viscosity of this layer smoothly  
 113 transitions to the olivine viscosity as the basalt composition transitions to eclogite. The lower mantle is modeled using  
 114 the diffusion-creep flow law for olivine, with a large intrinsic grain size, to assign a background viscosity of  $10^{22}$  Pa s.  
 115 The full range of viscosity allowed in the models is  $10^{18}$ – $10^{24}$  Pa s and occurs between the mantle surrounding the

116 slab, mantle wedge, or asthenosphere and the cold interior of the slab or lithosphere, respectively.

117 For models without a spreading ridge on the overriding plate, we impose a 500-km wide region of low viscosity  
118 at the domain boundary to allow the overriding plate to freely move toward or away from the model edge. The  
119 temperature within this region is also held fixed to prevent the formation of a downwelling during times of trench  
120 advance or a ridge during trench retreat (unless a ridge is explicitly included in the model). Note that because the low  
121 viscosity crustal layer is being tracked by tracer particles, the subduction location is free to move in response to the  
122 slab dynamics and tractions acting on the bottom of the plates.

123 *Phase Transitions.* A compositionally-dependent phase transition (CDPT) model tracks the proportions of the olivine,  
124 pyroxene and garnet minerals in each composition and the phase transitions for these minerals (Fig. 1d): for specific  
125 model parameters and complete references see Arredondo and Billen (2016) and Arredondo and Billen (2017). For the  
126 olivine component, the phase transitions include olivine → wadsleyite (410 km), wadsleyite → ringwoodite (520 km),  
127 and ringwoodite → bridgmanite+ferropericlase (660 km). For the pyroxene component, there is a non-temperature  
128 dependent dissolution to garnet starting at about 300 km depth that we do not include because it does not affect the  
129 density anomaly between the slab and the surrounding mantle. Calcium-perovskite starts to form at around 560 km  
130 depth from exsolution of calcium-rich garnet and dissolution of clinopyroxene. This transition occurs deeper in the  
131 eclogite layer (665 km) and does not occur in harzburgite. Majoritic garnet either transitions directly to bridgmanite, or  
132 in cold regions, first transitions to ilmenite above 660 km and then ilmenite transitions to bridgmanite below 660 km.

133 Compared to models assuming a 100% olivine composition, the single phase transition at 660 km occurs as a  
134 series of transitions with both positive and negative clapeyron slopes, which act to decrease the overall resistance to  
135 subduction at the base of the transition zone. And, unlike previous models considering only the phase transitions at  
136 410 km and 660 km, the additional phase transitions in the mid-transition zone add to the overall negative buoyancy  
137 of the slab (Arredondo and Billen, 2017).

### 138 **3. Results**

139 We first present the dynamical behavior of the models describing the relationship between slab dynamics, trench  
140 and plate motion and the model parameters (see statistics in Table 1). Second, we present the effect of time-variable  
141 slab dynamics on slab and wedge thermal structure. Movies (S1–S7) of the simulations are available in the Supple-  
142 mental Information.

#### 143 *3.1. Slab Folding, Trench and Plate Motions*

144 As has been shown in the previous work discussed in the introduction, we find that the resistance to sinking at 660  
145 km from a combination of the increase in viscosity and the phase transitions leads to episodic folding of the slab in the  
146 transition zone (models 1–6; Fig. 2b–d and movies in supplemental information). In contrast the slab in an identical

Table 1: Summary of Variable Model Parameters &amp; Plate Motion Results

Model	OP/SP Age (My)	Slab Dyn. -	Model Duration (my)	Trench Motion (a, r, s) %	Trench. Vel. +max, -max (cm./yr)	OP Vel. med, +max, -max (cm/yr)	SP Vel. med, maxlt (cm/yr)
No Overriding Plate Compositional Buoyancy							
1	20/80	f	49.5	76.1, 19.4, 4.5	3.4, -1.3	0.7, 3.1, -1.5	4.6, 11.1
1c	20/80	b	19.0	$\eta_{cr} = 10^{21}$ Pa s, slab breaks off			
2	40/80	f	40.6	47.9, 42.3, 9.8	4.2, -2.3	1.4, 4.6, -0.8	3.7, 15.5
With Overriding Plate Compositional Buoyancy							
3	20/80	f	61.6	41.3, 49.6, 9.0	1.6, -0.5	0.1, 1.8, -0.4	2.4, 6.2
3c	20/80	b	18.9	$\eta_{cr} = 10^{21}$ Pa s, slab breaks off			
With OP Buoyancy and OP Spreading Ridge							
4	20/80	f	65.2	35.8, 60.2, 4.1	2.6, -1.3	0.0, 2.4, -1.6	4.2, 9.7
5	20/40	f	44.8	49.5, 40.1, 10.4	2.3, -1.4	0.3, 2.2, -1.2	5.7, 10.3
6, $\eta_{min}$	20/80	f	62.1	9.5, 86.3, 3.4	0.5, -1.1	-0.3, 0.1, -1.1	2.6, 4.7
7, no PT	20/80	v	43.3	14.0, 75.0, 11.0	0.3, -1.1	-0.1, 0.7, -0.6	2.7, 5.0
7c, no PT	20/80	b	19.1	$\eta_{cr} = 10^{21}$ Pa s, slab breaks off			

Plate/trench velocities are positive in the direction of increasing longitude. Stationary is defined as a trench velocity less than  $\pm 0.1$  cm/yr. OP/SP: overriding plate/subducting plate. Slab Dynamics: folding (f), slab breakoff (b), vertical sinking (v). a/r/s: percentage of model run time that trench motion is in *advance*, *retreat*, or *stationary*. Plate/trench velocity ranges are determined starting 3 my after the peak SP velocity (maxtz) that occurs as the slab first enters the transition zone: med – median velocity, maxlt – maximum long term, +max – maximum positive velocity, -max – maximum negative velocity. no PT: no phase transitions were included in these models. The slab breaks-off in models 1c, 3c, and 7c (with a higher crustal viscosity) at the time listed as model duration: no further analysis is done for these models.

147 model with no phase transitions sinks directly into the mantle without folding showing that the viscosity jump of 100x  
 148 is not sufficient to cause folding given the strength of the slab (model 7, maximum yield strength of 1 GPa; Fig. 2a).

149 Slab folding leads to episodic motion of the trench and plates in which velocity increases and then decreases  
 150 during each folding event (Fig. 3). However, unlike several previous studies we find that there is no underlying  
 151 steady trench retreat. The subducting plate velocity is primarily controlled by the minimum crustal viscosity and  
 152 the relative viscosity contrast between the slab and the surrounding mantle, the latter of which varies because the  
 153 minimum viscosity of the upper mantle changes in response to the changing negative buoyancy of the slab (including

154 the added density due to phase transitions). The crustal viscosity value of  $10^{20}$  Pa s leads to a subducting plate velocity  
 155 range of 2–15 cm/yr after the initial decent of the slab through the upper mantle in agreement with observed rates of  
 156 plate motions (Table 1). However, if shear zone viscosity is increased by a factor of 5–10 the slab breaks off because  
 157 there is too much viscous resistance at the trench compared to the growing stress within the slab as it crosses the phase  
 158 boundaries in the transition zone (models 1c, 3c, 7c).

159 The folding of the slabs can be characterized by the fold frequency (how often a fold occurs) and the amplitude  
 160 of the fold (the horizontal distance between inflection points). A younger, mechanically thinner slab has a higher  
 161 frequency (25 my) compared to a mechanically stronger slab (50 my), but both have similar fold amplitudes (450 km;  
 162 Fig. 2b vs. c; model 5 versus model 4). The folding behavior is also affected by limiting the minimum mantle viscosity  
 163 to  $5 \times 10^{19}$  Pa s in two ways (Fig. 2d). First, the average sinking rate decreases (Table 1) causing the fold frequency  
 164 also to decrease. Second, the amplitude of the folding is smaller (250 km) due to the larger resistance to horizontal  
 165 motion of the slab through the upper mantle.

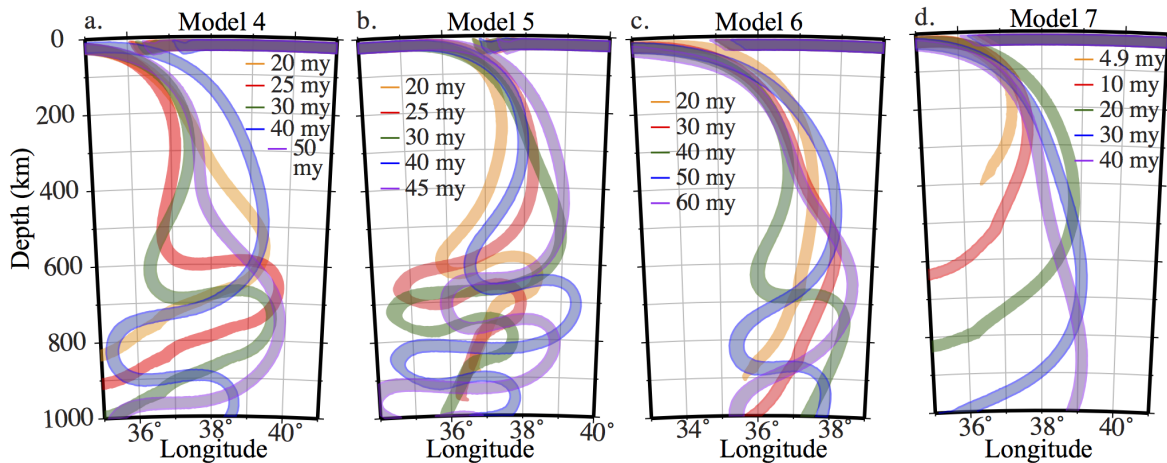


Figure 2: Comparison of slab dynamics in models with different phase change models. a) Model 4 with an older subducting plate (80 my) has less frequent folds than model 5. b) Model 5 with a young subducting plate (40 my) exhibits multiple folding events. c) Model 6 with a minimum viscosity cut-off of  $5 \times 10^{19}$  Pa s has smaller folding amplitudes and less frequent folding. d) Model 7 with no phase changes: slab sinks directly into the lower mantle and there is little time-dependent behavior.

166 To understand what might lead to the oscillatory-dominated trench motion, we explored several model parameters  
 167 that had been previously suggested to control trench motion. First, we compared the effect of overriding plate thermal  
 168 density (model 1 and 2) and found that following strong trench advance at the start of the simulation as the slab rapidly  
 169 sinks into the transition zone, the plate motions are similar until about 35 my (Fig. 3). At about 35 my an unusual arc  
 170 rifting event occurs in model 2 (see Billen, 2017) preventing further comparison of the models. During each folding  
 171 event the trench moves rapidly toward the overriding plate (advance). However, by adding compositional buoyancy  
 172 to the overriding plate there is a shift from 76% trench advance (model 1) to only 41% (model 3). The trench motion  
 173 is also slower overall and spends almost 9% of the time stationary (i.e., velocity less than  $\pm 0.1$  cm/yr; see Table 1).

174 Therefore, we conclude that overriding plate buoyancy does affect trench mobility by equalizing the isostatic pressure  
175 gradient between the overriding and subducting plates.

176 Next we add a spreading ridge to the overriding plate. Most numerical models that include temperature have  
177 included a spreading ridge on the overriding plate (Cížková and Bina, 2013; Garel et al., 2014; Agrusta et al., 2017),  
178 while mechanical models with prescribed density also include a ridge-push force because normal density material fills  
179 in behind the subducting slab (Stegman et al., 2010; Holt and Becker, 2017). However, the effect of this ridge-push  
180 force on trench motion has not been explicitly test. In addition, an analytical force-balance calculation has attributed  
181 trench advance to an unbalanced force from ridge push on the subducting plate (Capitanio, 2013). Therefore, the  
182 hypothesis that we are testing is that by only including a ridge on the subducting plate, there is a ridge-push force on  
183 the subducting plate side that is unbalanced causing the trench to always advance. By adding a spreading ridge on the  
184 overriding plate side, then the ridge push forces can be balanced. Comparison of models 3 and 4 show that adding  
185 the ridge push force only slightly decreases the amount of time spent in trench advance (from 41.3% to 35.8%), but  
186 it does decrease the amount of time the trench is stationary and therefore the trench is in retreat for 60% of the time  
187 (Table 1; Fig. 3). However, trench advance rates are still about twice as fast the trench retreat rates and therefore there  
188 is almost no net motion of the trench.

189 Other studies have concluded that trench retreat occurs for both weaker slabs or stiffer and less dense slabs (Funi-  
190 ciello et al., 2008; Schellart, 2008b; Ribe, 2010): because we use temperature-dependent rheology it is not straight-  
191 forward to make the slab less dense while also making it stiffer. In addition, while oceanic floor older than 80 my  
192 is commonly subducted, the observation of flattening of the age-subsidence curve suggest that oceanic lithosphere  
193 does not continue to thicken for ages greater than 80 my (Stein and Stein, 1992). However, we can test the effect of  
194 decreasing the stiffness of the slab by using a younger subducting plate. Here we test this by comparing the trench  
195 motion for an old subducting plate (80 my; model 4) to that for a young subducting plate (40 my; model 5). We  
196 find that because the younger slab folds more frequently, this actually leads to more time in trench advance or with a  
197 stationary trench.

198 Finally, while most previous models had not explicitly considered the role of asthenosphere viscosity on trench  
199 motion, a recent study showed that non-linear rheology could reduce trench retreat (Holt and Becker, 2017). In  
200 addition, other models with similar set-up and rheology, but with more trench retreat use a larger minimum viscosity  
201 cut-off of  $10^{19}$  Pa s (Cížková and Bina, 2013) or  $10^{20}$  Pa s (Garel et al., 2014). To test this effect in our model set-up,  
202 we changed the minimum viscosity cut-off in the model from  $10^{18}$  Pa s to  $5 \times 10^{19}$  Pa s (model 6). Increasing the  
203 minimum viscosity decreases the plate speeds from a median of 5.7 cm/yr to only 2.6 cm/yr and from a maximum of  
204 10.3 cm/yr to only 4.7 cm/yr. It also leads to more trench retreat (70%): in particular there is a fast pulse of trench  
205 retreat occurring at the start of the model as the slab sinks rapidly into the transition zone. However, after this initial  
206 phase, the trench retreats very slowly and the overall slab location remains relatively fixed with the folding event of  
207 the slab at 35–45 my overlapping the location of the slab at 10 my. Therefore, starting at about 35 my, trench retreat  
208 is mainly caused by a slow decrease in the shallow slab dip (increase in radius of curvature; see Fig. 7B) and, from

209 50–60 my, the growth of the accretionary prism.

210 One important remaining difference between our models and previous models is that the slabs sink into the lower  
 211 mantle, while in other models the slabs do not sink into the lower mantle for various reasons (resistance due to  
 212 viscosity jump, phase transition or a box boundary at 660 km). This leads to a different structure/geometry for the  
 213 return flow in mantle. In our models the return flow is primarily broad and deep, while in these other models the return  
 214 flow is isolated in the upper mantle. This difference in return flow structure may also be an important factor favoring  
 215 trench retreat.

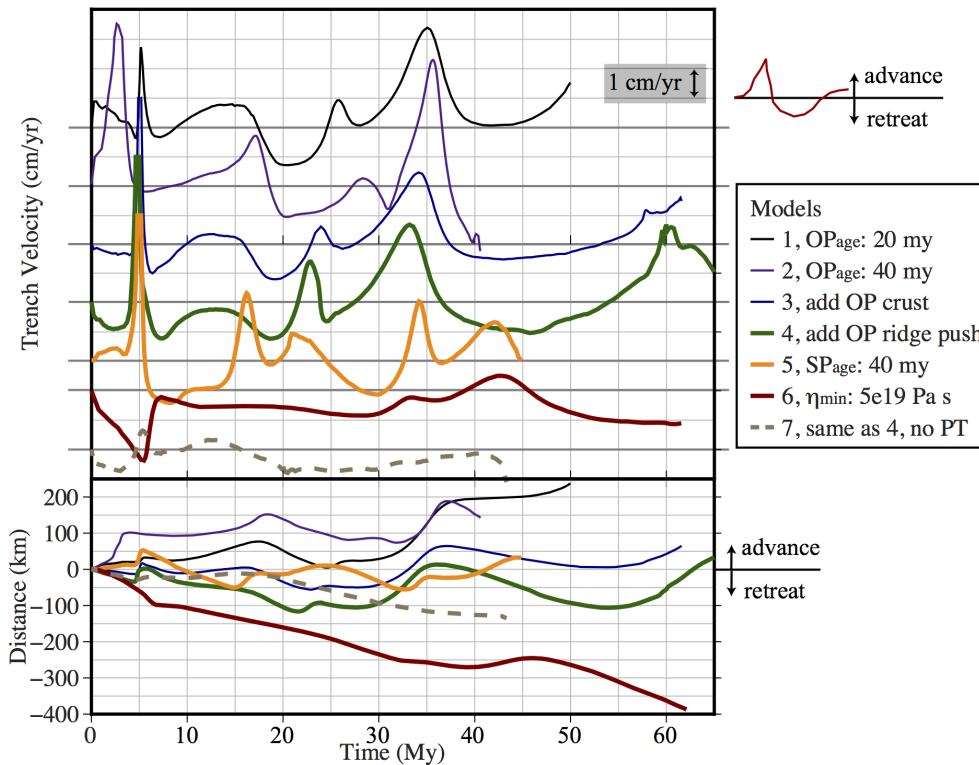


Figure 3: Trench motion. Top: trench velocity as a function of time illustrating the episodic variation of plate motion in response to folding of the slab for models 1–5. Model 7 has no slab folding. Model 6 has slab folding, but smaller corresponding changes in trench motion. Positive velocity is toward the overriding plate (advance). Negative velocity is toward the subducting plate (retreat). Bottom: distance moved by the trench with respect to the starting position. Positive distance is trench advance. Only models 6 and 7 exhibit a net trench retreat.

216 To further understand what is controlling the motion of the overriding plate and trench in the models, which differs  
 217 significantly from most previous models, we examined how deformation of the slab and the induced mantle flow is  
 218 coupled to motion of the overriding plate. In the case of steady sinking of the slab, we would expect that the sinking  
 219 motion of the slab pulls the shallow surrounding mantle in and pushes the deeper surrounding mantle down and away  
 220 (Billen, 2001; Gérard et al., 2012). It is also commonly assumed that asthenosphere flow beneath the overriding plate  
 221 is in the same direction as the plate motion (Forsyth and Uyeda, 1975; Royden and Husson, 2006; Schellart, 2008a;

222 Capitanio, 2013; Holt and Becker, 2017). In this case, if the asthenosphere is moving faster than the plate then it  
223 exerts a driving force on the plate dragging it toward the subduction zone. Whereas if the asthenosphere moves slower  
224 it exerts a resisting force slowing the overriding plate motion toward the trench. However, we find that there are times  
225 in which the overriding plate and the asthenosphere flow are in opposite directions.

226 Figure 4 shows snap-shots of the slab evolution with velocity vectors for three of the models (also see Movies  
227 S4–S6; Figures S4–S6). Each of the rows in this figure compares the models at a similar state of slab evolution. In  
228 the top row, folding of the slab has lead to rapid, near-vertical sinking of the slab, which drives rapid flow of the  
229 asthenosphere toward the slab, but the overriding plate is moving away from the trench (except in the model 6 with  
230  $\eta_{min} = 5 \times 10^{19}$  Pa s). We refer to this as *decoupled plate motion (mode 3)*, indicating that the *direction of flow* for  
231 overriding plate and underlying mantle flow are in opposite directions. In the middle row, the horizontal component of  
232 slab motion is directed toward the overriding plate following sinking of a previous fold and this drives asthenosphere  
233 and overriding plate motion away from the trench (coupled advance, mode 2). Finally, in the bottom row the slab is  
234 shown just before a folding event starts or is starting (model 6): both asthenosphere and overriding plate motion are  
235 toward the trench with the asthenosphere dragging (leading) the overriding plate (coupled retreat, mode 1). These  
236 snap-shots show that when the slab deformation includes folding, then coupling between mantle flow and the plates  
237 can be considerably more complex than expected from models with steady sinking and trench retreat.

238 The difference in behavior of models 4 and 5 compared to model 6 suggest a link between the non-linear rheology  
239 and decoupling of the asthenosphere flow from the overriding plate motion. Figure 5 compares plate motions with  
240 the asthenosphere flow and viscosity beneath the overriding plate, and the horizontal component of slab motion as  
241 function of depth for models 4–7. The supplemental information includes similar figures for all the models with  
242 additional parameters plotted (Figures S1–S7). The supplemental figures show that the subducting plate velocity and  
243 the slab velocity are always correlated and very similar in magnitude, and that the vertical component of slab velocity  
244 is remarkably constant in depth, although it varies a lot through time. This means that there is strong viscous coupling  
245 between the slab and the subducting plate and that the subducting slab must accommodate changes in stresses by  
246 horizontal motion. Therefore, a profile of subducting plate speed also shows how the magnitude of slab velocity is  
247 changing with time, while the horizontal component of slab velocity shows how the slab is responding to changing  
248 stresses. In figure 5 model 4 acts as a reference model, while model 5 shows the effect of changing the mechanical  
249 stiffness of the slab (through slab age), model 6 shows the effect of limiting the effect of the non-linear rheology  
250 (approximated by limiting the minimum viscosity), and model 7 shows the effect of not including phase transitions.

251 For models 4–7, we see that there are variations in subducting plate speed (slab sinking rate) with time, but these  
252 variations are smaller and occur less often for models 6 and 7. Also, the profiles of overriding plate velocity and  
253 trench velocity are very similar, with the small differences mainly being due to accretion of crustal material from the  
254 subducting plate. In some cases, the peaks in subducting plate (slab) velocity are matched by a switch to advancing  
255 motion of the subducting plate, but not always. The asthenosphere velocity below the overriding plate is sometimes  
256 moving with the overriding plate in retreat (mode 1) or advance (mode 2), while at other times it is directed in the

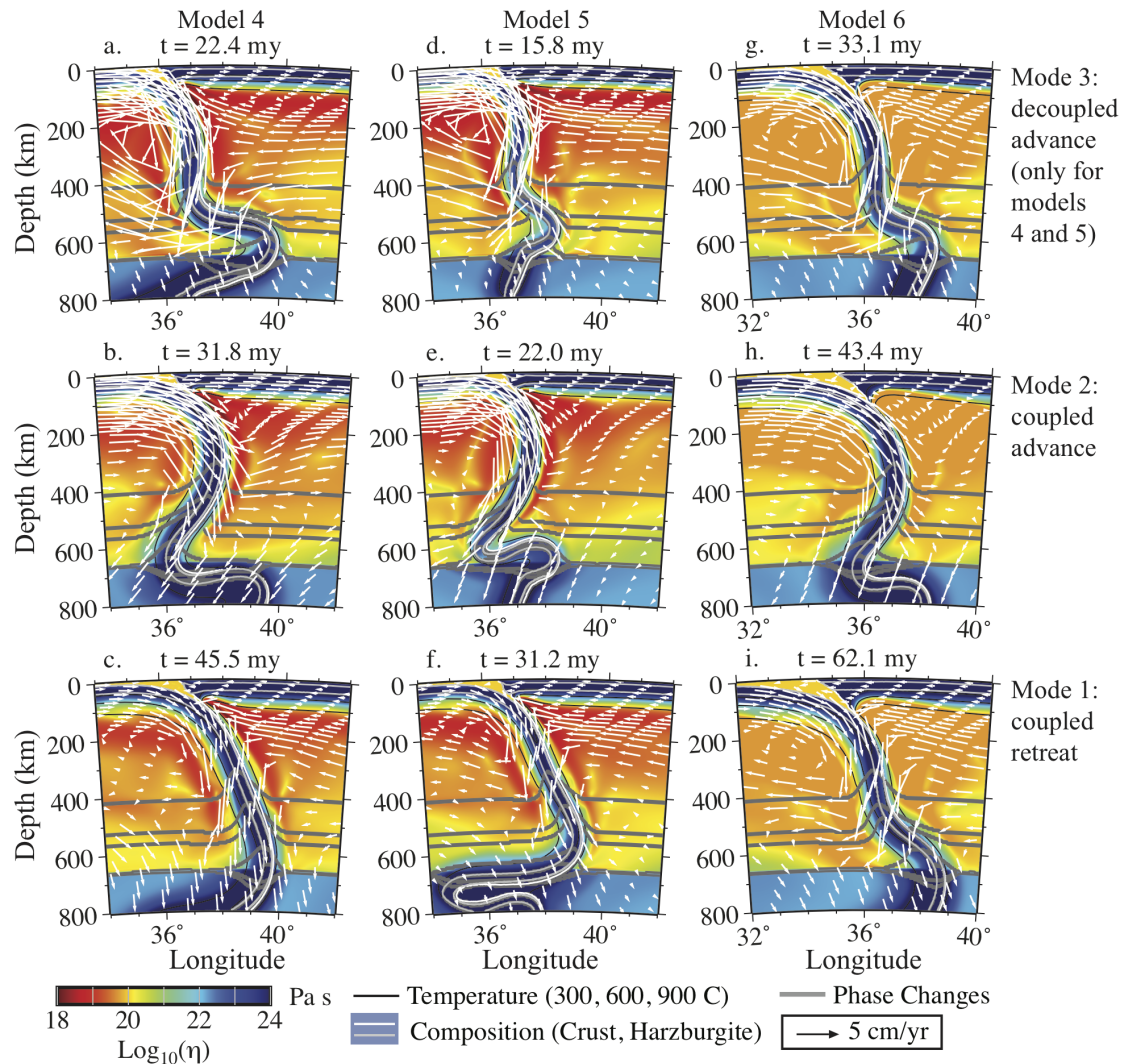


Figure 4: Time-evolution of coupling between slab, mantle and plate motion. a–c) Model 4: with overriding plate buoyancy, a spreading ridge and an old subducting plate (80 my). d–f) Model 5: same as model 4, but with younger subducting plate (40 my). .g–i) Model 6: same as model 4, but with a minimum viscosity cut-off of  $5 \times 10^{19}$  Pa s. Rows compare flow during similar states of slab evolution. Bottom row: mode 1 (coupled retreat) just before or at the start of slab folding both asthenosphere and overriding plate motion are toward the trench. Middle row: mode 2 (coupled advance) slab advance following sinking of a previous fold drives asthenosphere and overriding plate motion away from the slab. Top row: mode 3 (decoupled advance) slab buckling drives rapid asthenospheric flow toward the slab, but the overriding plate is moving in the opposite direction (except for model 6).

257 opposite direction (mode 3). For model 7 with no phase transitions and no slab folding, only coupled flow occurs  
 258 (mode 1 and 2), with a short period of advance followed by retreat. The subducting slab also has a larger component  
 259 of horizontal motion at shallow depth during coupled advance. Similar periods of coupled advance and coupled retreat  
 260 are seen in models 4–6, and are also correlated with the magnitude and depth extent of horizontal motion of the slab.  
 261 The fact that coupled advance occurs when there is a large component of positive horizontal (prograde) slab motion

262 indicates that the stiffness of the slab is also an important factor affecting the amount of trench advance in the models.  
263 For example, when the slab is mechanically weaker as in model 5, the prograde, horizontal motion of the slab and the  
264 advance rate of the overriding plate, are also smaller.

265 Models 4 and 5 also exhibit short periods of rapid trench advance while the underlying asthenosphere is flowing  
266 toward the slab (decoupled advance, mode 3). During this period, we see that three things are happening. First, the  
267 subducting plate velocity, and therefore also the slab sinking speed are high. Second, the slab is folding in opposite  
268 directions at the top of the slab (positive, prograde) compared to the bottom (negative, retrograde), as can be seen by  
269 horizontal slab motion. Third, there is stress-induced weakening of the asthenosphere beneath the overriding plate  
270 resulting in viscosity that drop below  $10^{19}$  Pa s. This suggests that the changing buoyancy forces and geometry of  
271 the slab results in internal weakening and folding of the slab. Folding of the slab causes rapid sinking and pulls the  
272 asthenosphere and mantle toward the folding slab, which has negative (retrograde) horizontal flow at larger depths.  
273 However, at shallow depths the slab has positive horizontal motion indicating that there is net subducting-plate and  
274 slab motion toward the overriding plate. Finally, because there is stress-induced weakening of the asthenosphere under  
275 the overriding plate, the motion of the overriding plate can decouple from the mantle flow being pulled toward the  
276 slab.

277 The important roll of the non-linear viscosity in allowing for this decoupled motion is seen by comparing models 4  
278 and 6. Limiting the minimum viscosity to  $5 \times 10^{19}$  Pa s prevents the decoupled motion from occurring, and the model  
279 exhibits net trench retreat. Also, we note that there is also stress-induced weakening in the asthenosphere beneath the  
280 subducting plate in models 4 and 5 (see supplemental Figs. S4 and S5). However, this weakening occurs for longer  
281 periods of time, not just during the times of decoupled advance, and not just during periods of faster subducting plate  
282 and slab motion. Therefore, this analysis supports the conclusion that it is the weakening beneath the overriding plate  
283 that is important in controlling the switch from mode 1 or 2 flow to mode 3 flow.

284 Together these three observations of model behavior during decoupled advance show that the sinking rate of  
285 the slab causes changes in asthenosphere viscosity through the non-linear rheology and that this in turn determines  
286 whether the asthenosphere and mantle are coupled or can flow in opposite directions. This suggests that when the mo-  
287 tion of the overriding plate and asthenosphere are decoupled, the excess prograde horizontal motion of the subducting  
288 plate is transferred directly to the overriding plate and can drive trench advance even though the asthenosphere is  
289 flowing in the opposite direction. Because the trench advance rate is faster in these periods of decoupled advance, the  
290 net trench motion is strongly affected by the short periods of decoupled plate motion. However, when the overriding  
291 plate and asthenosphere are coupled, asthenospheric drag on the overriding plate either works against the subducting  
292 plate to maintain trench retreat or helps to drag the overriding plate away from the subducting plate creating space for  
293 slow trench advance.

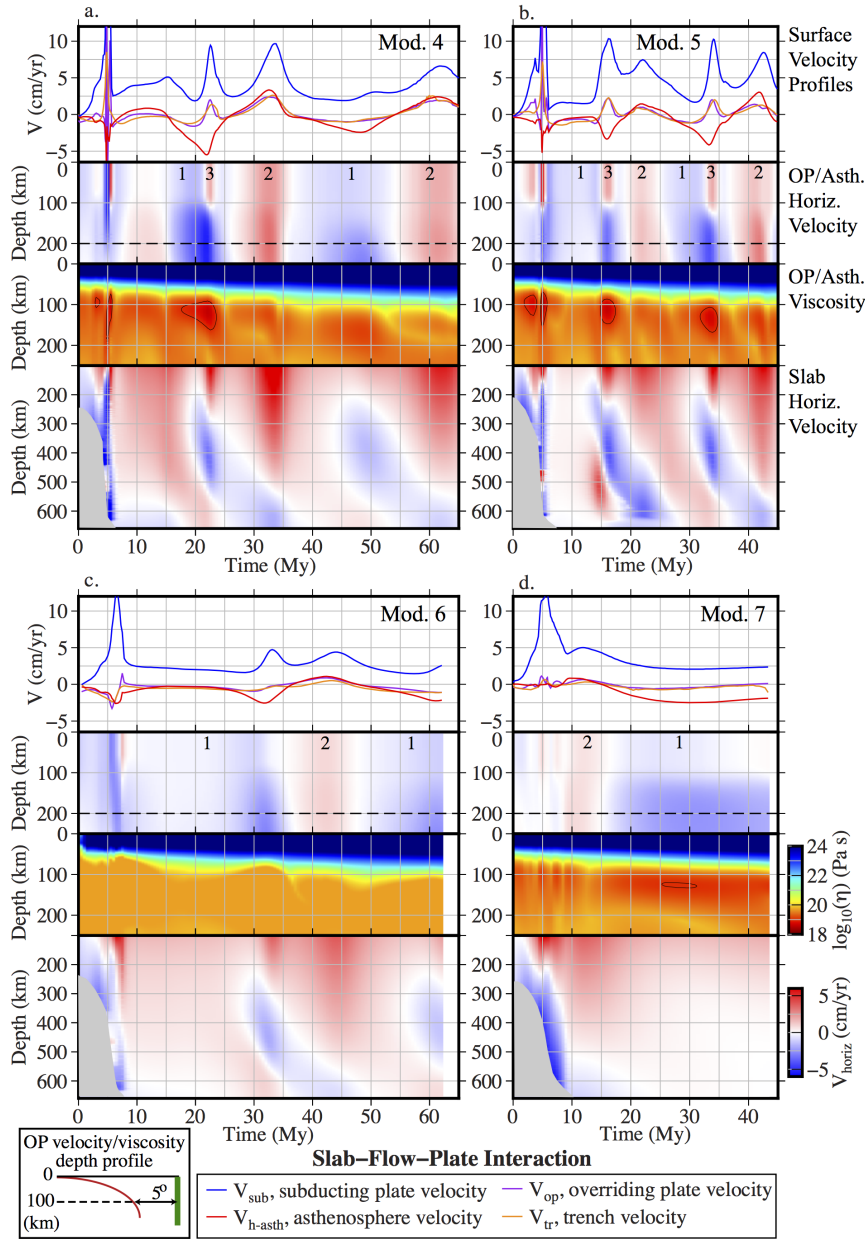


Figure 5: Effect of non-linear rheology on the coupling of slab, asthenosphere and plate flow. a–d) Models 4–7. Model 4 – reference case, model 5 – younger subducting plate, model 6 –  $\eta_{min} = 5 \times 10^{19}$  Pa s, model 7 – no phase transitions. Top: velocity profiles versus time. Positive velocity is away from the trench/slab. Negative velocity is toward the trench/slab.  $V_{h-asth}$  is the longitudinal (horizontal) component of flow in the asthenosphere at a depth of 200 km taken from the vertical profile located 5 degrees from the top of the slab-surface at 100 km depth (see green line in schematic). Numbers refer to mode 1–3 slab-flow-plate interaction. Middle-top: vertical profile of longitudinal (horizontal) component of flow versus time. Middle-bottom: vertical profile of viscosity versus time (same location as  $V_{h-asth}$  profile). Black contour outlines times when the viscosity decreases to  $\leq 10^{19}$  Pa s. Bottom: vertical profile (100–660 km depth) of the horizontal component of slab velocity versus time. The values are found by locating the point in the slab ( $T < 1000^\circ\text{C}$ ) with the maximum velocity magnitude at each depth.

### 294 3.2. Time-Dependent Thermal Structure

295 The episodic dynamics of the slab, plate and trench motion directly effects the time-dependent thermal structure  
296 of the subducting slab and the mantle wedge (Fig. 6). First, we find that the slab surface temperature (SST) depth  
297 profiles at a single time are quite similar to that found for steady-state kinematic wedge flow thermal models with  
298 similar parameters. The SST profiles are characterized by slow cooling of the slab at shallow depth, followed by rapid  
299 heating at the depth where the slab first comes into contact with hot mantle wedge material. In our models, this depth  
300 is primarily controlled by the basalt-to-eclogite transition, which determines the depth at which the crustal rheology  
301 is modeled to transition to that of olivine, but it also depends strongly on the minimum viscosity of the mantle. (This  
302 is similar in concept to case T550 from Syracuse et al. (2010) which modeled the partial-full coupling depth to occur  
303 when the slab surface temperature reached 550°C.)

304 For example, a model with relatively steady-state dynamics (model 7, no PT) has a thermal structure that is also  
305 uniform in time (Fig. 6a, i). Rapid heating occurs at 65–75 km depth with SSTs increasing from 300 to 750°C from  
306 80 to 100 km depth, but then only reaching to 900°C at 250 km. During a brief period (5–7 my) of faster subduction,  
307 which advects cold isotherms to greater depth, SSTs are colder by 75 °C at 100 km, but are substantially colder (almost  
308 250°C) at 75 km depth. However, after this time, there is little variation in the thermal profile as the slab dynamics  
309 are relatively steady. Interestingly even though this model has a relatively old slab (80 my), because it is subducting  
310 slowly (only 2.5 cm/yr), the SSTs predict melting of hydrated basalt over a narrow depth range (Fig. 6i).

311 In contrast, models with episodic slab folding exhibit substantially more variation in SSTs that primarily vary with  
312 the speed of the subducting plate (Fig. 6b, c). The hottest temperatures (up +225°C relative to the median temperature)  
313 occur early on during a period of very slow subduction immediately after the slab enters the lower mantle: at these  
314 times melting of the slab crust is predicted. The effect of slab age can be seen in slightly hotter SSTs for model 5  
315 (40 my) compared to model 4 (80 my), but this is a smaller effect than variations in subduction rate from 2 to 15  
316 cm/yr. During folding events the slab sinks rapidly (5–10 cm/yr) and SSTs drop by up to 250°C (at 100 km). These  
317 cold pulses can be followed along the slab surface to depths greater than 250 km and are reflective of much colder  
318 temperatures throughout the slab interior (see crust-mantle-boundary (CMB) profiles in Fig. 6i–l).

319 In addition to the coupling depth, the mantle viscosity has a strong effect on SSTs. In model 6, the higher minimum  
320 mantle viscosity causes the mantle wedge corner to deepen over time and this also shifts rapid heating of the slab  
321 surface from 75 km down to 100 km depth (Fig. 6d). Similar to model 7, because there is little variation in subduction  
322 rate, the SST profile is relatively constant in time. Also the median SST (deeper than the decoupling depth) is higher  
323 because the subduction rate is slower on average. At shallow depths, there is a slight shift to colder temperatures after  
324 35 my coinciding with a shift to larger slab curvature (smaller slab dip) and growth of the accretionary prism above  
325 the shallower slab (see Fig. 4j–l).

326 Mantle wedge temperature is also affected by the minimum mantle viscosity, through the effect on shallow slab  
327 shape, rate of mantle flow into the wedge corner, and growth of the accretionary wedge. In models with a minimum  
328 viscosity of  $10^{18}$  Pa s, flow in the mantle wedge corner is consistently shallower, predicting melting as shallow as 35

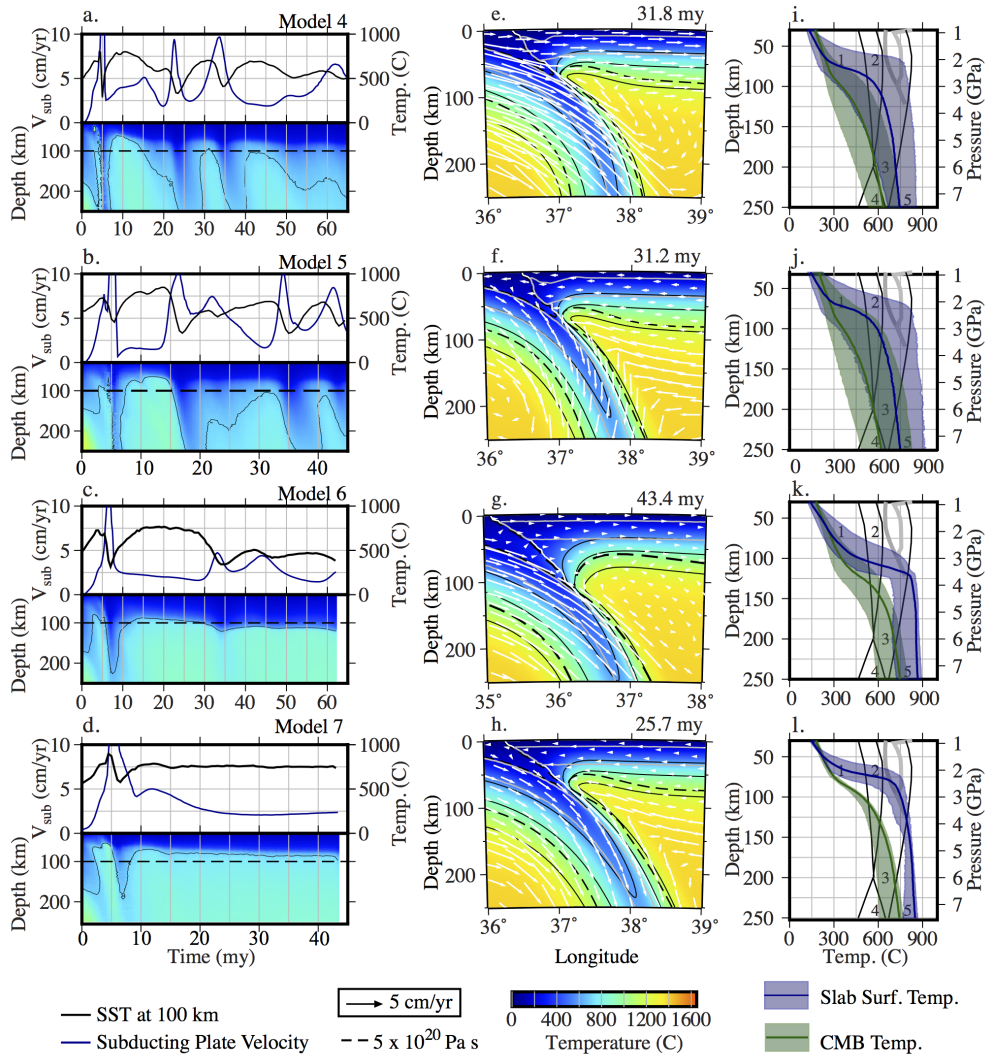


Figure 6: Time-evolution of slab-surface temperature (SST). Rows: 1–4 are Models 4–7. a–d) Bottom: temperature (color) along slab surface (depth) versus time. Black contour: 700 °C (temperature of wet basalt solidus at  $\approx 80$  km depth). Top: subduction velocity (blue) and SST at 100-km depth (black). Location of slab surface profile is shown in Fig. 1b and follows the crust contour. e–h) temperature in the slab and mantle wedge corner at times specified. Contours are the same as in Figure 4 with addition of a viscosity contour (dashed line). i–l) SST (blue) and crust-mantle boundary (CMB) temperature versus depth/pressure. Solid line: median temperature. Shading: range of minimum and maximum temperature. Dark/light gray lines are the solidi for wet/dry basalt (Vielzeuf and Schmidt, 2001). Black lines separating regions 1–5 are dehydration reaction for hydrated harzburgite (Hacker et al., 2003a): dehydration of serpentine/chlorite/brucite (1: 14.8% water), serpentine/chlorite dunite (2: 6.2 wt% water), chlorite/harzburgite (3: 1.4 wt% water), phase A (4: 6.8 wt% water), and dry garnet harzburgite (5).

329 km for wet peridotite and times when dry melting is possible (e.g., model 4; Fig. 7a, c). Note, the vertical profile  
 330 is located where the slab-surface crosses 100 km depth. In contrast, when the minimum viscosity of the mantle is  
 331 increased to  $5 \times 10^{19}$  Pa s, there is slow cooling of the mantle as the slab curvature increases from  $\approx 200$  km to  $\approx 400$   
 332 km (model 6, 10–30 my; Fig. 7b, d). This is followed by a more rapid increase in slab curvature from  $\approx 400$  km to

333  $\approx 600$  km (40–45 my) and a shift from hot mantle wedge conditions to a cold mantle wedge. In the cold state, even wet  
 334 melting of peridotite is not possible along this profile (but could occur along vertical profiles locate further down the  
 335 slab surface). However, even in the "hot" mantle state, the wedge profiles are cooler by  $\approx 100^\circ\text{C}$  (Fig. 7b) compared  
 336 to the models with a lower minimum mantle viscosity. In kinematic wedge thermal models, the mantle viscosity can  
 337 be changed independent of slab geometry; these fully-dynamic models demonstrate that these parameters are not in  
 338 fact independent and can lead to pronounced changes in slab-wedge thermal structure when there is self-consistent  
 339 time-dependent evolution of slab geometry.

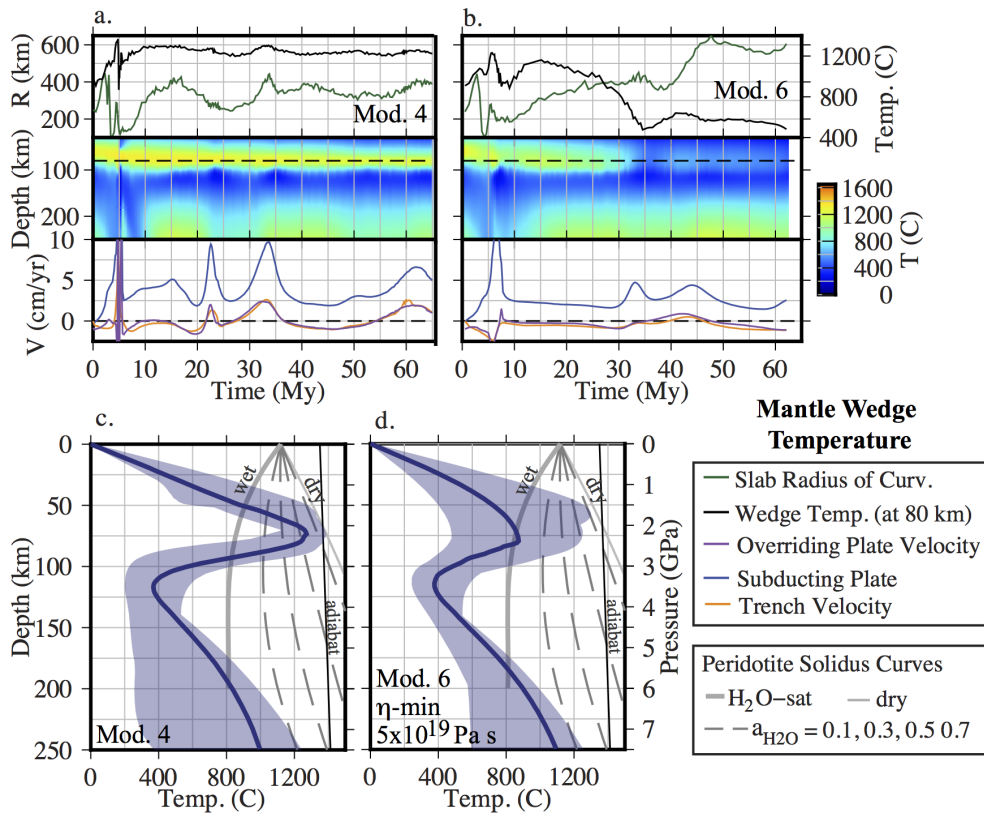


Figure 7: Time-evolution of mantle wedge temperature. a, c) Model 4. b, d) Model 6. a–b) Top: comparison of mantle wedge temperature at 80 km depth (black) to the radius of curvature of the slab (green: larger  $R$  indicates shallower dip). Middle: vertical profile of mantle wedge temperature versus time. Location of mantle wedge profile is shown in Fig. 1b. Bottom: plate and trench motion (see legend). c–d) Mantle wedge temperature versus depth/pressure. Dark blue: median temperature. Light blue shading: range of minimum and maximum temperature. Gray lines are the solidi (see legend) for wet to dry peridotite (Hirschmann, 2000; Till et al., 2010).

#### 340 4. Discussion

341 Determining the relationship between plate and trench motion, and slab deformation is important for determining  
 342 the key physical processes that govern coupling of large-scale mantle convection and plate tectonics (Conrad and

343 Lithgow-Bertelloni, 2002; G erault et al., 2012). From the comparison of the plate motions at the surface, the astheno-  
344 sphere flow and viscosity and the slab motion at depth, we find that there are three modes of slab-flow-plate interaction  
345 in the models:

346 Mode 1 (coupled retreat) occurs when the motion of the slab is primarily vertical sinking. In this case, slab sinking  
347 pulls the asthenosphere flow beneath the overriding plate toward the slab and the overriding plate is dragged  
348 toward the trench.

349 Mode 2 (coupled-advance) occurs when there is a large component of forward (prograde) motion of the slab along  
350 most of its length in the upper mantle. In this case, the prograde motion of the slab drives the asthenosphere  
351 flow away from the subducting slab, which also drags the overriding plate away from the trench. In all cases,  
352 there is also a horizontal component of forward motion across the shear zone pushing the overriding plate. Only  
353 in the mode 1 case is this motion overcome by the opposing flow-induced drag on the overriding plate. The  
354 proportion of forward motion needed to cause mode 2 flow depends on the stress-induced weakening of the  
355 asthenosphere.

356 Mode 3 (decoupled advance) occurs during initial stages of folding when the slab has a large component of prograde  
357 (positive) horizontal motion at shallow depths, while at deeper depths there is a large retrograde (negative)  
358 horizontal component. If the stresses induced by folding of the slab (seen as larger sinking rates) are large  
359 enough, then the non-linear weakening of the asthenosphere allows this prograde shallow motion to push the  
360 OP into advance, while the deeper retrograde motion pulls the underlying asthenosphere toward the slab.

361 There are three aspects of the physical model that combine to cause the three modes of slab-flow-plate interaction.  
362 The first aspect is the slab rheology, including both the total stiffness (integrated strength across its thickness) and the  
363 plastic yielding under large stress. The large stiffness of the slab is a result of the strong-temperature dependence of  
364 the olivine viscosity, a large yield stress (1 GPa) used to mimic Peierls creep at high stress (Billen and Hirth, 2007) and  
365 a maximum viscosity cut-off of  $10^{24}$  Pa s. The large stiffness leads to a significant component of prograde horizontal  
366 motion of the slab compared to a less stiff slab because the slab is strong enough to support its own weight when the  
367 bottom of the slab is supported by higher viscosity in the lower mantle. This effect can be seen by comparing  $V_{horiz}^{slab}$   
368 for an older, stiffer slab (Model 4) and a younger, less stiff slab (Model 5): the older slab has larger  $V_{horiz}^{slab}$  that extend  
369 to deeper depths. The plastic yielding of the slab is also important because it allows the slab to weaken and localize  
370 deformation under large stresses: this allows the folding to occur even though the rest of the slab remains strong.

371 The second aspect is the compositionally-dependent phase transition (CDPT) model. Compared to simpler pa-  
372 rameterizations of the phase transitions (e.g., olivine only with only two phase transitions), this more earth-like model  
373 provides less buoyancy-resistance at 660-km because of the counteracting effect of the pyroxene phase transitions at  
374 this depth. This difference is important because the slab is able to sink into the lower mantle and therefore it also  
375 creates a large return flow that includes the lower mantle: if the slab stagnates at the base of the transition, the return

376 flow is isolated in the upper mantle and must drive more horizontal flow to balance the same amount of slab sinking  
377 in the upper mantle. The CDPT model also leads to a more distributed increase in the density of the slab starting at  
378 410-km down through 520-km. A model with a the same density anomaly located only at the 410-km would tend to  
379 steepen the shallow slab (same force applied at shorter distance leads to more torque). This effect is demonstrated  
380 by the test model with olivine-only phase transitions: the slab stops sinking when it reaches 660-km and the stress  
381 transferred to the slab above 410-km is sufficient to cause the slab to break.

382 Finally, the third key aspect of the model that is the slab-induced, non-linear weakening of the mantle. The slab  
383 density is the main source of stress (e.g., besides ridge push) driving flow in the model. The viscosity structure con-  
384 trols how this stress is distributed (instantaneously) across the fluid, roughly decreasing exponentially with distance.  
385 However, when the viscosity is stress-dependent, this same stress can be accommodated at higher strain-rate leading  
386 to a lower effective viscosity. Because the non-linear viscosity also depends on temperature and increases with depth,  
387 the weakening response occurs primarily directly next to the slab and in the mantle wedge, where the stress is the  
388 highest. However, the weakening can also occur farther from the slab in the asthenosphere, because at shallow mantle  
389 depths and mantle temperatures the strain-rate accommodated by the dislocation creep (non-linear) is significantly  
390 higher than for diffusion creep (linear). This difference in strain-rates gets smaller with depth because the dislocation  
391 creep mechanism has a larger activation volume (viscosity increases faster with depth).

392 The significant impact of non-linear viscosity on slab dynamics and its ability to decouple mantle and plate flow  
393 direction has also been noted in instantaneous regional subduction models (Jadamec and Billen, 2010) and models of  
394 LPO (lattice-preferred orientation) development during the early stages of subduction (Jadamec, 2016; MacDougall  
395 et al., 2017). In addition, because changes to the strength of non-linearity (approximated in our models by limiting  
396 the minimum viscosity) affects trench motion and the subducting plate speed in the models, it may be possible to use  
397 these observations, together with other constraints on slab geometry, to better constrain lateral viscosity variations and  
398 absolute viscosity of the asthenosphere. For example, while model 6, does a better job of generating trench retreat,  
399 the maximum subducting plate speed is only 2.6 cm/yr, far below the maximum subducting plate speeds observed in  
400 the present-day (Lallemand et al., 2005). Therefore, the viscosity structure would need to change in such a way as  
401 to allow for faster subducting plate speeds, and more trench retreat. In particular, the models presented here assume  
402 uniform grain-size, do not include the effects of variations in water content (e.g., a wetter mantle wedge) or presence  
403 of melt, and use the same flow law for all the major compositions (except basalt). Thus, these model results provide  
404 the motivation for future studies that can systematically test the effects of more comprehensive rheological models,  
405 with a suite of observations that are sensitive to different parameters.

406 These results also present new challenges to applying an analytical force-balance approach to understanding how  
407 slabs drive plate motions. First, rather than ignoring the overriding plate in force-balance calculations (Capitanio,  
408 2013), it must be considered in the analysis because slab-induced mantle flow affects the motion of the overriding  
409 plate and the resulting force works across the shallow plate interface to affect trench motion. During coupled phases,  
410 when the mantle flow pulls the overriding plate toward the trench it helps to prevent trench advance and drive trench

411 retreat, whereas in decoupled phases, even though the asthenosphere flows toward the trench the overriding plate can  
412 be pushed by the advancing subducting plate. The ability of the overriding plate to affect trench motion has been  
413 demonstrated in other studies, for example, using a similar model set-up and non-linear rheology, but limiting the  
414 minimum viscosity to  $10^{19}$  Pa s (Cížková and Bina, 2013). This conclusion is also supported by instantaneous models  
415 of subduction in a 2D cylindrical geometry with linear viscosity, which showed that the ability of strong slabs to  
416 drive trench advance is increased when there is a weak asthenosphere beneath the overriding plate (Gérault et al.,  
417 2012): that study concluded that this was because the weak asthenosphere partially decouples the plate and mantle  
418 and emphasizes the interaction between adjacent plates.

419 Second, rather than assessing the instantaneous force balance given a specified slab geometry, it is necessary to  
420 take into account how the evolving slab-induced stresses on the mantle can change the mantle viscosity structure. Both  
421 this study and the study by Jadamec (2016) show that the viscosity can be reduced by 10–100x, not just immediately  
422 around the slab, but reaching 500–1000 km from the slab beneath the plates. The models presented here show that  
423 these changes in viscosity structure can be rapid, and are happening continuously. However, the largest changes in  
424 viscosity in the models, those leading to complete decoupling of plate motion direction, are short-lived and relate to  
425 large changes in slab geometry, suggesting that such events may also be less common in the Earth (e.g., the time-  
426 scale of folding found in the models: 25–50 my). For example, if such a weakening event is necessary to allow for  
427 trench advance, than this could be one reason why trench advance is less prevalent in present-day observations of  
428 trench motion. More generally, however, other processes besides slab folding could also lead to weakening in the  
429 asthenosphere.

430 Analysis of the shallow thermal structure is in general agreement with the slab surface and mantle wedge tem-  
431 peratures predicted by kinematic-slab wedge-flow thermal models (Wada et al., 2008; Syracuse et al., 2010): thermal  
432 structure is primarily controlled by subduction rate, with secondary effects due to slab age and moderate changes  
433 in slab dip. However, unlike kinematic models there is a feedback between the evolving temperature and viscosity  
434 structure and the slab geometry. In particular, limiting the minimum viscosity within the mantle wedge causes the  
435 overriding plate to be dragged down into the mantle creating a broad, cold mantle wedge corner. While similar struc-  
436 tures sometimes appear in published subduction models, the dynamic effects and the implications for melting in the  
437 mantle wedge (Elkins-Tanton et al., 2001; Kelemen et al., 2003; Cagnioncle et al., 2007; England and Katz, 2010),  
438 are not commonly addressed. Our results show, that the shallow wedge thermal structure is another important con-  
439 straint on the upper mantle rheology. For example, from the behavior of model 6, it can be argued that if the effect of  
440 non-linear viscosity is generally more limited in the upper mantle, the wedge viscosity is probably kept weaker by the  
441 effects of water and/or melt (Billen and Gurnis, 2001; Arcay, 2017).

442 The models presented here also highlight the degree to which slab thermal structure can vary in response to  
443 changes in slab sinking rate. First, slab sinking rates vary from 2 to  $> 10$  cm/yr in response to changing slab geometry  
444 during folding, buckling and recovery of the slab in the transition. Such long-term variation in the subduction rate  
445 (25–50 my) could contribute to the long-term cyclical patterns observed in arc magmatism (Haschke et al., 2002;

446 DeCelles et al., 2009). In particular, most of the models predict possible melting of basaltic crust at the surface of the  
447 slab during periods of very slow subduction ( $< 2$  cm/yr). However, they also predict substantial drops in SST during  
448 periods of fast subduction (5–10 cm/yr), which could severely limit dehydration of the slab (van Keken et al., 2002),  
449 loss of other volatiles (e.g., carbon Dasgupta and Hirschmann, 2010), and water-induced melting in the mantle wedge  
450 (Schmidt and Poli, 1998; Till et al., 2010, 2012). These results also show that the amount of water retained in slabs  
451 sinking deeper into the mantle could be highly variable. Such variability in water content could impact intermediate  
452 depth seismicity in the slab (Hacker et al., 2003b; Omori et al., 2004), release of water in the transition zone (Richard  
453 et al., 2007; Leahy and Bercovici, 2007; Myhill et al., 2017) and transport of water into the lower mantle (Hirschmann,  
454 2006).

455 The model analysis highlights how the differences between the model design used in our studies and that used  
456 by other research groups has lead to the different slab dynamics, and plate and trench motions. The next question is  
457 then, how relevant are these models to understanding slab dynamics and plate motions in the Earth. As of yet, neither  
458 our models, nor other existing model do a satisfactory job of self-consistently matching the suite of observations  
459 characterizing slab dynamics and plate motion on Earth: the variation in slab shape reflecting differences in time-  
460 dependent deformation, correlations between slab shape, density, sinking rates, and subduction duration, the range of  
461 plate speeds, the relationship between different directions and rates of trench motion and slab deformation, and the  
462 thermal structure of the shallow slab and mantle wedge. And, we know of several simplifications that are likely to  
463 affect the ability of a model to reproduce not just a couple, but all of these observations. Most important amongst  
464 these simplifications for our models are: 1) the use of a 2D rather than 3D domain, 2) a model domain with sidewalls  
465 rather than a full sphere, 3) multiple surface plates and slabs and 4) a passive, lower mantle lacking density structures  
466 and thermal boundary conditions that would lead to both large-scale and small-scale upwelling.

467 However, we suggest that the models presented here bring us closer to earth-like subduction for several reasons.  
468 First, the models use an earth-like non-linear rheology, which is known to be active in the upper mantle, and which  
469 we have now shown plays a fundamental role in the coupling of plate-flow-slab interactions. Models that do not  
470 include non-linear rheology may be able to match some observations, but they tend to have very cold mantle wedges.  
471 More studies on the effect of non-linear rheology are necessary to better capture and understand the variables that  
472 control slab deformation: e.g., grain-size, Peierls creep versus a yield criterion, effects of water and melt in the mantle  
473 wedge. Second, we use the CDPT, which better represents an earth-like density distribution in the slab. Many studies  
474 have clearly shown that slab buoyancy is a fundamental control on slab dynamics and trench motion (Schellart, 2008b;  
475 Stegman et al., 2010; Ribe, 2010): therefore, using an earth-like density distribution and not just an averaged density is  
476 necessary to apply model results to slab dynamics on Earth. Third, our models produce both retreating and advancing  
477 trench motion with rates of about 1–2 cm/yr, and plate velocities in the range of 1–15 cm/yr after the slab reaches the  
478 lower mantle, in agreement with present-day observaitons. However, the models do not exhibit steady, longer term  
479 trench retreat or fast trench retreat rates. Therefore, further analysis are needed to understand how these other modes  
480 of slab-flow-plate interaction are achieved in models with more earth-like rheology and phase transitions. Fourth,

481 our models produce folded slabs that sink slowly into the lower mantle, in agreement with the interpretation of many  
482 thick slab anomalies in the lower mantle (van der Hilst et al., 1997; Ritsema et al., 2004; Fukao et al., 2009; Simmons  
483 et al., 2015). Finally, the models presented here have slab and mantle wedge thermal structure that are consistent  
484 with previous kinematic thermal models that match a variety of temperature-related observations (e.g., dehydration,  
485 melting, surface heat flow). This is further evidence that the non-linear rheology, which is a key factor controlling the  
486 slab-wedge thermal structure, through its effect on the slab sinking rates and mantle flow rates, is a first order feature  
487 of an earth-like model of a subduction zone.

## 488 **5. Conclusions**

489 Using 2D fully dynamic models of subduction we have shown that there is a feedback between slab deformation  
490 and stress-induced weakening of the asthenosphere underlying the overriding plate through non-linear rheology. We  
491 find that there are three modes of interaction between the slab deformation, mantle flow and overriding plate motion.  
492 In mode 1 and mode 2 the overriding plate and underlying mantle have coupled motion with either trench retreat or  
493 advance, respectively. In mode 3, weakening of the asthenosphere allows the mantle to flow toward the retrograde  
494 motion of the folding slab, while the overriding plate is pushed by prograde advancing motion of the shallow slab  
495 and subducting plate. This kind of dynamically-controlled viscous resistance in the upper mantle complicates force-  
496 balance analysis of trench and plate motion, but also highlights the fact that the overriding plate is not passive and  
497 can help to drive trench retreat or advance (coupled) or allow rapid trench advance (decoupled). In addition, we find  
498 that modifying the overriding plate structure to include the ridge push force and crustal buoyancy leads to more trench  
499 retreat. However, because of the decoupling effects of the rheology and the effects of compositionally-dependent  
500 phase transitions on slab density, trench motion oscillates in response to slab folding without net trench retreat over  
501 time. Therefore, we conclude that the effects of non-linear rheology may be less-pronounced in the Earth's upper  
502 mantle in order to match observations of trench retreat. However, non-linear viscosity also affects the subducting  
503 plate speed and thermal structure in the mantle wedge. Therefore, reducing the effects of the non-linear rheology  
504 could also have the unwanted effect of producing plate speeds that are too slow and a mantle wedge that can not  
505 produce melts. Therefore, an increase in asthenosphere viscosity beneath the overriding plate should be added to  
506 models, while still allowing the mantle to weaken around the slab and in the mantle wedge. Finally, the episodic  
507 sinking rate of the slab causes significant time variability in thermal structure of the shallow slab system predicting  
508 periods of reduced volatile-induced melting in the mantle wedge and increased transport of volatiles to the transition  
509 zone and deep mantle.

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