

# Effects of composite rheology on plate-like behavior in global-scale mantle convection

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## Key Points:

- Uppermost mantle viscosity variations induced by composite rheology control surface tectonics
- Composite rheology can impede or enhance plate mobility depending on lithospheric strength
- Composite rheology does not facilitate the onset of subduction for large yield stress

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**Abstract**

Earth's upper mantle rheology controls lithosphere-asthenosphere coupling and thus surface tectonics. Rock deformation experiments and seismic anisotropy measurements indicate that composite rheology (co-existing diffusion and dislocation creep) occurs in the Earth's uppermost mantle, potentially affecting convection and surface tectonics. Here, we investigate how the spatio-temporal distribution of dislocation creep in an otherwise diffusion-creep-controlled mantle impacts the planform of convection and the planetary tectonic regime as a function of the lithospheric yield strength in numerical models of mantle convection self-generating plate-like tectonics. The low upper-mantle viscosities caused by zones of substantial dislocation creep produce contrasting effects on surface dynamics. For strong lithosphere (yield strength  $>35$  MPa), the large lithosphere-asthenosphere viscosity contrasts promote stagnant-lid convection. In contrast, the increase of upper mantle convective vigor enhances plate mobility for lithospheric strength  $<35$  MPa. For the here-used model assumptions, composite rheology does not facilitate the onset of plate-like behavior at large lithospheric strength.

**Plain Language Summary**

Understanding uppermost mantle flow and deformation is important to study Earth's surface evolution, since plate tectonics and mantle convection are intertwined processes. Observations and experiments provide important - yet uncertain - constraints suggesting that uppermost mantle viscosity should be at least partially controlled by dislocation creep (i.e. its rheology should vary non-linearly with stress). However, most studies have not included dislocation creep. Here, we incorporate different amounts of this deformation mechanism in global-scale numerical models of mantle convection featuring Earth-like tectonic plates. We demonstrate that fast-evolving low-viscosity areas containing dislocation creep arise around slabs and plumes. Moreover, large amounts of dislocation creep alter surface tectonics in several ways: for a weak lithosphere, subductions become shorter-lived and plate velocities increase. For a strong lithosphere, in contrast, plate tectonics is inhibited. This study therefore demonstrates the key role of composite rheology in understanding mantle-lithosphere interactions.

## 1 Introduction

The lithospheric behavior of terrestrial bodies notably depends on their mantle properties and dynamics (e.g. Alisic et al., 2012; Coltice et al., 2017; Garel et al., 2020). In particular, mantle rheology determines the coupling between the convecting mantle and the lithosphere, therefore affecting surface heat transfer, plate velocities and continental motions (e.g. Stein et al., 2004; Rolf et al., 2018). Rock-deformation laboratory experiments conducted at upper-mantle conditions (Fig 1a-b, e.g. Hirth & Kohlstedt, 2003; Karato & Wu, 1993) show that mantle rheology is composite, meaning that deformation is driven by a coexistence of different creep mechanisms such as diffusion creep (linear or Newtonian stress/strain-rate dependence) and dislocation creep (non-linear power-law or non-Newtonian stress/strain-rate relationship). These experimental results are corroborated by the observed spatial heterogeneity in the strength of uppermost-mantle seismic anisotropy (e.g. Beghein et al., 2014; Debayle & Ricard, 2013), which could be at least partially explained by different amounts of olivine lattice preferred orientations (LPO), possibly caused by the heterogeneous development of dislocation creep in the uppermost mantle (e.g. Becker et al., 2006; Hedjazian et al., 2017; Nicolas & Christensen, 1987).

While mantle composite rheology is typically considered in regional-scale geodynamics models (e.g. Billen & Hirth, 2005; Garel et al., 2020; Neuharth & Mittelstaedt, 2023), it is often neglected in global-scale models (e.g. Coltice et al., 2017; Li & Zhong, 2019; Stein et al., 2004), or simply mimicked by reduced activation energy in pure diffusion creep rheology (Christensen, 1983, 1984). However, this latter approximation causes differences in the planform of stagnant-lid convection compared to using full composite rheology (e.g. Schulz et al., 2020). Moreover, prescribing pure diffusion creep makes it difficult to fully capture Earth's lithosphere and mantle behavior, such as observed plume swells' shapes (Asaadi et al., 2011), trench retreat rates (Holt & Becker, 2016), seismic anisotropy patterns around slabs (Jadamec & Billen, 2010), surface dynamic topography amplitudes (e.g. Bodur & Rey, 2019), and subduction geometry during its initiation (e.g. Billen & Hirth, 2005). Numerical studies prescribing pure dislocation creep in the upper-mantle have shown its importance for all these processes. However, in a composite formulation, the spatiotemporal distribution of the different creep mechanisms is not determined a priori, but arises self-consistently. Accounting for it therefore allows us to evaluate where substantial dislocation creep may occur in the mantle and to fur-

79 ther study its effects on geodynamic processes. Some global models of mantle convec-  
 80 tion with plate-like behavior recently included composite rheology (e.g. Dannberg et al.,  
 81 2017; Rozel, 2012), but these computationally-demanding models used a single set of rhe-  
 82 ological activation parameters based on experimental values, while estimates vary over  
 83 a large range (e.g. Ranalli, 2001; Korenaga & Karato, 2008; Jain et al., 2018, 2019). More-  
 84 over, these numerical studies focussed on the effect of grain-size evolution on the plan-  
 85 form of convection and on the lithospheric behavior. Therefore, a systematic exploration  
 86 of the effects of composite rheology in the upper mantle is still needed.

87 Here, we explore how the temperature-, depth- and stress-dependent diffusion/dislocation  
 88 creep partitioning impacts the planform of convection and the tectonic regime in 2D-cartesian  
 89 whole-mantle convection models with composite rheology and static grain-size self-generating  
 90 plate tectonics. Our goal is not to use Earth-like rheological parameters, but rather in-  
 91 vestigate the geodynamic effects of different parametrizations of composite rheology and  
 92 capture qualitative convective and tectonic trends relevant for the Earth (Fig. 1). We  
 93 find that composite rheology influences both mantle convective planform and surface tec-  
 94 tonics due to its spatio-temporal dynamic effect on uppermost mantle viscosity, either  
 95 enhancing or altering plate mobility and plateness depending on lithospheric strength.  
 96 These results demonstrate that uncertainties in experimentally-determined rheological  
 97 parameters lead to substantial geodynamical effects, and calls for further consideration  
 98 of composite rheology in studies of mantle-lithosphere interactions.

## 99 2 Methods

### 100 2.1 On the use of composite rheology

101 Mantle viscosity varies with temperature ( $T$ ), pressure ( $P$ ), grain-size ( $d$ ) and stress  
 102 ( $\sigma$ ) (e.g. Hirth & Kohlstedt, 2003; Karato & Wu, 1993):

$$\eta_{mech} = A_{mech} d^m \sigma^{1-n} \exp\left(\frac{E_{mech} + PV_{mech}}{RT}\right). \quad (1)$$

103  $R$  is the gas constant,  $m$  is the grain-size exponent and  $n$  is the stress exponent.  $E_{mech}$ ,  
 104  $V_{mech}$  and  $A_{mech}$  are respectively the activation energy, the activation volume and a pre-  
 105 exponential factor (accounting for all other effects on mantle rheology, such as water and  
 106 melt content) for the rheological mechanism ( $mech$ ) considered (diffusion or dislocation  
 107 creep).

108 Diffusion creep dominates below and dislocation creep dominates above the tran-  
 109 sition stress ( $\sigma_t$ ) at which the strain-rates due to the two different mechanisms are equal  
 110 ( $\dot{\epsilon}_{diff} = \dot{\epsilon}_{disl}$ , e.g. Christensen, 1984; Hall & Parmentier, 2003):

$$\sigma_t = \left( \frac{A_{diff}}{A_{disl}} \right)^{\frac{1}{n-1}} d^{-\frac{m}{n-1}} \exp \left( \frac{(E_{disl} - E_{diff}) + P(V_{disl} - V_{diff})}{RT} \right)^{\frac{1}{n-1}}. \quad (2)$$

111  $E_{diff}$ ,  $E_{disl}$ ,  $V_{diff}$  and  $V_{disl}$  can be determined for olivine from rock experiments (e.g.  
 112 Karato & Wu, 1993) and vary respectively between 240–450 kJ/mol, 430–560 kJ/mol,  
 113 0 – 20 cm<sup>3</sup>/mol and 0 – 33 cm<sup>3</sup>/mol (Hirth & Kohlstedt, 2003; Karato & Wu, 1993;  
 114 Ranalli, 2001), depending on water content. Despite those uncertainties,  $E_{diff} < E_{disl}$   
 115 and  $V_{diff} < V_{disl}$  (Fig. 1a-b, e.g Karato & Wu, 1993). Those experiments predict that  
 116 dislocation creep should dominate in hot regions of the uppermost-mantle and areas sub-  
 117 mitted to high stresses (Fig. 1a-b).

## 118 2.2 Numerical model setup

119 We solve the non-dimensional equations of mass, momentum and energy conser-  
 120 vation under the Boussinesq approximation using StagYY (e.g. Tackley, 2000a) on a 2D-  
 121 cartesian 512x128 or 768x192 grid (aspect ratio 4:1). Grid cells are refined near the ther-  
 122 mal boundary layers. Top and bottom boundaries are free-slip, lateral boundaries are  
 123 periodic. We use a reference Rayleigh number of 10<sup>7</sup>. The mantle is heated both from  
 124 below and from within (constant internal heating rate  $H = 8.6 \times 10^{-12}$  W kg<sup>-1</sup>, Ta-  
 125 ble S1).

126 We use a pseudoplastic rheology to model plate-like behavior (e.g. Trompert & Hansen,  
 127 1998; Tackley, 2000a), and vary the surface yield stress  $\sigma_{Y_0}$ , which represents lithospheric  
 128 strength, between 12 and 234 MPa. The yield stress varies with depth at a rate of  $\sim 0.3$   
 129 MPa km<sup>-1</sup>. The surface yield stress is bounded by the typical stress drop during earth-  
 130 quakes (10 MPa, Allmann & Shearer, 2009) and the yield stress of pristine lithospheric  
 131 rocks measured in experiments (Brace & Kohlstedt, 1980). Over the modeled range of  
 132 yield stresses, diverse tectonic behaviors are expected for pure diffusion creep: from mo-  
 133 bile plates at low yield stress to stagnant-lid at high yield stress (e.g. Arnould et al., 2018).

134 In StagYY, the transition stress  $\sigma_t^*$  between diffusion and dislocation creep is de-  
 135 fined in analogy to Eq. 2 as:

$$\sigma_t^* = \sigma_0 \left( \frac{B_{disl}}{B_{diff}} \right)^{\frac{1}{n-1}} \left( \frac{d}{d_0} \right)^{-\frac{m}{n-1}} \exp \left( \frac{(E_{disl} - E_{diff}) + P(V_{disl} - V_{diff})}{R(T + T_0 - T_{surf})} \right)^{\frac{1}{n-1}}. \quad (3)$$

136  $T_0 = 0.64$  is the non-dimensional reference temperature, equivalent to 1,600 K and  $T_{surf} =$   
 137 0.12 is the non-dimensional surface temperature, equivalent to 300 K.  $\sigma_0$  is a reference  
 138 transition stress.  $B_{diff}$  and  $B_{disl}$  differ from  $A_{mech}$  in Eq. 1 and ensure that mantle vis-  
 139 cosity equals the non-dimensional reference viscosity  $\eta_0 = 1$  ( $9.8 \times 10^{21}$  Pa s) at refer-  
 140 ence conditions (temperature of 1,600 K and surface pressure). As we do not account  
 141 for grain-size evolution,  $d = d_0$  unless explicitly mentioned otherwise (see Discussion  
 142 in section 4). For dislocation creep,  $m=0$  and  $n=3.5$  while for diffusion creep,  $m=2$  and  
 143  $n=1$ .

### 144 **2.3 Computed cases**

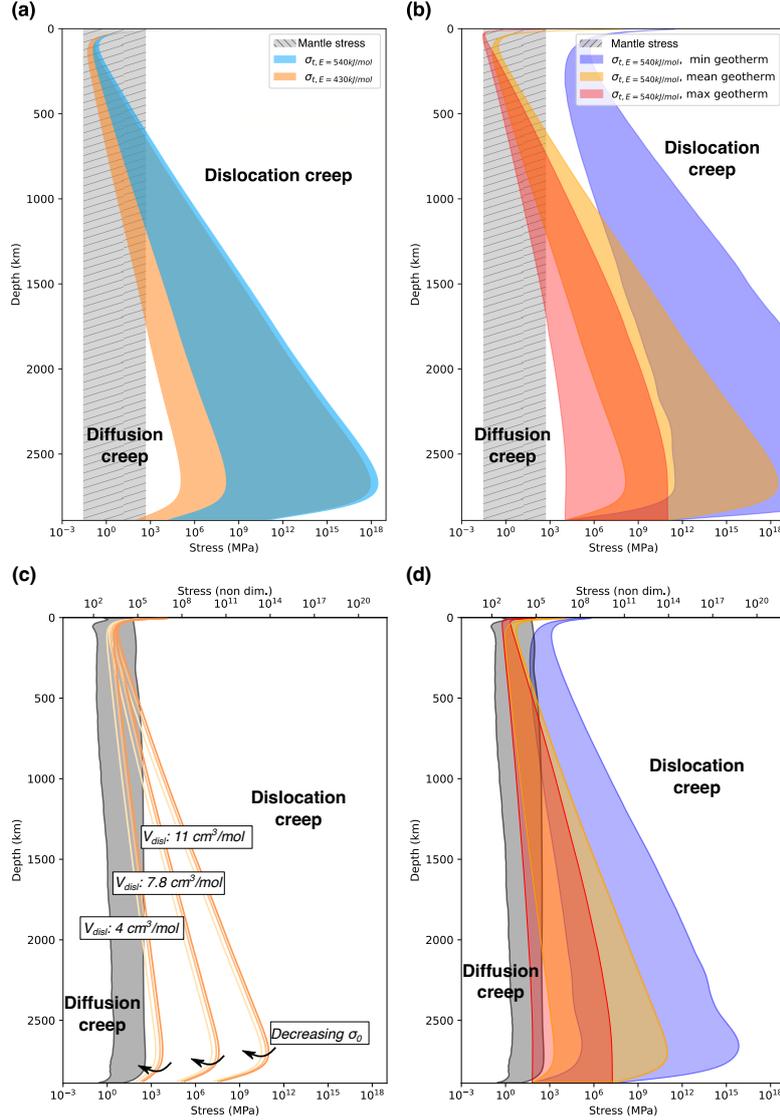
145 For each value of  $\sigma_{Y_0}$ , we fix  $E_{diff}$ ,  $V_{diff}$ , and  $E_{disl}$ , but vary  $V_{disl}$  by a factor of  
 146  $\sim 3$ , since its experimental value is subjected to the largest uncertainties (e.g. Karato &  
 147 Wu, 1993; Korenaga & Karato, 2008). We also vary  $\sigma_0$  between 1.2 and 3.5 MPa to en-  
 148 sure that dislocation creep is mostly restricted to the upper mantle. We choose lower ac-  
 149 tivation parameters than experimentally determined for pristine olivine for reasons of  
 150 numerical feasibility. Instead, we preserve ranges of variation for  $E_{disl} - E_{diff}$  and  $V_{disl} -$   
 151  $V_{diff}$  similar to rock experiments (Karato and Wu (1993), Fig. 1) since these differences  
 152 matter the most in Eq. 2 and 3. The spatio-temporal evolution of mantle convection self-  
 153 consistently partitions the mantle into areas dominated by dislocation creep or diffusion  
 154 creep, depending on the value of stress.

155 For each yield stress, we first ran models in pure diffusion creep over 3 Gyr, start-  
 156 ing from a stratified thermal field with small perturbations to initiate convection. We  
 157 then restarted from the final thermal field of these models while including composite rhe-  
 158 ology and ran those new models over 3 Gyr. Since we do not model evolutionary mod-  
 159 els, this procedure ensures that the models are in quasi-statistical steady-state (Fig. S2)  
 160 during the last 400 Myr of each simulation that we analyse. Detailed model parameters,  
 161 with their non-dimensional and dimensional values are given in Table S1.

## 162 **3 Results**

### 163 **3.1 Spatio-temporal distribution of dislocation creep**

164 Decreasing both  $V_{disl}$  and/or  $\sigma_0$  results in a thicker and more continuous layer de-  
 165 forming in dislocation creep in the upper mantle (Fig. 2). As a consequence, upper man-

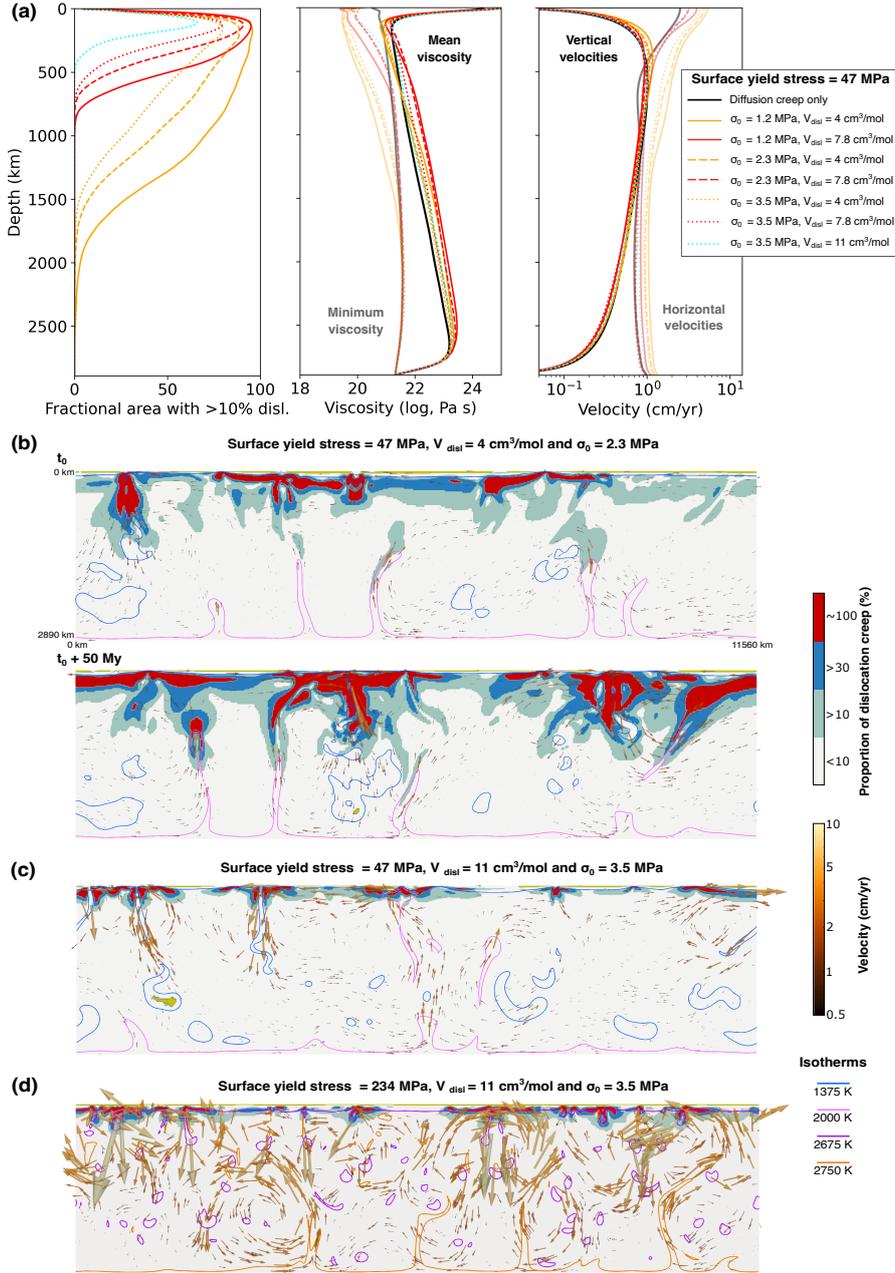


**Figure 1.** **Top:** Range of olivine transition stress measured by Karato and Wu (1993), assuming a grain-size of 1 mm. (a) Sensitivity to  $E_{disl}$  (blue: 430 kJ mol<sup>-1</sup>, orange: 540 kJ mol<sup>-1</sup>) and  $V_{disl}$  (10-25 cm<sup>3</sup> mol<sup>-1</sup>), using an average geotherm from a reference model in pure diffusion creep (Fig. S1e). (b) Sensitivity to temperature using  $E_{disl} = 430$  kJ mol<sup>-1</sup> and  $10 < V_{disl} < 25$  cm<sup>3</sup> mol<sup>-1</sup> (blue=cold, yellow=average, and red=hot geotherm (Fig. S1e)). **Bottom:** Same as above, but for our modeling setup. (c) Sensitivity of the model transition stress to  $V_{disl}$  (4-11 cm<sup>3</sup> mol<sup>-1</sup>) and  $\sigma_0$  (1.2-3.5 MPa), using an average geotherm (Fig. S1e). (d) Sensitivity to temperature. In all panels, gray-striped areas show the stress range expected in Earth’s mantle (top) and predicted in our reference model (bottom, Fig. S1b).

166 the viscosity decreases by at least one order of magnitude on average. Moreover, aver-  
167 age horizontal and vertical velocities increase by a factor of 3 depending on the amount  
168 of dislocation creep (Fig. 2a), irrespective of the surface yield stress (Fig. S3), showing  
169 that composite rheology enhances convective vigor locally. Due to its location and low  
170 viscosity signature, the layer containing  $>10\%$  dislocation creep is here-after referred to  
171 as an “asthenosphere” in models with composite rheology, although it sometimes locally  
172 reaches lower-mantle depths (low  $V_{disl}$  and  $\sigma_0$ ).

173 Areas strongly affected by dislocation creep show a high spatio-temporal variabil-  
174 ity within the asthenospheric layer (Fig. 2b-d and Supplementary Movie 1), which pro-  
175 duces large lateral viscosity variations in the upper mantle, as shown by e.g. Alisic et  
176 al. (2012); Billen and Hirth (2007); Semple and Lenardic (2020). In models featuring plate-  
177 like behavior, dislocation creep mainly occurs around slabs and plumes in the uppermost  
178 mantle. Indeed, ambient mantle shearing by sinking slabs is responsible for the highest  
179 convective stresses, and thus for a higher proportion of dislocation creep around them.  
180 In contrast and depending on their thickness, slab interiors deform mostly through dif-  
181 fusion creep (Fig. 2b-c) because of their much colder state (Fig. 1b and d). The evolu-  
182 tion of individual slabs is significantly affected by composite rheology, consistent with  
183 regional thermo-mechanical models (e.g. Garel et al., 2020): slabs tend to sink faster through  
184 an upper mantle with more abundant dislocation creep and thus a more pronounced low  
185 viscosity zone (Fig. 2). Moreover, they tend to buckle and/or break-off more easily de-  
186 pending on their strength and the mantle viscosity structure (Fig. 4a and d). In fact,  
187 both the amount of dislocation creep around slabs and the thickness of the asthenosphere  
188 are responsible for creating a viscosity contrast between the upper and the lower man-  
189 tle, which hinders the sinking of slabs and affects their evolution (Fig. 2a, e.g. Billen and  
190 Hirth (2007)).

191 Around plumes, hot mantle more likely deforms through dislocation creep (Fig. 1b  
192 and d), although shearing is less important than around slabs. Plumes are thus also sur-  
193 rounded by lower viscosities than pure diffusion creep cases, which favors fast rising (Fig.  
194 2 and Fig. S4). Plume material further tends to feed fast lateral asthenospheric channeled-  
195 flow (as proposed by e.g. Phipps Morgan et al., 1995) in which dislocation creep occurs  
196 more likely due to high temperatures and stresses, favoring even lower viscosity in these  
197 areas than in diffusion creep models (Fig. 2a and b). This occurs preferentially when new  
198 plume heads reach sub-lithospheric depths. Over a few million to a few tens of million



**Figure 2.** (a) Time-averaged profiles of (left) mantle fractional area with >10% dislocation creep, (middle) minimum and mean viscosity, and (right) vertical and horizontal velocity for models with a surface yield stress  $\sigma_{Y0} = 47$  MPa. (b-d) Proportion of dislocation creep and mantle velocity field (arrows scaled and coloured by magnitude) in three models. In (b), a 50 Myr-evolution is shown. In (b-c), blue lines show slabs and magenta lines contour plumes. In (d), purple lines contour dripping lithosphere and orange lines show hotter-than-average upwellings.

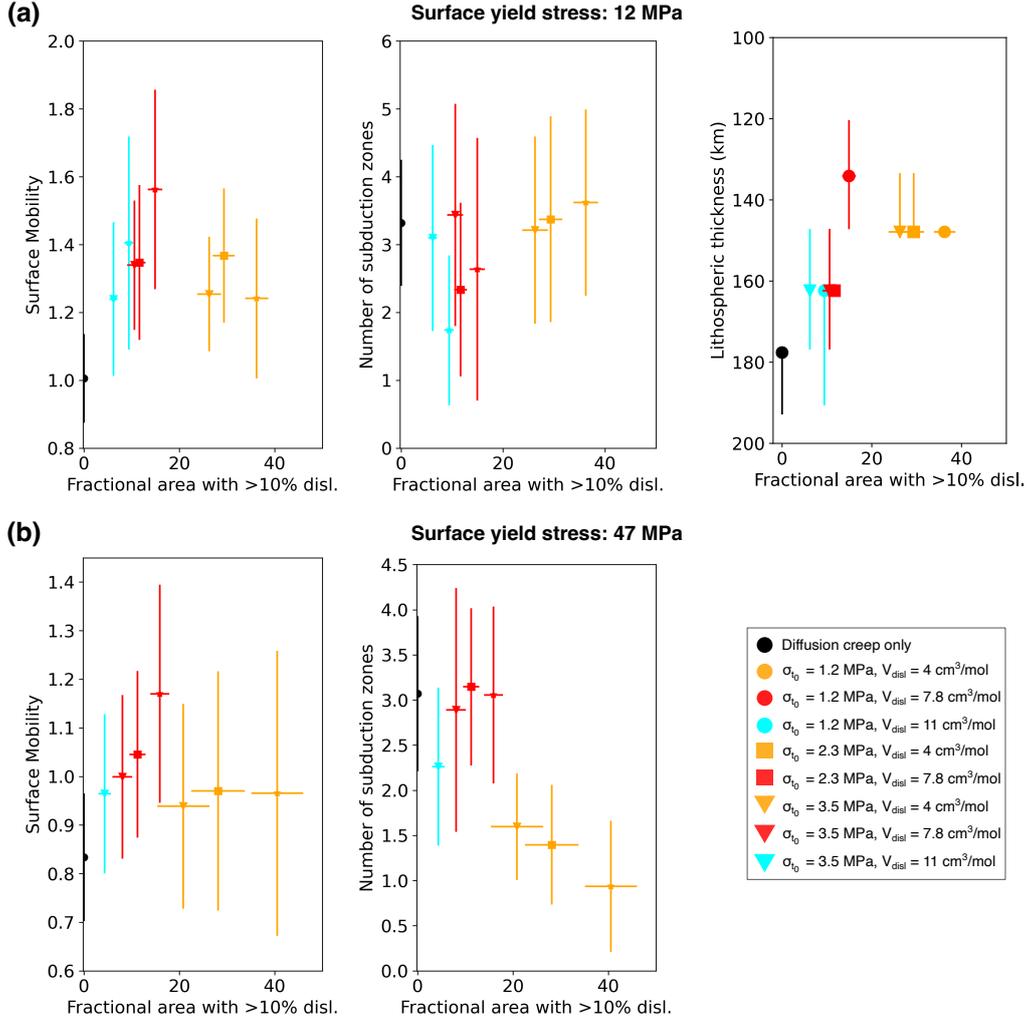
199 years, the geometry and abundance of dislocation creep can therefore vary considerably  
 200 (Fig. 2b and c), controlled by the dynamics of convective thermal heterogeneities.

201 Models with surface yield strength larger than 120 MPa experience stagnant-lid  
 202 convection. In these models, the mantle is much warmer due to limited heat loss, thus  
 203 favoring more vigorous and smaller-scale convection than in cases with plate-like behav-  
 204 ior (Fig. 2). Higher temperature and increased convective vigor promote dislocation creep,  
 205 which emerges in areas of basal lithosphere dripping, or around hotter-than-average up-  
 206 wellings in the shallow mantle (Fig. 2c). The large variability of these processes controls  
 207 the spatio-temporal distribution of dislocation creep.

### 208 **3.2 Effects on the tectonic regime**

209 The effect of composite rheology on the surface tectonic regime is quantified through  
 210 surface mobility  $M = \frac{v_{surf}}{v_{rms}}$  (with  $v_{surf}$  the average surface velocity and  $v_{rms}$  the vol-  
 211 ume root-mean-square velocity) and plateness  $P = 1 - \frac{def_{90}}{def_{90,iso}}$  (with  $def_{90}$  being the  
 212 fractional surface area containing 90% of deformation, and  $def_{90,iso}$  being the value for  
 213 an isoviscous model, Tackley, 2000a). These proxies are close to 1 for the mobile-lid regime  
 214 and tend to 0 in the stagnant-lid regime, with episodic transitioning between these end-  
 215 members. In addition, we track the number of active subduction zones, detected from  
 216 surface downward velocity peaks, and the lithospheric thickness, defined from the inflec-  
 217 tion point of the time-averaged temperature profile (Fig. 3 and Fig. S5).

218 Regardless of the surface yield strength, lithosphere thickness decreases as the pro-  
 219 portion of dislocation creep increases (Fig. 3a and Fig. S5), by up to 60% compared to  
 220 diffusion creep models. In the asthenospheric areas strongly affected by dislocation creep,  
 221 increased convective vigor tends to impede lithospheric growth due to more efficient con-  
 222 vective erosion. Therefore, the thicker the layer with substantial dislocation creep, the  
 223 thinner is the lithosphere for a given surface yield stress compared to pure diffusion-creep  
 224 models. Besides the major control of surface lithospheric yield strength, composite rhe-  
 225 ology has two contrasting effects on the tectonic regime. These effects are summarized  
 226 on Fig. 4a-c and described below.



**Figure 3.** Effect of composite rheology on surface tectonic regime (temporal average and standard deviation of surface mobility, number of subduction zones and lithosphere thickness as a function of the time-averaged mantle fractional area containing >10% dislocation creep) in models with  $\sigma_{Y_0} = 12$  MPa (a) and 47 MPa (b).

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### 3.2.1 Models with a weak lithosphere (<35 MPa)

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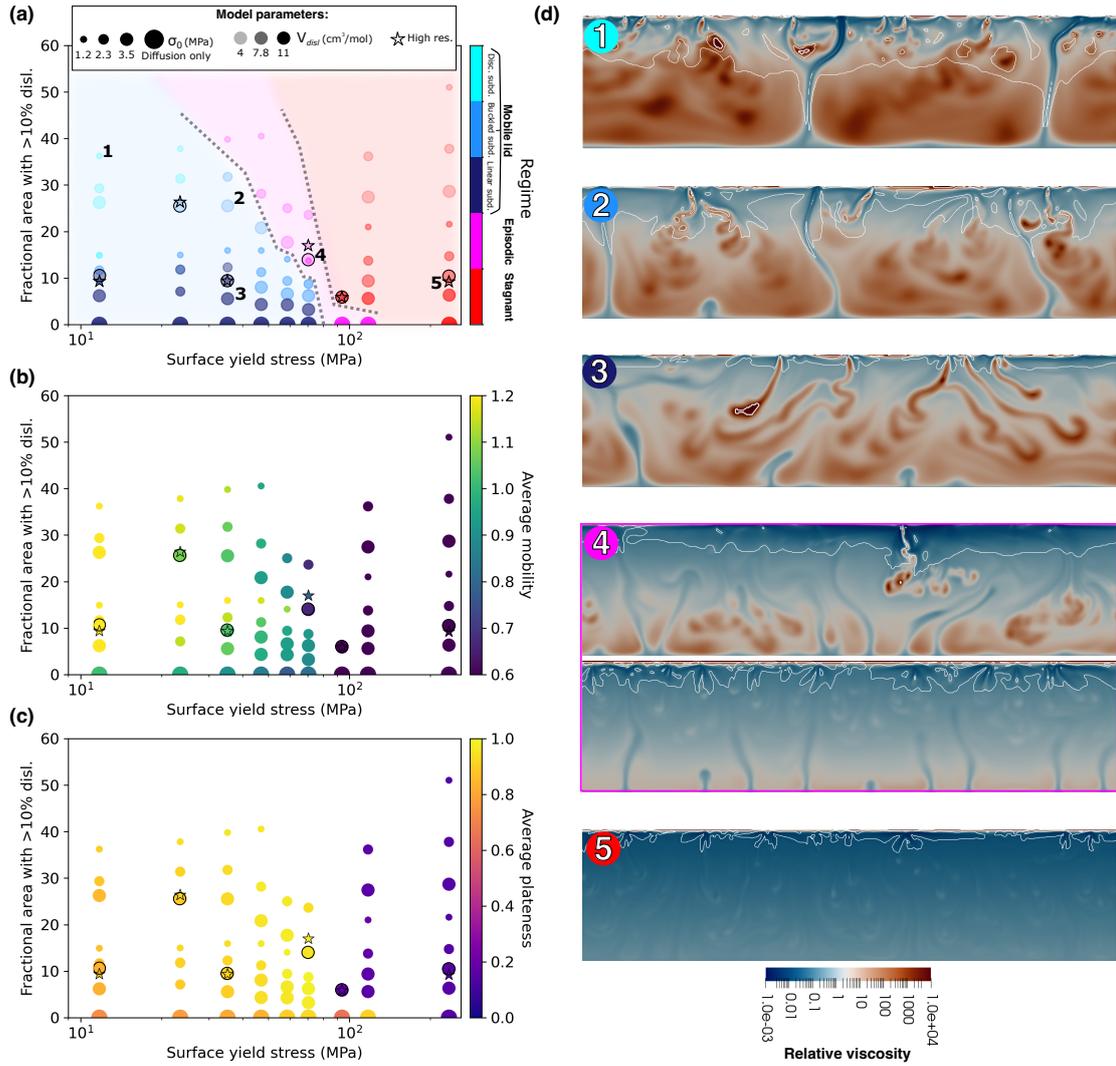
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For yield stresses below  $\sim 35$  MPa, models in pure diffusion creep are in the mobile-lid regime. Composite rheology enhances surface mobility (up to 1.6) and plateness. Active subduction zones tend to be shorter-lived (Fig. 4d). In these models, the viscosity reduction in the uppermost mantle induced by dislocation creep leads to the decoupling of lithosphere from the asthenosphere via lubrication, and to reduced stress acting on the lithosphere although local convective vigor increases (Tackley, 2000b). This decou-



**Figure 4.** (a) Regime diagram of all models. Mobile-lid models have discontinuous and short-lived subductions (cyan), buckled slabs (blue), or mostly linear slabs (deep-blue). Episodic models (magenta) have intermediate plateness and mobility. Stagnant-lid models (red) have low plateness and mobility. Qualitative boundaries are drawn between each regime. (b-c) Similar to (a) but with colours representing time-averaged surface mobility and plateness, respectively. (d) Snapshots of viscosity of selected models referred as numbers in (a). White lines contour low-viscosity regions with >10% dislocation creep.

234 pling contributes to the observed increase in mobility. Since dislocation creep also fa-  
 235vors lithosphere thinning, the plastic strength at lithospheric base is reduced compared  
 236to models in pure diffusion creep. Therefore, an increasing amount of dislocation creep

237 enhances thin slab break-offs. Accounting for composite rheology in models with a low  
 238 lithospheric strength thus enhances mobile-lid convection.

### 239 **3.2.2 Models with a strong lithosphere ( $>35$ MPa)**

240 Models in pure diffusion creep with surface yield stresses comprised between  $\sim 35$  MPa  
 241 and  $\sim 120$  MPa are also still in the mobile-lid regime. Including composite rheology with  
 242 decreasing values of  $V_{disl}$  and/or  $\sigma_0$  results in up to 40 % of the mantle being affected  
 243 by dislocation creep (Fig. 3b and 4a-c).

244 For small amounts of dislocation creep in the mantle ( $<20\%$ ), both plateness and  
 245 surface mobility tend to increase by a factor of up to 1.4 and the number of slabs remains  
 246 stable (Fig. 3b). In these models, thin low-viscosity asthenospheric areas tend to lubri-  
 247 cate the base of the lithosphere, enhancing plate mobility and plateness (e.g. Tackley,  
 248 2000b).

249 When the proportion of dislocation creep exceeds 20%, the number of active sub-  
 250 ductions, plateness, and surface mobility decrease (Fig. 3b, 4a-c, and S5b). Lithosphere-  
 251 asthenosphere decoupling promotes episodic and stagnant-lid convection (Fig. 4). This  
 252 strengthening phenomenon due to large viscosity contrasts between the convecting man-  
 253 tle and the lithosphere has long been demonstrated using Newtonian rheology (e.g. Moresi  
 254 & Solomatov, 1995; Solomatov, 1995; Höink et al., 2012, although the latter study in-  
 255 voked a flow channelization effect as being responsible for stagnant-lid convection) and  
 256 non-Newtonian rheology in the asthenosphere (Semple & Lenardic, 2020, although they  
 257 did not employ temperature- and depth-dependent viscosity, in contrast to the present  
 258 study).

259 We further tested higher surface yield stresses ( $>120$  MPa), which led to contin-  
 260 uous stagnant-lid behavior irrespective of our choice of activation parameters. Like in  
 261 models with a lower yield stress, decreasing  $V_{disl}$  and/or  $\sigma_0$  produces a thickening of the  
 262 layer containing dislocation creep. Although the convective regime remains unchanged  
 263 in these models, changing the amount of dislocation creep can strongly decrease the vis-  
 264 cosity in the asthenosphere and decrease lithospheric thickness by up to 60%. These ef-  
 265 fects could have a large impact on the distribution of partial melting and the rates of  
 266 magmatism on stagnant-lid planets (e.g. Schulz et al., 2020; Tosi & Padovan, 2021). How-  
 267 ever, these models also suggest that once a stagnant-lid is established with a pure dif-

268 fusion creep rheology, adding composite rheology in the upper-mantle does not promote  
 269 the generation of more plate-like behavior.

## 270 4 Discussion and conclusion

### 271 4.1 Model assumptions

272 Model setup simplifications potentially alter mantle flow and therefore the spatio-  
 273 temporal diffusion/dislocation creep partitioning. Our models are limited to 2D-cartesian,  
 274 have a reference Rayleigh number  $\sim 10$  times lower than Earth's, lower lithospheric strengths  
 275 than inferred from laboratory experiments (e.g. Brace & Kohlstedt, 1980), and lower ac-  
 276 tivation parameters for olivine than those predicted by rock experiments (e.g. Hirth &  
 277 Kohlstedt, 2003). We do not consider multiple mantle and lithosphere compositions and  
 278 phases (e.g. King, 2016). We also only tested one initial thermal state for our models  
 279 with composite rheology although different initial conditions could lead to distinct regime  
 280 boundaries for diffusion-creep-only and composite rheology, as shown in e.g. Semple and  
 281 Lenardic (2021); Weller and Lenardic (2018).

282 However, our mobile-lid models still produce mantle velocities of the order of the  
 283 cm/yr (Fig. 2a), oceanic lithosphere thickness of 100-200 km, and successfully generate  
 284 dislocation creep where it is expected to occur from rock-deformation experiments (Fig.  
 285 1). We also note no significant difference when increasing the resolution (Fig. 4a-c, star  
 286 symbols). Therefore, we anticipate that the general convective and tectonic trends (Fig.  
 287 4) and physical mechanisms described in this study still apply using more Earth-like se-  
 288 tups. In particular, we obtain a self-generated and self-evolving low-viscosity astheno-  
 289 sphere without invoking water and/or partial melting (King, 2016; Semple & Lenardic,  
 290 2020), the latter being often called on to justify the use of weakening laws to improve  
 291 plateness in whole-mantle Newtonian models (e.g. Tackley, 2000a; Bello et al., 2015).

292 Importantly, we assumed a uniform static grain-size, although rock-deformation  
 293 experiments indicate that diffusion creep should strongly depend on grain-size evolution  
 294 (Eq. 1). In some stagnant-lid models, we increased the static grain size, which produced  
 295 an increase in mantle average viscosity, stress, and proportion of dislocation creep in the  
 296 uppermost mantle (Fig. S6), associated with lithospheric thickening, as already described  
 297 in Schulz et al. (2020). This test, applied to models without dynamic grain-growth and  
 298 reduction, reveals the competing effects of large grain size (which tends to increase man-

299 tle viscosity) and large amounts of dislocation creep (which tend to decrease it) on litho-  
300 sphere thickness, at least up to a doubling of static grain-size with our setup (Fig. S6).  
301 Further exploring the role of grain-size evolution in mobile-lid scenarios is therefore needed  
302 to further understand the role of composite rheology on mantle and lithosphere dynam-  
303 ics.

## 304 **4.2 Earth’s observations and composite rheology in the uppermost man-** 305 **tle**

306 On Earth, seismic anisotropy, through the generation of dislocation creep-induced  
307 LPO (e.g. Nicolas & Christensen, 1987), can provide complementary insight on the lat-  
308 eral variations of mantle rheological properties (e.g. Becker et al., 2008). Although 3D  
309 modeling is required to quantitatively compare the diffusion/dislocation creep partition-  
310 ing in our models with observed seismic anisotropy, our results already potentially ex-  
311 plain its observed orientation and strength variations (e.g. Debayle et al., 2005), as well  
312 as high strength around slabs (e.g. Jadamec & Billen, 2012) and in the thermal trail of  
313 plumes (e.g. Barruol et al., 2019). The correlation between strong anisotropy and fast  
314 plate velocities described in Debayle and Ricard (2013) could also partly result from the  
315 fact that these plates are attached to fast sinking slabs, thus favoring more dislocation  
316 creep due to lithosphere basal shear. One future direction would therefore be to estimate  
317 seismic anisotropy in more Earth-like models with composite rheology and compare it  
318 to Earth’s observations (e.g. Kendall et al., 2022). Together with the consideration that  
319 some rheological parameters are inter-dependent (e.g. Jain et al., 2019), this should pro-  
320 vide complementary constraints on the range of rheological parameters applicable to Earth.  
321 Finally, another independent constraint could come from the study of how composite rhe-  
322 ology affects the spatio-temporal distribution of surface dynamic topography, both on  
323 the long-term (e.g. Bodur & Rey, 2019) and on shorter glacial-isostatic-adjustment timescales  
324 (Kang et al., 2022).

325 In this study, we show that our choice of composite rheological parameters impacts  
326 uppermost mantle spatio-temporal viscosity variations and dynamics, therefore affect-  
327 ing convection and surface tectonics in a non-linear way: at low lithospheric strength,  
328 increasing the proportion of mantle deforming through dislocation creep promotes plate  
329 mobility as well as numerous, weaker and short-lived slabs. In contrast, increasing the  
330 proportion of mantle containing dislocation creep in models with large lithospheric strength,

331 results in episodic to stagnant-lid convection. This shows the potential geodynamical in-  
 332 fluence of experimental uncertainties of the rheological parameters and calls for both fur-  
 333 ther experimental refinement of mantle rheological parameters such as  $V_{disl}$ , and further  
 334 exploration of the effects of composite rheology on mantle convective planform and sur-  
 335 face tectonics in more sophisticated planetary-scale models.

## 336 Open Research

337 The convection code StagYY (Tackley, 2008) is property of ETH Zurich and Paul  
 338 J. Tackley. Data files used in this study can be downloaded from Arnould (2022).

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