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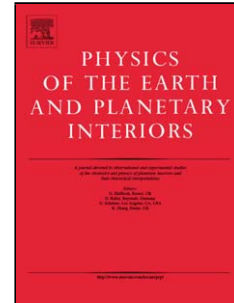


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1 **Morra et al. simulated the fate of subducting plates**
2 **interacting with the mid-lower mantle with variable**
3 **viscosity**

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29 *viscosity*

30 **Abstract**

31 In the last two decades it has been proposed several times that a non-monotonic
32 profile might fit the average lower mantle radial viscosity. Most proposed profiles
33 consist in a more or less broad viscosity hill in the middle of the mantle, at a depth
34 roughly between 1,200 km and 2,000 km. Also many tomographic models display
35 strong signals of the presence of “fast” material lying at mid mantle depths and a recent
36 spectral analysis of seismic tomography shows a very clear transition for degree up to
37 around 16 at a less than 1,500 km depth. Finally latest works, both theoretical and
38 experimental, on the high-to-low spin transition for periclase, have suggested that the
39 high-spin to low-spin transition of Fe⁺⁺ might lie at the heart of all these observations.
40 To verify the dynamical compatibility between possible mantle profile and observed
41 tomographic images and compare them with possible mineral physics scenarios, such as
42 the spin transition, we employ here a recently developed Fast Multipole-accelerated
43 Boundary Element Method (FMM-BEM), a numerical approach for solving the viscous
44 momentum equation in a global spherical setting, for simulating the interaction of an
45 individual slab with a mid-mantle smooth discontinuity in density and viscosity. We
46 have focused on the complexities induced to the behaviour of average and very large
47 plates O (2,000 km – 10,000 km), characteristic of the Farallon, Tethys and Pacific plate
48 subducting during the Cenozoic, demonstrating that the a mid mantle density and/or
49 viscosity discontinuity produces a strong alteration of the sinking velocity and an
50 intricate set of slab morphologies. We also employ the Kula-Farallon plate system
51 subducting at 60 Ma as a paradigmatic case, which reveals the best high-resolution
52 tomography models and clearly suggests an interaction with a strong and/or denser layer
53 in the mantle. Our 38 models show that a plate might or might not penetrate into the
54 lowest mantle and might stall in the mid lower mantle for long periods, depending on
55 the radial profiles of density and viscosity, within a realistic range (viscosity 1, 10 or
56 100 times more viscous of the rest of the mantle, and a change of differential density in
57 the range -2% to 2%), of a transitional layer of 200 km or 500 km. We conclude that a
58 layer with high viscosity or negative density would naturally trigger the observed
59 geodynamic snapshot. We finally propose a scenario in which the long time
60 accumulation of depleted slabs in the mid-mantle would give rise to a partially

61 chemically stratified mantle, starting from the less prominent high-spin to low-spin
62 contribution on the basis of mantle density and rheology.

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64 1. Introduction

65 The radial profile of the lower mantle viscosity is still largely unknown due to
66 incomplete knowledge of composition, grain size, volatiles content, and to the lack of
67 laboratory experimental data at such high pressure and temperatures and the difficulty
68 of performing exhaustive ab-initio models of diffusion and dislocation creep for
69 polycrystalline materials. However, geophysical inversion data based on geoid and
70 post-glacial rebound (Forte and Mitrovica, 2001; Mitrovica and Forte, 1997) has
71 suggested at the first order a peak of high viscosity in the middle of the lower mantle at
72 around 2000 km and in a range between 10 and 100 times than the one in the upper
73 mantle.

74 Few viscosity radial profiles for the lower mantle have been put forward. Among
75 them (Ricard and Wuming, 1991) proposed that a peak of viscosity in the mid mantle
76 might explain a set of geoid and topographical data. From geophysical considerations
77 (Wen and Anderson, 1997) imposed a chemical barrier at 1200 km. (Kellogg et al.,
78 1999) introduced the concept of a barrier at 1800 km depth based on geochemical and
79 radiogenic heat production arguments, combined with tomographical evidences. Later a
80 number of works came to similar conclusions, for example (Forte and Mitrovica, 2001)
81 who combined satellite and ground data, (Ito and Toriumi, 2007) have found a peak of
82 activation energy for vacancy diffusion at mid mantle pressures (both experimentally
83 and numerically from large-scale molecular dynamics simulations) and Peltier (personal
84 comm., 2008) looked at the GRACE dataset and found that the hypothesis of the
85 viscosity hill in the mid mantle is still compatible with the most up-to-data gravity data,
86 although this presents a controversy, as other studies, e.g., (Soldati et al., 2009)
87 indicated that lower mantle viscosity structure can hardly be constrained from gravity
88 data.

89 1.1 Tomographical Evidences

90 A clearer picture of how anomalous the mid-lower mantle appears from mantle
91 tomography. One of the better seismically resolved areas of the Earth's mantle is under
92 North America, where the Farallon slab is arguably the best-known subduction system
93 characterized by likely penetration into the lower mantle, e.g. (Grand, 1994). The set of

94 panels on the left of Figure 1 illustrates the paleo plate reconstructions for the period
95 from 80 Ma to 40 Ma (Müller et al., 2008) of the splitting of the wide oceanic plate with
96 over 8,000 km width subducting under North-America into two subplates called
97 Farallon and Kula. A recent high resolution tomography (Sigloch et al., 2008),
98 reproduced in figure 1, right panel, offers a very detailed image of the Farallon slab at
99 mid mantle depth. The most prominent features are the thickening of the slab at depth
100 between 1,200 km and 1,500 km and the segmentation into two plates, most probably
101 the signature of the division of the paleo Farallon-Kula plate, also confirmed by the
102 combination of paleo-tectonic reconstruction and global tomography, as in (Ren et al.,
103 2007).

104 Such anomalous mid-lower mantle behaviour is confirmed by the visual inspection
105 of a global tomography as the the one shown in Figure 2 where the isosurface
106 corresponding 0.8% heterogeneity, associated with the shear-velocity model of (Grand,
107 1994) is originally displayed with the Paraview software (Henderson, 2007). While
108 also here the intriguing pattern of the Farallon slab is visible, with its shape that largely
109 broadens at 1,500–2,000 km depths, other similar patterns appear in other regions of the
110 lower mantle. Finally the picture shown in figure 1 and 2 for the Farallon slab is shown
111 to be robust and independent on the tomographic models assumption through the
112 comparative analysis of five independent global shear velocity models in Figure 3:
113 tx2007 (Simmons, 2006), pri-s05 (Montelli et al., 2006), rmsl-s06 (Li et al., 2007),
114 saw642an (Panning and Romanowicz, 2006) and smeant (Becker and Boschi, 2002). All
115 such models convey very clearly the picture of a Farallon slab that does not cross
116 straightaway in the lower mantle, but instead flattens at depth of 1500–2000 km, as it
117 would encounter an obstacle.

118 The 3-D seismic structure of the lower mantle can be also analyzed globally, looking
119 at the spectral signal of all the slabs together. Detailed analysis of spherical harmonics
120 results as the one shown in the Figure 4 (modified from figure 1 of (Boschi et al., 2008))
121 illustrates an unexpected non-monotonic radial pattern. The logarithm of the ratio of
122 positive-to-negative shear-velocity spectra from model SMEAN (Boschi et al., 2008)
123 recently also observed in (Houser and Williams, 2009) elucidates, in fact, how the
124 positive (blue) fast anomalies dominate until the depth of ~1500km up to spherical
125 harmonic degree 15, while the situation is reversed below where negative (red) values

126 dominate at lower depths. This unquestionable tomographical observation is the premise
127 for searching the causes and consequences of the phenomenon.

128 *1.2 Mineral physics insights*

129 The dichotomy between the deepest 1000km and the rest of the lower mantle has
130 been already proposed to be of compositional nature (Van der Hilst and Kárason, 1999).
131 However this picture has been enriched by recent works that combine the interpretation
132 of seismic data with the knowledge of the elastic properties of silicates whose spin
133 varies at mid mantle conditions (da Silva et al., 2000). Such novel works depart from
134 the most recent experiments on ferropericlase during its spin transition (Marquardt et
135 al., 2009) and exploit the surprisingly transient and anomalous character change of the
136 bulk modulus during the transition, that largely softens. Because the observed pattern of
137 vertical gradient of bulk Velocity (dV_p/dz) is instead smooth and apparently adiabatic
138 (Cammarano et al., 2003; Matas et al., 2007), the only explanation of this phenomena is
139 a mid-lower mantle thermo-chemical transition at the same depth of the spin transition
140 in order to compensate for the bulk-modulus weakening (Cammarano et al., 2010).

141 If this hypothesis proves to be correct, this groundbreaking interpretation would then
142 demand a thorough investigation of its geodynamical consequences and above all, of the
143 conditions that might have produced it. Such a mid-mantle transition is clearly not as
144 sharp as the upper lower mantle one but it is believed to be smooth. The inversion of the
145 laboratory data indicates a thickness of several hundred kms (Marquardt et al., 2009),
146 which is less than the prior spin transition thickness i.e. up to 1000 km (da Silva et al.,
147 2000). The combination of seismic and mineralogical investigations strongly suggests
148 that the thermo-chemical should happen at depths around 1200–1600 km (Cammarano
149 et al., 2010; Marquardt et al., 2009).

150 All these indications are converging toward a scenario in which the viscosity profiles
151 in the lower mantle might be non-monotonic, which would be entirely different from
152 the canonical picture one gets from an Arrhenius type of activation energy, while it
153 would be better described by compositional differentiation. A fundamental point to
154 clear is the origin of the hill shape of the mantle viscosity arising from some inversion.
155 If the rising of viscosity can be associated with the differentiated composition, i.e. with
156 more viscous rocks at higher depth, the following lowering viscosity would be simply

157 due to the rising temperature, i.e. to the Arrhenius law again. In this sense a selection of
158 stronger rocks at higher depth is the natural outcome of a very long-term dynamics in
159 which the deep earth is naturally less mobile.

160 With the exception of the work by (Wen and Anderson, 1997), who postulated a
161 chemical layer at around 1200 km depth based on geoid inversion, this is probably the
162 first attempt to fit the observed mantle observable with an explicit compositional well
163 defined radial transition. The details of the generation of such a three-layered mantle
164 require careful study and are still speculative, however the strength of the tomographic
165 indications listed above are so neat that require models able to explain them. Yet there is
166 only one conclusive evidence of the existence of such dramatic alteration, i.e.
167 (Cammarano et al., 2010) and more independent studies are required to confirm this
168 observation. Finally definitive estimates of this new radial profile have not been
169 presented.

170 *1.3 Geodynamic scenario and our parameterization*

171 If the above mineral-physics and seismological interpretations are both correct, it
172 remains to be explained the geodynamic origin of the lower mantle transition and what
173 is its impact for rising plumes and sinking slabs. This work aims at understanding just
174 the fate of a slab in the lower mantle. The novel geodynamic scenario here suggested is
175 the one in which a sinking slab must cross three distinct mantle layers in its way to the
176 core: the upper mantle, the external-lower mantle (i.e. above 1500–2000 km) and the
177 interior-lower mantle (i.e. below 1500–2000 km). In particular, we propose that the
178 viscosity hill often suggested in the literature was indicating such a mid-lower mantle
179 viscosity transition.

180 We consider 4 fundamental parameters influencing the fate of the slab. Other factors
181 taken fixed in all models are the thickness of the plate, set to 200 km, which should
182 account for the broadening that the plate undergoes due to diffusion mechanisms and
183 maybe thickening in the early stages of subduction. The plate starts each model from an
184 initial straight shape, inclined of 20 degrees from the vertical, just above the hill.

185 The first examined parameter is the differential density between slabs and mantle.
186 Because the slab composition cannot change in its way to the core (except some
187 devolatilization phenomena, mostly in the upper mantle), and the spin transition

188 happens in the slab at shifted depths compared to the rest of the mantle, due to
189 lithosphere's cooler temperature, we consider the entire range of possible differential
190 densities, 2%, 0% and -2%, from the lower mantle during the transition. Before and
191 after the mid-mantle transition we assume instead unvaried 2% negative buoyancy for
192 the slab (Figure 5). Real values can be different for each real case, but our end-member
193 scenarios should encompass all possible events.

194 The second parameter is mantle viscosity. We investigate a non-homogeneous
195 mantle with a mid mantle peak viscosity 1, 10 and 100 times of the far field mantle, as
196 they are end members and are of lower or equal order of magnitude to the proposed
197 mantle viscosity hills (Figure 5). We do not vary the mantle-lithosphere viscosity ratio.
198 Such ratio has been amply studied in most geodynamic simulations of subduction that
199 give a value of about 100 (Capitanio et al., 2007) in the upper mantle. In the lower
200 mantle such a value is typically estimated as being much lower. We assume therefore a
201 slab with a viscosity ratio of 10 between slab and mantle. Interesting tests of how a
202 homogeneously viscous slab might evolve in the lower mantle have been proposed by
203 (Káráson, PhD Thesis) showing a complexity of shapes, also seismically detected and
204 interpreted (Ren et al., 2007).

205 In the technical implementation of the viscosity hill, as explained in the appendix,
206 only the high shear stresses due to the highly viscous mantle have been implemented.
207 Several reasons justify this assumption: a) the spin transition should take place at
208 different depths inside and outside the slab, due to the differential temperature between
209 slab and mantle; b) we follow the conceptual model that the viscosity rises in the mantle
210 due to a compositional variation, not directly due to the spin change (however, this is
211 not clear yet); c) we are interested in the fate of the slab in the particular context of their
212 ability to cross a mid mantle viscosity peak and reach the core-mantle boundary
213 contributing to the D'' composition or to a more or less thick differentiated mid mantle
214 layer, and this is due more to the mantle shear stresses applied to the slab than to the
215 slab strength itself. In other words, we are interested only in the case in which the
216 lithosphere remains a coherent entity, while we can only detect, and not directly model,
217 the case in which it becomes unstable, thus producing fragments that drop in the
218 underlying mantle.

219 The last two parameters are geometrical, i.e. the thickness of the transition, in which
220 two end-members cases are tested: a thin 200 km and a thick 500 km. and two different
221 slab width cases: very wide, 10,000 km and relatively narrow, 2,000 km, corresponding
222 respectively to the largest plates on Earth and to a typical subduction zone.

223 The total of 36 possible combinations gives origin to a very large set of possible
224 outcomes that we summarize in three categories: (1) Stokes flow, for slabs that directly
225 penetrate through the transition and do not feel the presence of mantle peak effects, (2)
226 transient, for slab that sensibly slow down at the mid mantle transition and then go
227 through it arriving at the lowest mantle and (3) stalling for slabs that remain entrapped
228 in the mid-mantle and never cross the transition.

229 **2. Numerical Simulations**

230 The simulation of the flow of creeping systems with large and sharp viscosity and
231 density variations presents formidable challenges, due to the extremely complex
232 evolving heterogeneities. We employ an indirect boundary integral formulation
233 representing the boundary between a strong slab a less viscous mantle. The resulting
234 formulation, based on bounded Stokes flow laws, is a set of Fredholm integral equations
235 of the second kind. While the methods based on boundary integral equations (Brebbia,
236 1978) reduce the effort necessary for representing complex geometries, their standard
237 formulation is computationally very expensive for large problem due to the construction
238 of dense matrices. In order to accelerate the solution, we replace the calculation of the
239 dense matrix with the implementation of an algorithmic “matrix multiplier operator”,
240 based on the multipole expansion of the integral terms (Greengard and Rokhlin, 1987)
241 as detailed in the appendix.

242 In our numerical scheme, the slab surface is discretised into linear triangular
243 elements. It has been shown that the linear system arising from its related equation set is
244 well-conditioned and dense (Zhu, 2006). The system is then solved employing the
245 GMRES Algorithm which exploits the potentialities of the FMM fast multipole
246 approach (Greengard and Rokhlin, 1987). Once the single and double layer integrals are
247 calculated, the position of each node is updated with a simple explicit forward step in
248 time. The heat or chemical diffusion equation is not solved in this work because it
249 would unnecessarily complicate our procedures, but it can be straightforwardly

250 implemented either with a MP-BEM solver or coupling the code with FD finite-
251 difference or FEM finite-elements methodologies, as it is in the implementation stage.

252 Standard implementations of the boundary element method allow only the study of
253 homogeneous flow as discussed in (Morra et al., 2007)(Morra et al., 2009) and
254 illustrated in the first part of the appendix. In practical terms this means that for each
255 element, the density difference and the viscosity ratio between slab and mantle are
256 constant. In this work, for the first time, we employ a non homogenous viscosity and
257 density boundary formulation, which allows us to model the interaction of the slab with
258 a non-homogeneous mantle, characterized by non-monotonous radial density and
259 viscosity profiles. Both requirements are exploited employing different approaches,
260 described in more detail in the technical Appendix:

261 a) the viscosity ratio is straightforwardly mapped on each boundary element
262 separately, depending on the local position of the boundary, but the implementation
263 requiring the calculation of the viscosity ratio between inner and outer slab requires a
264 careful implementation. The approximation here employed is explained in detail in the
265 appendix, formula (9), and has been tested with excellent results, giving only local
266 errors in proximity to the mantle-hill zone. In practice, it corresponds to a mean-field
267 approximation, neglecting the effects of the far field viscosity transition;

268 b) the non-homogeneous differential density requires the calculation of a one
269 dimensional radial finite difference implementation of the pressure, applied to the
270 boundary element formulation, replacing the component of the stress in the Stokeslet
271 component, as explained in recalculating the formula (10). This is done at each step,
272 with minimum computational time consumption, by integrating over the entire mantle
273 density profile.

274 **3. Model Results**

275 In this section we describe in detail the results of our experiments in numerical
276 modelling; we provide in the Discussion section a geophysical explanation of our
277 theoretical insights. Let's consider first the interaction of a small plate (2000 km x 2000
278 km) that encounters a purely 500 km thick viscosity hill, without considering any
279 density variation, i.e. only the case in which $\Delta\rho_{\text{hill}}$ is 2% higher in the slab compared to
280 the surrounding mantle. The transition from the 10x to the 100x viscosity hill model is

281 paradigmatic of the dramatic change due to the pure mantle shear stresses to the slab
282 sinking. Three steps whose time evolution for the two models are illustrated in Figure 6.
283 The slab in both cases crosses the viscosity hill, but the time required for going through
284 it is one order of magnitude higher for the 100x model. Furthermore the impact with the
285 high viscosity layer changes dramatically the morphology of the down welling, which
286 converts from a slab like structure into a cylindrical drop, growing as Rayleigh-Taylor
287 (RT) instability. We call the case shown above figure 6 as being Stokes sinking, while
288 the one on the lower mantle is defined as being transient.

289 In order to show the effect of a density variation superposed to a viscosity hill, the
290 following figure 7 illustrates the time sequences of two 10x viscosity models with a
291 small 2,000 km x 2,000 km plate testing 2 different smooth density variations $\Delta\rho_{\text{hill}} =$
292 $\rho_{\text{litho}} - \rho_{\text{mantle}}$ at the peak either 0% or -2% compared to surrounding mantle. The 0%
293 model, whose time evolution is illustrated in figure 7 (top panels), displays an
294 intermediate outcome between the two models shown in figure 6, with the final
295 morphology (after the hill is crossed) resembling partly such as slab and partly such as a
296 RT instability, falling in a ring shape. The lower panel of figure 7 shows that in this
297 simple configuration, the opposing buoyancy model ($\Delta\rho_{\text{hill}}=-2\%$) is able to halt the plate
298 at the viscosity transition. The two models here displayed are paradigmatic for a
299 transition from a transient to a stalling case.

300 The figure 8 illustrates the final stage of 6 different modeled slabs, focusing on the
301 role of differential density, summarizing their morphology. Only the mild viscosity hill
302 (10x) is considered as they show a larger variation compared to the 100x cases. Here
303 upper and lower panels represent the comparison between a small (2,000 km x 2,000
304 km, top panels) and a very wide plate (10,000 km x 2,000 km, bottom panels). Although
305 the timing of the deformation is very different for small and large plates (see figure 9),
306 the final configurations are similar. The most surprising result is that, while a variation
307 from 2% to 0% triggers little effect, the negative-buoyancy ($\Delta\rho_{\text{hill}}=-2\%$) model is
308 characterized by a rounded long-term shape of the plate, with down welling at the sides
309 of the plate itself. As it is shown in Figure 9, such a down welling arises a long time
310 after the plate has met the mid-mantle hill. It is therefore foreseeable that for a
311 morphological complex hill this instability might grow from segments inducing several
312 mesoscale down welling (Cizkova and Matyska, 2004), but possibly also being an

313 origin for upwelling plumes. In general, the models with inverse density hill (able to
 314 invert the sign of the slab buoyancy) display an unstable upside-down effect for which
 315 the slab splits and then fall laterally in form of drops.

316 All models are finally summarised in figure 9 where it is displayed the radial position
 317 of the center of mass of each plate versus a common non-dimensional time (a 2,000 km
 318 x 2,000 km plate in top panels and a 10,000 km x 2,000 km plate in the bottom panels).
 319 The trajectories of the centre of mass for the small and the 5x wider plate display a
 320 similar pattern but the mean-velocity is about double for the wider plate (see x-axis of
 321 the upper and lower panel). This can be explained by the prediction of (Capitanio et al.,
 322 2007) based on the classical work of (Happel and Brenner, 1983), that the drag of a
 323 plate in the mantle is proportional to $[1+\ln(L/S)]$ where L is the maximum length of the
 324 plate and S is the shortest. In our case $D_L=1+\ln(10,000 \text{ km} / 2,000 \text{ km})=1+\ln(5)=2.6$
 325 while $D_S=1+\ln(2,000 \text{ km} / 2,000 \text{ km})=1+\ln(1)=1$. The sinking velocity is given by the
 326 ratio between plate total buoyancy (volume* $\Delta\rho$) and drag, $v_L\sim(5x1)/2.6\sim 1.9$ and
 327 $v_S\sim(1x1)/1=1$, therefore the ratio $v_L/v_S=1.9$ explains the observed sinking velocity.

328 The non-dimensional time employed in figure 9 generates a range of possible sinking
 329 rates that depend on the physical parameters involved in the system. In order to detect
 330 the equivalent order of magnitude for the modelled plates, we can convert the non-
 331 dimensional time using the velocity renormalization factor $2/9*\Delta\rho g a^2/\eta$ and the time
 332 renormalization $9/2*\eta/\Delta\rho g a$, where η is the mantle viscosity, $\Delta\rho$ is the differential
 333 density, g is the gravity and a is the effective length of the slab $a=S*[1+\ln(L/T)]$, where
 334 S and L are as above, 2,000 km and 2,000–10,000 km. Assuming a lower mantle
 335 background viscosity in the range of $10^{22} - 3x10^{22}$ Pa s and an average differential
 336 density in the range 1-2% ($\sim 50 \text{ kg/m}^3$) for the sinking lithosphere, one finds that a time
 337 unit is equivalent to 2 to 6 Myrs, which can be linearly renormalized for a different
 338 lower mantle viscosity. This implies that we predict a crossing time of the lower mantle
 339 of the order of 20–60 Myrs for very wide plates, as Farallon, Tethys or Pacific, and of
 340 40–120 Myrs for small plates as single pacific slabs, both for the models that directly
 341 cross the mid mantle transition. Comparing this time with the diffusive timescale of a
 342 plate in the mantle, this could explain why only the largest plates have been detected in
 343 the deepest mantle.

344 For both, wide and narrow plates, the trajectories can be grouped in three families
345 (Figure 9 and table 1), defined roughly by the same parameter subset. The first family is
346 specified by the cluster of trajectories that cross the hill in an almost unperturbed form
347 and is called Stokes. For a hill of only 200 km, such models are defined by the ones that
348 cross a small or no viscosity transition hill and have a lithosphere-mantle density
349 difference positive or null, while for a hill thickness of 500 km already when the density
350 difference is null the trajectory is perturbed. The second family is composed by the
351 plates that remain permanently or for extremely long time stalled at the hill. We call this
352 dynamics Stalling. This involves all the models with a hill-thickness of 500km and
353 negative $\Delta\rho$ and high viscosity and negative $\Delta\rho$ (see table 2 for better visualizing the
354 transition). Collected into a third family there is then a large set of models that are
355 characterized by moderate anomalous viscosity and moderate density anomaly,
356 displaying a transient behaviour. They sensibly slow around the viscosity/density hill
357 for periods of the same order of magnitude of the ones that penetrate down to the
358 bottom of the lower mantle and then sink down, often displaying a new morphology.
359 Those models are called Transient in the tables. Although their definition is here
360 slightly arbitrary, the direct analysis of the models shows their actual existence. Their
361 importance, as discussed in the next section, resides in the possibility that the long
362 transient time spent around the mid mantle transition might act as a reservoir that in the
363 life of the Earth evolution would originally cause the compositional anomaly observed
364 in the lower mantle, as better discussed in the next session.

365 **4. Discussion and new scenarios for the Earth's mantle**

366 Convergent results taken from gravity data, and more recently mineral physics and
367 seismology, indicate that a non-monotonic profile of viscosity/density might better
368 describe the mid lower-mantle dynamics. In particular, it has been proposed that a high
369 viscosity layer several hundred km thick might exist at depths 1,200–1,600 km. More
370 recently a study that compares mineral physics and seismological data introduces
371 evidences for a compositionally distinct layering below $\sim 1,500$ km depth (Cammarano
372 et al., 2010).

373 We find little difference between the dynamics of a very wide plate and a small plate,
374 except for the sinking-velocity, which respects a logarithmic law (Capitanio et al., 2007)

375 (Morra et al., 2009). This terminal velocity diversity explains why only very large-scale
376 plates (Farallon, Tethys and Pacific), while smaller slabs would tend cross, although
377 being slowed down, by the non-monotonic radial pattern of the lower mantle viscosity.
378 This filtering effect could be enhanced by a mid mantle layering that induces a transient
379 dynamics (Figure 9), which induces small slabs to stall in the mid-mantle after a period
380 of several tens Myrs, diffusing heat and thereby dissipating away their differential
381 buoyancy.

382 We modelled numerically the effects of a wide but realistic range of thicknesses,
383 densities and viscosities for a non-homogeneous layer in the lower mantle and on the
384 lower-mantle subduction of a relatively small and a very wide plate. Depending on such
385 parameters, a variety of different behaviours is possible. The change in hill thickness
386 induces alone stalling if a sufficiently strong negative buoyancy (to the mantle) is
387 applied, while the transient behaviour is common to both 200 km and 500 km models
388 (table 1), changing only the timing of the transition. We conclude that even a mild
389 viscosity and density transition, if broad enough, can dramatically change the fate of the
390 lower mantle slabs.

391 An unexpected outcome of the model is the fate of slabs that during their sinking
392 encounter a combination of a high viscosity zone with a negative density anomaly. We
393 predict that they might stall a long time in the mid-mantle transition zone, furthermore
394 they might tear and/or fragment because of the opposite buoyancy forces encountered
395 and then amplified by the local high viscosity of the mantle. This would reduce the
396 wavelength of the seismic anomalies observed below in the deep mantle, which can
397 explain the observation of Figure 4.

398 Plotting the hill-mantle viscosity ratio vs. slab-hill differential density in table 2
399 shows the most distinct transition between the three dynamic domains that we show in
400 this work, with a banded configuration that illustrates how both the negative buoyancy
401 and the high viscosity ratio are able to create a transient and a stalling dynamics. This
402 somehow predictable result is here directly quantified, illustrating how a) if the
403 viscosity peak proposed before for the mid mantle hill is realistic (100x) then the slab
404 must either stall or have a long residential time in the mid mantle (see bottom line of the
405 table 2); b) if the compositional heterogeneity in the deep lower mantle is more dense

406 then the above one (as it is predictable) the dynamics of the lower portion of the lower
407 mantle will be strongly inhibited and the slabs will always display a transient or stalling
408 behaviour (see right column of the table 2); c) if the two effects act in a combined way,
409 with close to neutral buoyancy and a mild hill-mantle viscosity ratio, the thickness of
410 the transitional layer will play an important role controlling the slab behaviour and also
411 the slab width will be relevant because inducing a larger mantle flow.

412 If a density and viscosity non-homogenous layer of thickness of 200 km or 500 km
413 exists in the lower mantle, this must be necessary due to a compositional heterogeneity,
414 either within locally in the layer, or between a sandwich consisting of a top and a
415 bottom layer in the lower-mantle, because the spin transition in the ferropericlase alone
416 cannot induce it (Wentzcovitch et al., 2009). Furthermore, the bulk modulus softening
417 in ferropericlase during the spin transition is not observed in the dV_p/dz data, which
418 also implies that a compositional transition might compensate for the lack of dV_p/dz
419 jump. Such a configuration, if confirmed, requires a geodynamic system that at steady
420 state is able to create such compensation. We envisage here this new mechanism,
421 proposing that many sinking slabs would have transient time or completely stall in the
422 mid lower mantle. Such slabs represent a compositional differentiated reservoir, as it
423 has been for example already proposed for an hazburgite component in the D" layer
424 (Hirose et al., 1999). In the long term, a differentiation in the mid lower mantle would
425 dynamically develop by a feedback due to the already existing compositional lowest
426 mantle segregation, in a self-sustained form. This mechanism could induce over a long
427 term the viscosity-density hill itself, causing a radial separation in two layers of the
428 lower mantle, therefore preventing global mantle convection and favouring a more
429 layered form of mantle dynamics.

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437

438

439 **Appendix. Fast Multipole Boundary Element Method for Stokes Flow**

440 In order to capture the most fundamental modes that characterize how a slab can or
 441 not cross a density and/or a viscosity transition we simplified the system to its
 442 fundamentals. The effective viscosity $\eta(T,C)$ (Kellogg et al., 1999) that characterizes
 443 rheological behaviour of subducted plates has been replaced by two mean values: η_{int}
 444 inside the plate and η_{ext} for the external mantle. Using such rheological definition the
 445 gradient of velocity and $\boldsymbol{\sigma}$ is indicated by the viscous stress tensor:

$$446 \quad (1) \quad \boldsymbol{\sigma} = -pl + \eta (\nabla \mathbf{u} + \nabla^t \mathbf{u}) = -pl + \eta \dot{\boldsymbol{\epsilon}}$$

447 We solve the generalized Stokes equations that comprise the momentum
 448 conservation and incompressibility condition:

$$449 \quad (2) \quad \nabla \cdot \boldsymbol{\sigma} + \rho \mathbf{b} = 0 \quad \nabla \cdot \mathbf{u} = 0$$

450 For our problem, this amounts to subdividing the mantle into several closed regions,
 451 each characterized by a homogenous density and viscosity. When such an
 452 approximation is acceptable, it is possible rewrite (2) as a boundary integral equation for
 453 each sub-domain. This constitutes a decrease in the problem dimensionality and
 454 therefore potential computational gains. Moreover, this approach offers inherent
 455 multiscale capabilities as the surface mesh resolution can vary dynamically and track
 456 the physics of interest.

457 We show later how a system characterized by a perturbed viscosity η and/or density
 458 ρ in function of space:

$$459 \quad (3) \quad \eta = \eta'(z)$$

$$460 \quad (4) \quad \rho = \rho'(z)$$

461 as is our case, can be solved using a perturbative approach to our boundary
 462 equations.

463 *Boundary Equations*

464 We consider Stokes flow as described by the equations (1) and (2). Assuming a
 465 constant η in each domain, the velocity of each point inside the domain can be written
 466 as a sum of two surface integrals (Pozrikidis, 1992), called single and double layer
 467 integrals respectively, representing the effects of the forcing $\sigma_{ik}(\mathbf{x})\mathbf{n}_k$ and velocity $\mathbf{u}_i(\mathbf{x})$

$$468 \quad (5) \quad -\frac{1}{8\pi\eta} \int_{\partial D} \sigma_{ik}(\mathbf{x})n_k G_{ij}(\mathbf{x}, \mathbf{x}_o) dS(\mathbf{x}) + \frac{1}{8\pi} \int_{\partial D} u_i(\mathbf{x})n_k T_{ijk}(\mathbf{x}, \mathbf{x}_o) dS(\mathbf{x}) = \begin{cases} u_i(\mathbf{x}_o) & \text{if } \mathbf{x}_o \in D, \\ 0 & \text{otherwise} \end{cases}$$

469 where \mathbf{G}_{ij} \mathbf{T}_{ijk} are the steady Green's functions for velocity and stress respectively,
 470 also known as the Stokeslet and the Stresslet:

$$471 \quad (6) \quad G_{ij}(\mathbf{x} - \mathbf{x}_o) = \frac{\delta_{ij}}{r} + \frac{\hat{x}_i \hat{x}_j}{r^3}; \quad \hat{\mathbf{x}} = \mathbf{x} - \mathbf{x}_o \text{ and } r = |\hat{\mathbf{x}}|$$

$$472 \quad (7) \quad T_{ijk}(\mathbf{x} - \mathbf{x}_o) = -6 \frac{\hat{x}_i \hat{x}_j \hat{x}_k}{r^5}.$$

473 In turn, the above equation is implemented both for the inner and the outer fluid, in
 474 this way the boundary equations are cast into a form more appropriate for a quasi-steady
 475 multiphase flows in the presence of a gravity field. Hence for a point \mathbf{x} on the surface S
 476 that separates different fluids, we obtain:

$$477 \quad (8) \quad \frac{1 + \lambda}{2} \mathbf{u}(\mathbf{x}) - \frac{1 - \lambda}{8\pi} \int_S^{PV} \mathbf{n} \cdot \mathbf{T} \cdot \mathbf{u} dS = -\frac{1}{8\pi\eta_0} \int_S \mathbf{G} \cdot \Delta \mathbf{f} dS,$$

478
 479 where PV denotes the principal value of the integral, η_0 is the viscosity of the
 480 external fluid, taken as a reference and $\lambda = \eta_{\text{int}}/\eta_0$ is the viscosity ration between inner
 481 and outer fluid and $\Delta \mathbf{f}$ is a normal stress jump that in our case accounts for gravity.

482 The equations have been widely used and tested for homogenous media. In this work
 483 we considered two types of inhomogeneities parameterized into a spatial variation of the
 484 coefficients 1) viscosity ratio $\lambda = \eta_{\text{int}}/\eta_{\text{ext}}$; 2) density differential $\Delta \rho = \Delta \rho_{\text{int}} - \Delta \rho_{\text{ext}}$ We
 485 assumed that a zone of thickness between 200 km and 500 km, within a domain of
 486 several 1000's km, could have viscosities of 1-2 orders of magnitude higher than the
 487 surrounding and $\Delta \rho_0 = 0$ or even negative. In practice for each panel point \mathbf{x}_0 at the center
 488 of the Panel(\mathbf{x}_0), λ has been rescaled as:

489 (9)

$$490 \quad \lambda[Panel(\mathbf{x}_o)] = \begin{cases} \eta_{int}/\eta_{ext}(\mathbf{x}_o) = \eta_{int}/[\eta_0^{ext} + \eta'_{ext}(\mathbf{x}_o)] & \text{if } \mathbf{x}_o \in ViscosityHill, \\ \eta_{int}/\eta_{ext}(\mathbf{x}_o) = \eta_{int}/\eta_0^{ext} & \text{otherwise} \end{cases}$$

491 The implementation of non-homogeneous density is slightly more complex.
 492 Exploiting a first-order hydrostatic approximation the non-uniform density is
 493 implemented as a space dependent shift of the pressure term in the single layer integral
 494 (in our case, the space variable is the radius and the pressure is defined by a profile of
 495 the pressure of the mantle to the lithosphere). We recalculated the entire radial pressure
 496 profile at each time step and the external forces applied in the single layer integral to the
 497 system. Hence the pressure P at the radius $r=(\mathbf{x}_{ii}^2)^{1/2}$ becomes:

$$498 \quad (10) \quad P[Panel(\mathbf{x}_o)] = \int_{r(\mathbf{x}_o)}^{\infty} g\Delta\rho[r'(\mathbf{x})]dr'$$

499 that is applied in the single layer of the boundary equation, replacing $\Delta\mathbf{f}$ with \mathbf{Pn} .

500 *Multipole Approach*

501 The sinking slab surface S in Figure 5 and the supported quantities velocity \mathbf{u} , and
 502 stress tensor at the boundary $\boldsymbol{\sigma}$ are discretized with boundary elements (also called
 503 panels). The boundary integral equation thus becomes a linear system

$$504 \quad (11) \quad ((1 + \lambda)/2 + T) \mathbf{U} = \mathbf{F}.$$

505 Many approaches carry out the construction of the matrix; which scales as N_{pan}^2 both
 506 memory- and computation time-wise though, making it impractical for large systems.

507 We use a Fast Multipole Method (FMM) (Barnes and Hut., 1986; Greengard and
 508 Rokhlin, 1987) in eq (8). The FMM algorithm, illustrated in figure 10, dramatically
 509 reduces the complexity of matrix-vector multiplication involving a certain type of dense
 510 matrix, which can arise out of many physical systems.

511 The FMM scales as $N \log(N)$, which is far more tractable and still allows the use of a
 512 Generalized Minimized Residual method GMRES or any Krylov space based method
 513 that does not rely on the storage of the full matrix. By treating the interactions between
 514 asymptotic basis functions using the FMM, the corresponding matrix elements do not
 515 need to be explicitly stored, resulting in a significant reduction in required memory.

516 A multipole method exploits the decay of the kernel to convolve and makes a
 517 controlled approximation. It does this by expanding the system Green's function using a
 518 multipole expansion, which allows one to group sources that lie close together and treat
 519 them as if they are a single source.

520 More explicitly, let us compute

$$521 \quad (12) \quad u(\mathbf{x}_o) = \int_D G(\mathbf{x}_o - \mathbf{x})\rho(\mathbf{x})dV(\mathbf{x})$$

522 We consider the contribution from D_i , a part of D that is far enough from our
 523 evaluation point \mathbf{x}_o and proceed with a Taylor expansion of the kernel \mathbf{G} about $\mathbf{x}_c \in D_i$

$$\begin{aligned} 524 \quad u(\mathbf{x}_o) &= \int_{D_i} G(\mathbf{x}_o - \mathbf{x})\rho(\mathbf{x})dV(\mathbf{x}) \\ 525 &\simeq \int_{D_i} (G(\mathbf{x}_o - \mathbf{x}_c) - \nabla G(\mathbf{x}_o - \mathbf{x}_c) \cdot (\mathbf{x}_o - \mathbf{x}_c) + \dots) \rho(\mathbf{x})dV(\mathbf{x}) \\ 526 &\simeq G(\mathbf{x}_o - \mathbf{x}_c) \int_{D_i} \rho(\mathbf{x})dV(\mathbf{x}) \\ 527 &\quad - \nabla G(\mathbf{x}_o - \mathbf{x}_c) \cdot \int_{D_i} (\mathbf{x}_o - \mathbf{x}_c)\rho(\mathbf{x})dV(\mathbf{x}) + \dots \end{aligned}$$

529

530 We note that the equation involves successive moments of the ρ distribution in D_i .

531 The present FMM code can handle convolutions with the Green's functions for the
 532 Poisson equation, the Stokeslet or the Stresslet. It employs up to the second order
 533 moments of the source distributions (quadrupoles). The reader is referred to (Barnes and
 534 Hut., 1986; Greengard and Rokhlin, 1987; Warren and Salmon, 1993) for general
 535 information on multipole methods and the work of (Tornberg and Greengard, 2008) for
 536 the transposition of harmonic multipoles to the evaluation of the Stokeslet and Stresslet.

537 The FMM algorithm thus sorts the sources in a tree structure whose cells contain the
 538 moment integrals--or multipoles-- and carries out a field evaluation through a tree
 539 traversal (figure 11). The refinement of the interactions is determined by a tree traversal
 540 stopping criterion based on a prescribed tolerance.

541 The FMM-BEM drastically improves the computational cost of the method. For the
 542 coarse resolutions, the method displays the nominal N^2 scaling of a direct interaction

543 code. This is attributed to the relatively few elements and tree cells. The scaling then
544 quickly approaches a nearly linear one $N\log(N)$ for the finer resolutions.

545 The FMM-BEM has been parallelized using MPI. The parallel efficiency has been
546 tested on a Opteron cluster with Quadrics connections. The scaling is very good up to
547 64 CPUS, still keeping 90% of efficiency. In its current implementation the FMM-BEM
548 uses a shared (not distributed) tree, thus reducing the communication load at the
549 expense of memory requirements.

550

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- 656
657
658

659 **Figures**

660 **Fig.1.** The left panel displays three snapshots of the evolution of the northern region
661 of the Farallon-Kula plate system between 78Ma and 42 Ma. The black ellipse indicates
662 the location of the spreading ridge. The right panel shows the most up-to-date high
663 resolution 3D slab tomography reproduced from (Sigloch et al., 2008). The [SG] line
664 traces the Farallon-Kula paleo-plate boundary. At 1500km depth the two plates
665 superpose, due to less space at depth (Morra et al., 2007). The broadening and flattening
666 of the slab in the mid lower mantle confirms the the lower resolution outlook of the
667 global tomography models displayed in figures 2 and 3.

668 **Fig.2.** 3D isosurface of the 0.8% anomaly to the radial profile for the global
669 tomographic model of (Grand et al., 1997). The three dimensional morphology of the
670 Farallon slab is well distinguishable in the top portion of the figure. It is complex,
671 flattening laterally and apparently reaching the core only in a very confined location.
672 There are also other prominent features of large and broad flat anomalies at mid mantle
673 depth.

674 **Fig.3.** A comparison of 5 global S-wave tomographic models around the Farallon slab
675 at 4 different depths. Models include tx2007 (Simmons, 2006), pri-s05 (Montelli et al.,
676 2006), rmsl-s06 (Li et al., 2007), and saw642an (Panning and Romanowicz, 2006). The
677 smean (Becker and Boschi, 2002) model is an average the main proposed models and
678 summarizes them. The cusp of the Farallon slab is a common feature of most
679 tomographic models. Also the broadening of the image (indicating slab flattening) at
680 depth between 1,500km and 2,000km is independent on the inversion approach
681 employed.

682 **Fig.4.** This figure is mostly reproduced from figure 1 of (Boschi et al., 2008). It is
683 repropoed here as the main evidence of a dramatic transition at mid-lower mantle
684 depths. It shows the logarithm of the ratio of positive-to-negative shear-velocity spectra
685 from model SMEAN of (Becker and Boschi, 2002), as a function of harmonic degree l
686 (horizontal axis) and depth from Moho to core-mantle boundary (vertical axis). (Boschi

687 et al., 2008) computed independent harmonic expansions of negative and positive
 688 velocity anomalies from SMEAN, and took the logarithm of the ratio of the resulting
 689 spectra. See also the discussion proposed by (Houser and Williams, 2009). Positive
 690 (blue) values correspond to dominance of fast anomalies at a given depth and harmonic
 691 degree; vice-versa for negative (red) values: at relatively low harmonic degrees, fast
 692 anomalies dominate the pattern of seismic heterogeneity in the mid mantle, but the
 693 situation is reversed in the lower mantle, with the transition at $\sim 1,500$ km depth.

694 **Fig.5.** Viscosity and density radial profiles for the lower mantle. Above are shown two
 695 models proposed in the past by (Ricard and Wuming, 1991) and (Forte and Mitrovica,
 696 2001), both displaying a non-monotonous viscosity profile in the lower mantle, with a
 697 maximum at mid mantle depth. The lower panel illustrates instead the viscosity and
 698 density profiles that we test in this work. Our lower mantle model exhibit a less
 699 prominent viscosity peak respect to the one proposed above, as we observe in our model
 700 that 100x has already a strong influence on the geodynamic evolution. The density
 701 profile shown in bottom right figure illustrates both the hill-mantle and litho-mantle
 702 density differential. Those values are end-member, corresponding to typical seismic
 703 fluctuations. We model all the nine combinations.

704 **Fig.6.** Model results for the subduction of a small plate (2,000km x 2,000km) that
 705 encounters a mid lower mantle discontinuity. The differential density between slab and
 706 mantle is kept constant, 2%, while the viscosity peaks are tested, 10x and 100x, for a
 707 peak width of 500km. The viscosity transition is able to stop the slabs for a certain time,
 708 and retard largely the penetration into the lowest mantle, but the penetration in the deep
 709 mantle is very different for the viscosity peak of 10x or 100x, being the former a slab
 710 avalanche where the slab shape of the plate is maintained, while the last one is a
 711 Rayleigh Taylor instability type. The 100x is a much slower process, producing much
 712 longer residential times in the mid lower mantle.

713 **Fig.7.** Evolution of a sinking slab through a viscosity peak with two different
 714 lithospheric-mantle differential densities $\Delta\rho$ for a not very large plate (2,000km) and
 715 very broad peak zone (500km). The viscosity hill is chosen 10x the background lower

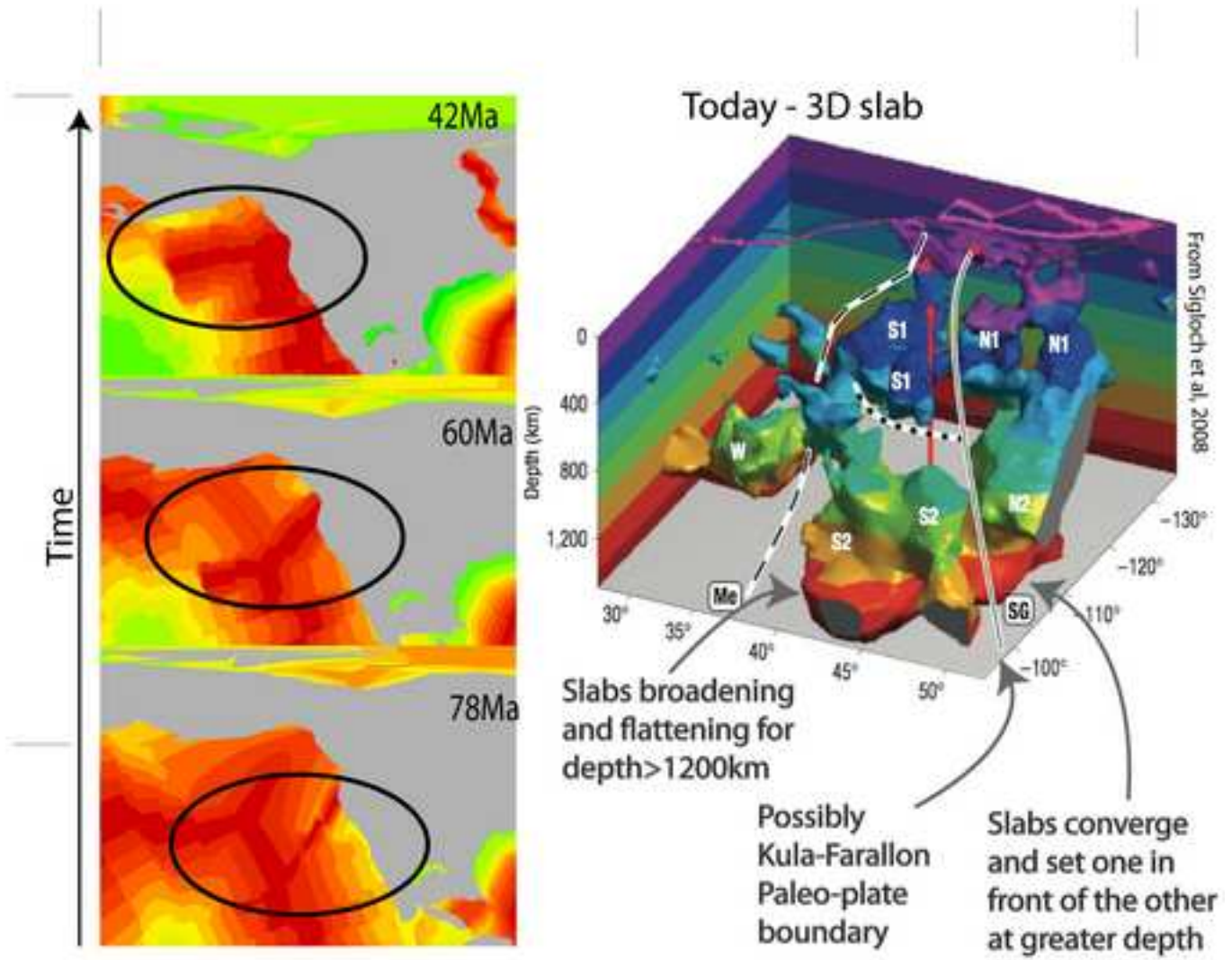
716 mantle viscosity. When the differential density between slab and hill density is null, the
 717 transition is crossed but as the slab morphology changes dramatically, generating a
 718 Rayleigh Taylor instability from the center of the stalling plate, with a ring shape. The
 719 time evolution shown in the lower panels illustrates how negative buoyancies (i.e. hill
 720 density higher than lithosphere density) inhibits the crossing inducing slab stalling. In
 721 the long term, if the differential buoyancy remains respect to the lowest mantle, how
 722 small scale or even a subduction-like instability triggered, as illustrated in Figure 8.

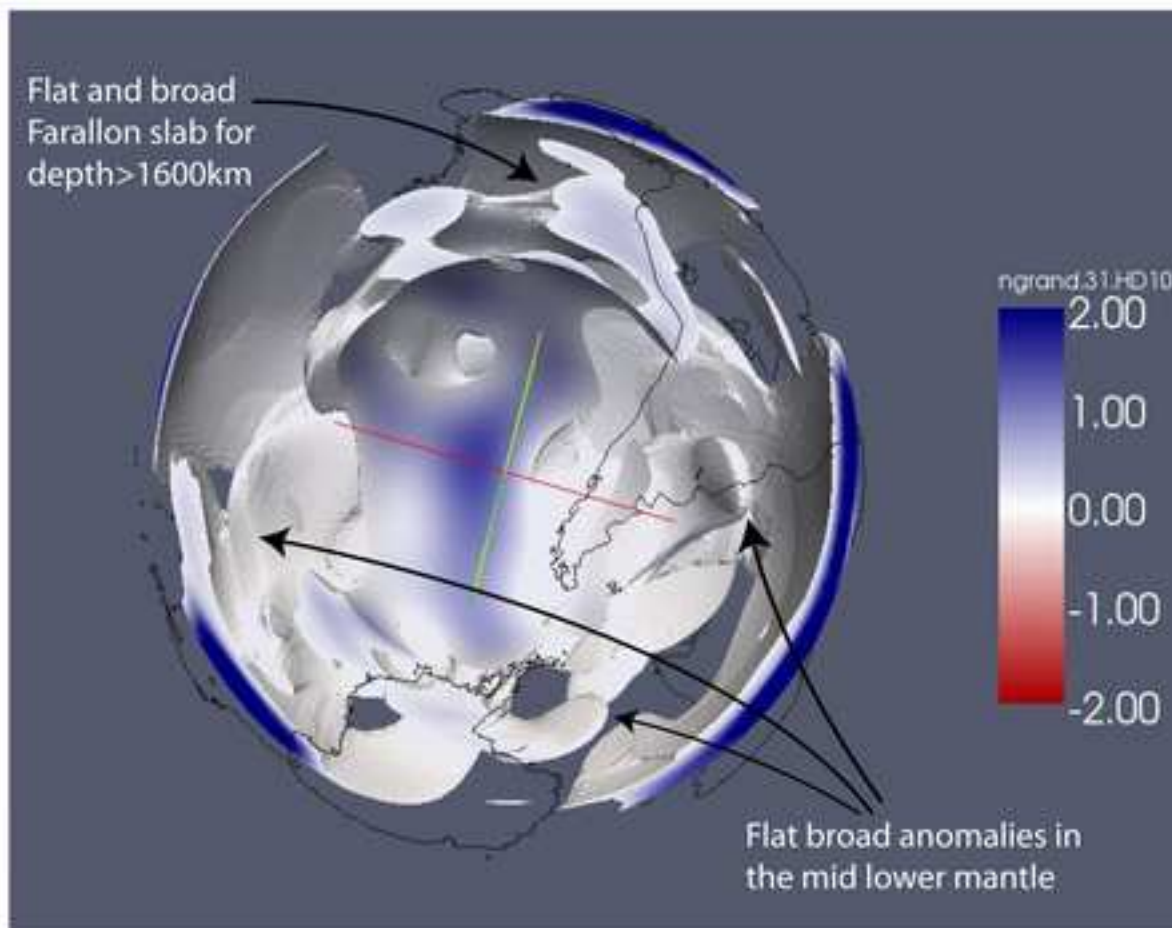
723 **Fig.8.** The final crossing morphology of 6 different models, 3 characterized by a small
 724 2,000km wide plate and 3 by a wide one (10,000km), all crossing a 200km thick
 725 transition zone and viscosity hill of 10x the background mantle. The models from left to
 726 right change through differential lithosphere-mantle density. The positive to null
 727 lithosphere-hill differential density comparison shows little effect, while the negative
 728 differential density (Hill Density > Slab Viscosity) instead triggers in the long term long
 729 residency times in the mid mantle and an unstable behaviour, coherently found for small
 730 or very wide plates, at different scales.

731 **Fig.9.** Summary of all models tested, through the radial trajectory of center of mass of
 732 the sinking slab. Small plates (2,000km wide) are displayed in the panel above, while
 733 very wide plate (10,000km) in the lower panel. The three families of models also
 734 recalled in the tables are (a) Stokes-like, which sink down to the core almost not
 735 interacting with the viscosity hill; (b) Stalling, with plates entrapped close to the
 736 viscosity-density hill for a period of the order of lower mantle Stokes crossing time or
 737 longer; (c) Transient, spend a moderate time around the viscosity hill, typically no
 738 longer of the lower mantle Stokes crossing time. Meaning of the legend: the line style
 739 indicates the thickness of the hill: dotted 500km (T=5) and continuous 200km (T=2).
 740 Colors instead display the hill peak viscosity ratio: blue 100 (V=100), green 10 (V=10)
 741 and red 1 (V=1). The dimensionless lithosphere-hill differential density is shown by
 742 symbols, which translated to standard differential density values becomes: circle +2%
 743 (D=1), triangle 0% (D=0) and rhombic -2% (D=-1). Externally to the transition the
 744 litho-mantle density difference (D=1) and the viscosity ratio (V=1) are kept constant.

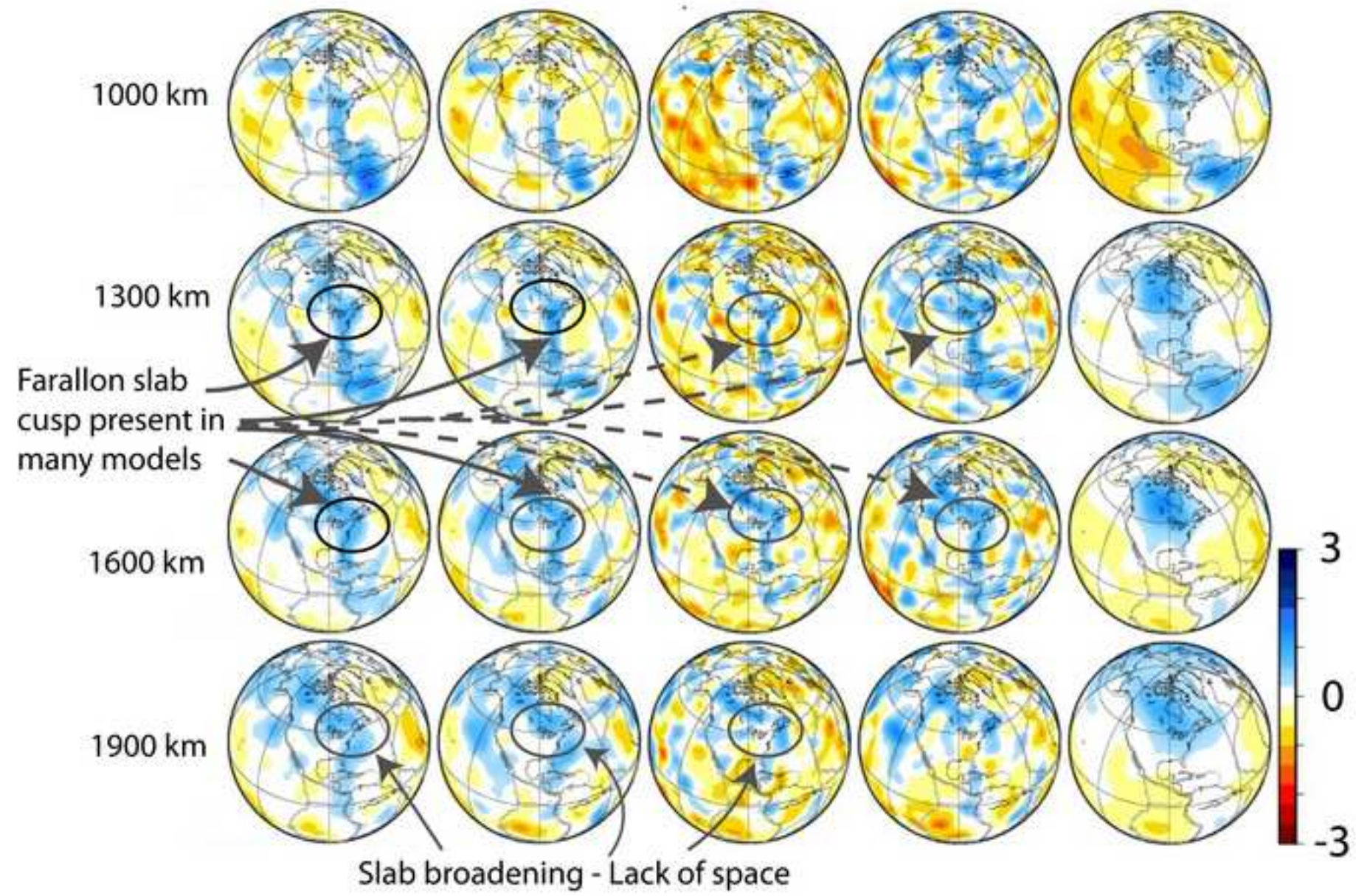
745 **Fig.10.** A general sketch of the multipole method employed. The summation on each
746 surface is done using local (Point) or Multipole terms, depending on the relative
747 position of different centers. Right Panel: The component of the summation are the
748 integrals on each panel (=boundary element). Summing the integral contribution within
749 a certain domain, the contribution of a set of panels is compressed into the one of one
750 pole. Left Panel: There are three possible summation methods. Local(=Point)-
751 Multipole, Local-Local, Multipole-Multipole. They depend on the relative contribution
752 of the green function.

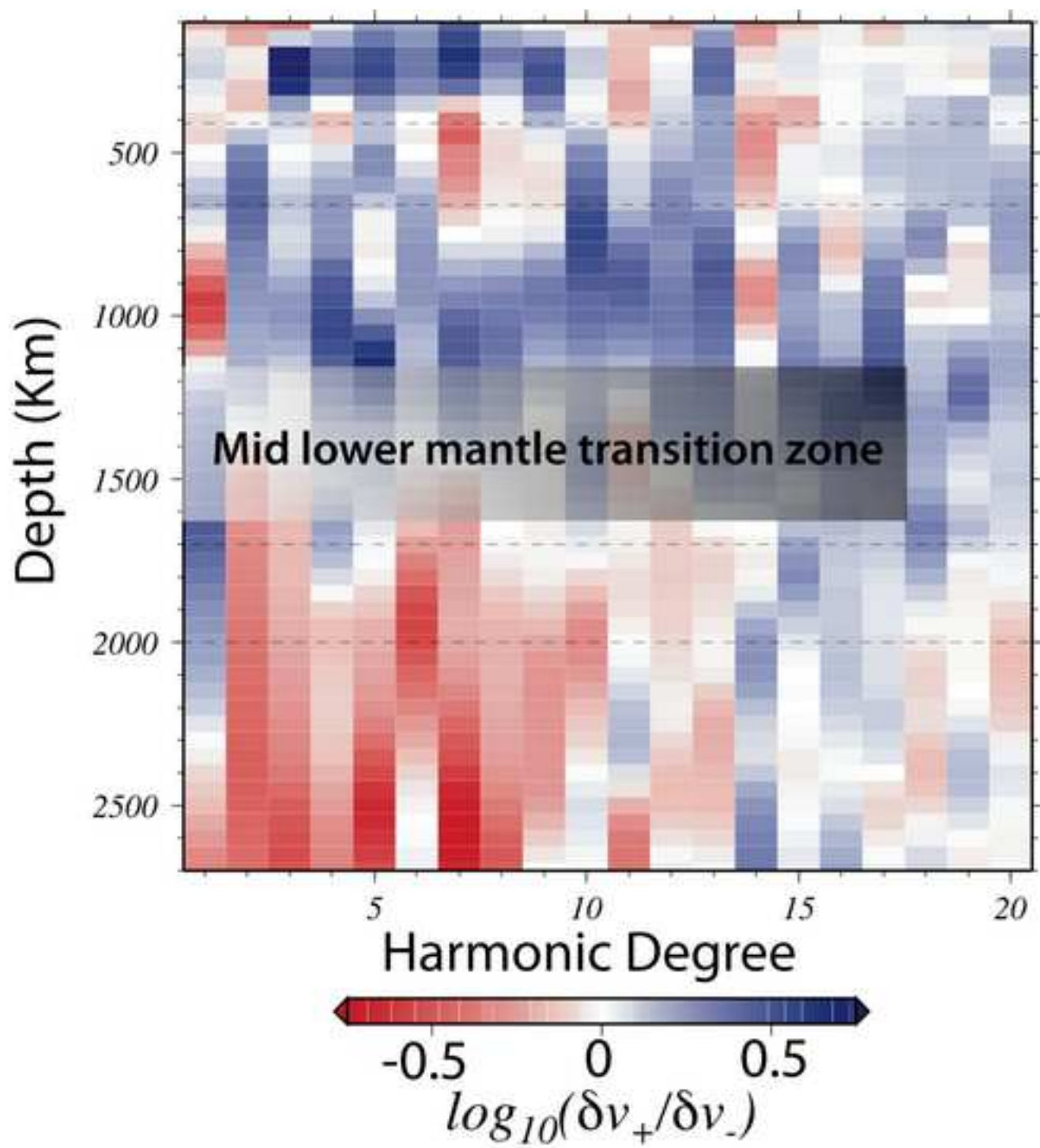
753 **Fig.11.** Sketch of a oct-tree and the hierarchical organization of its poles. Each internal
754 node has up to eight children in order to partition the space by recursively subdividing it
755 into eight octants until the required resolution is reached.

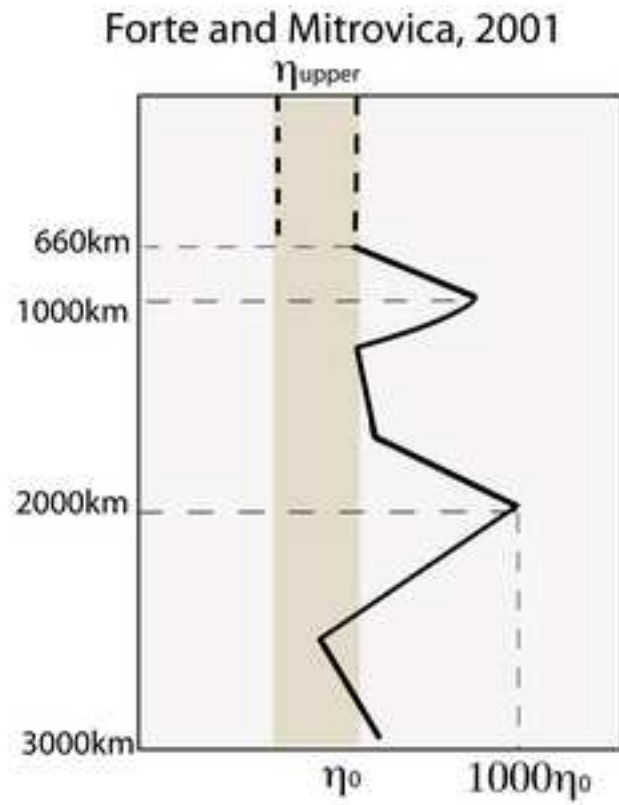
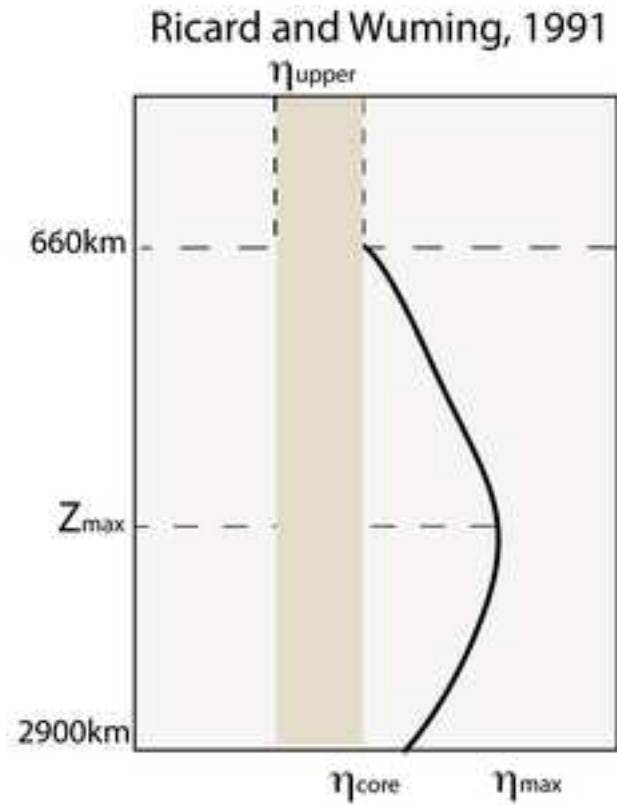




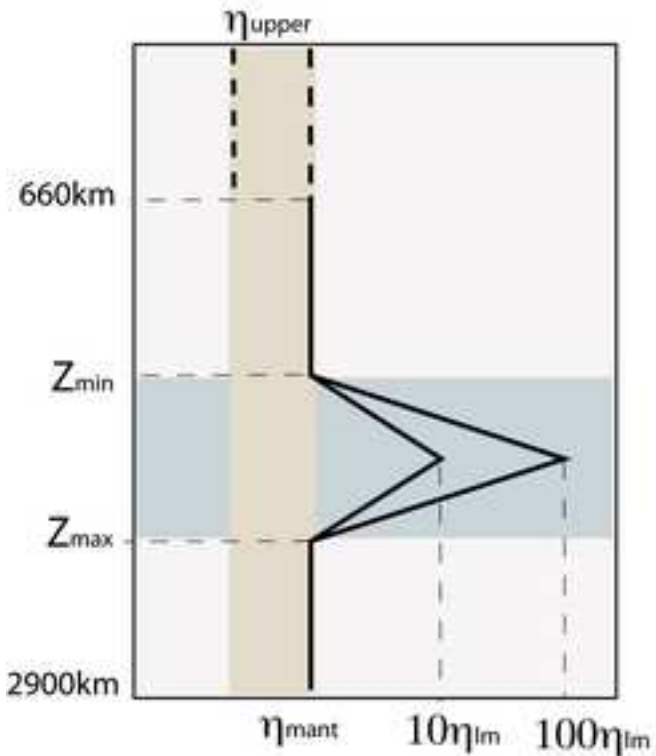
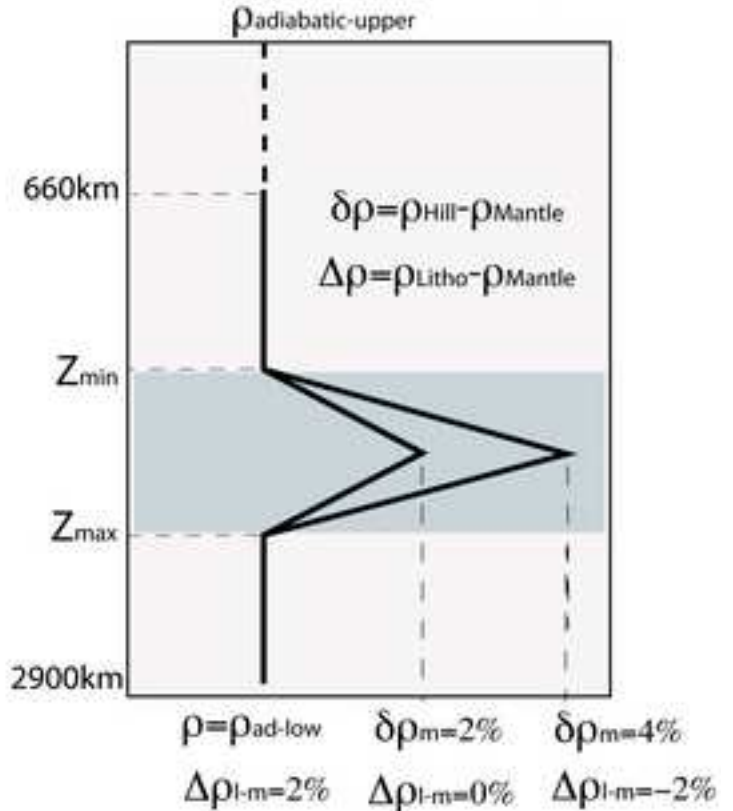
Farallon Slab from global tomographic models



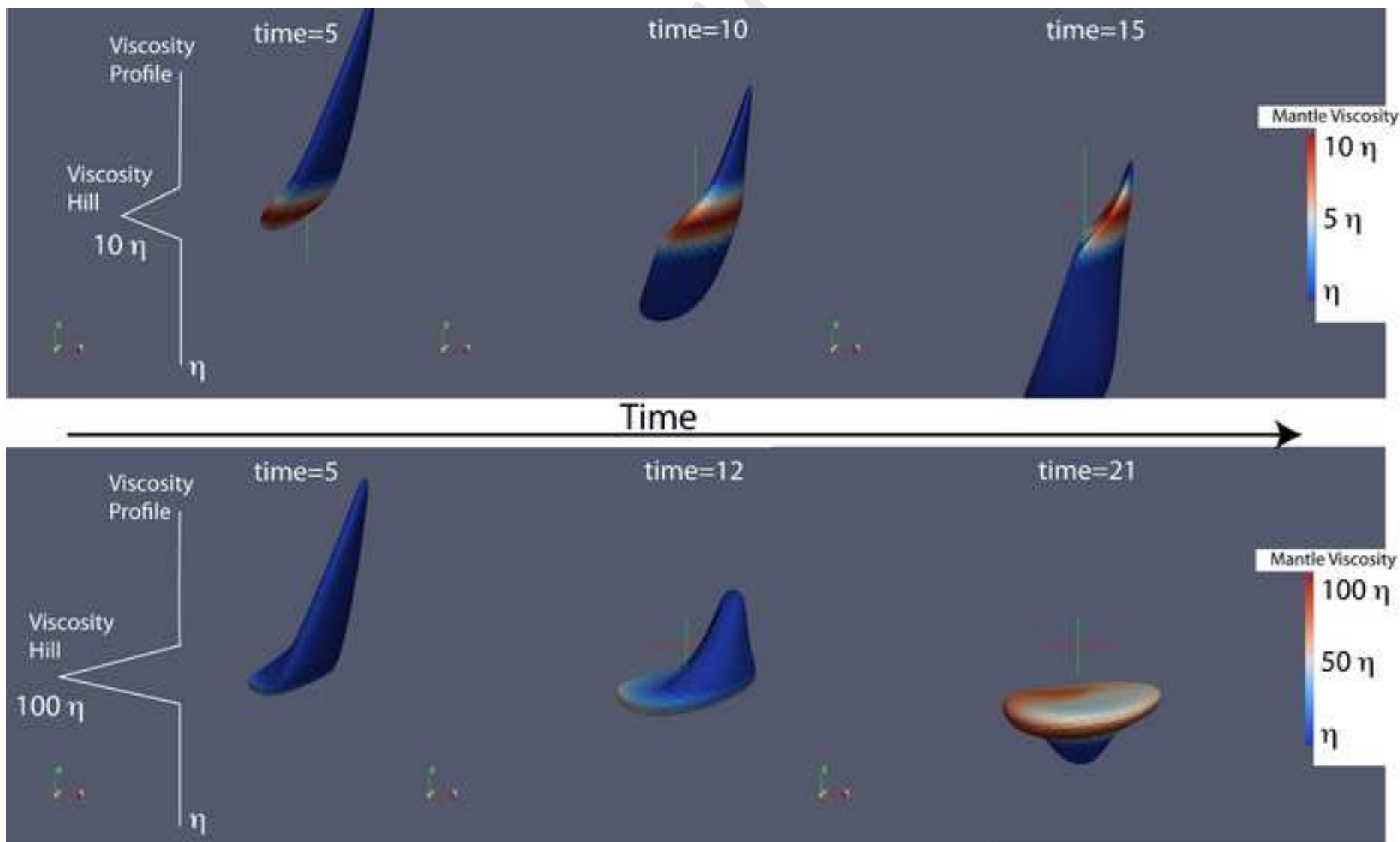


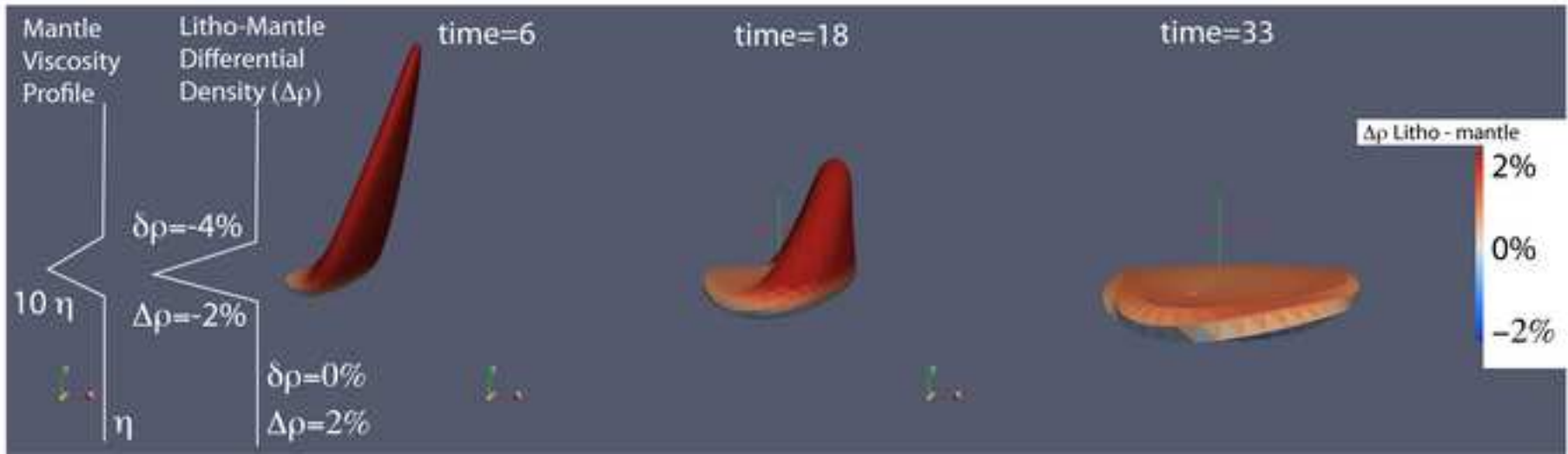
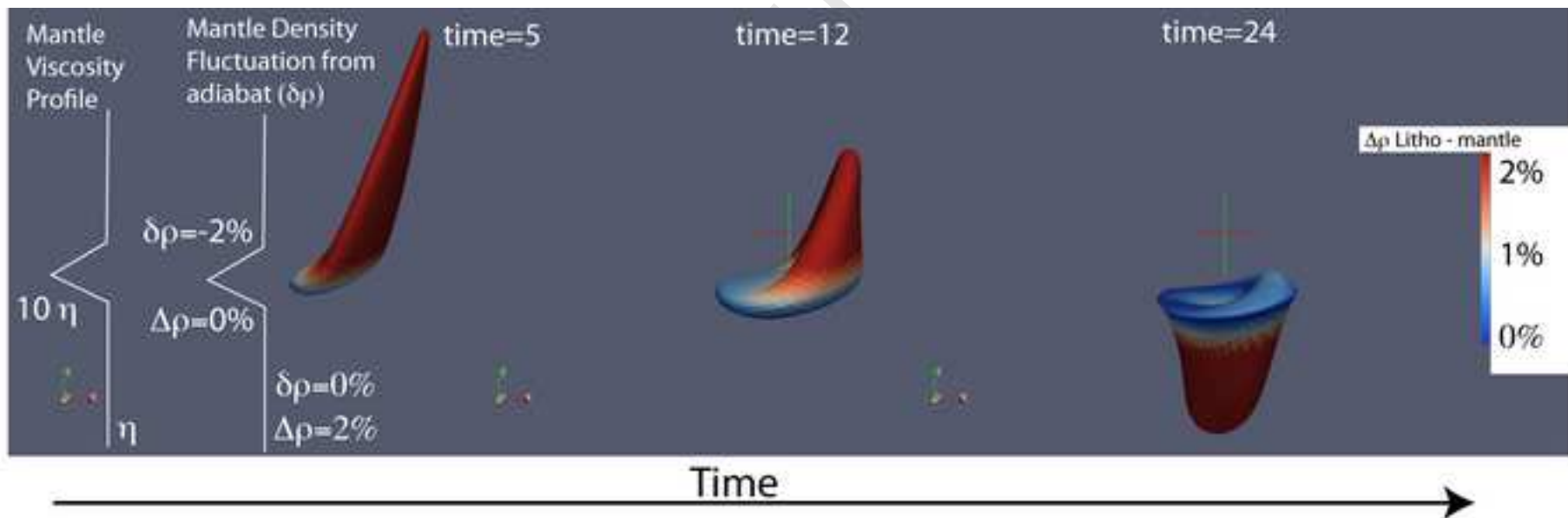


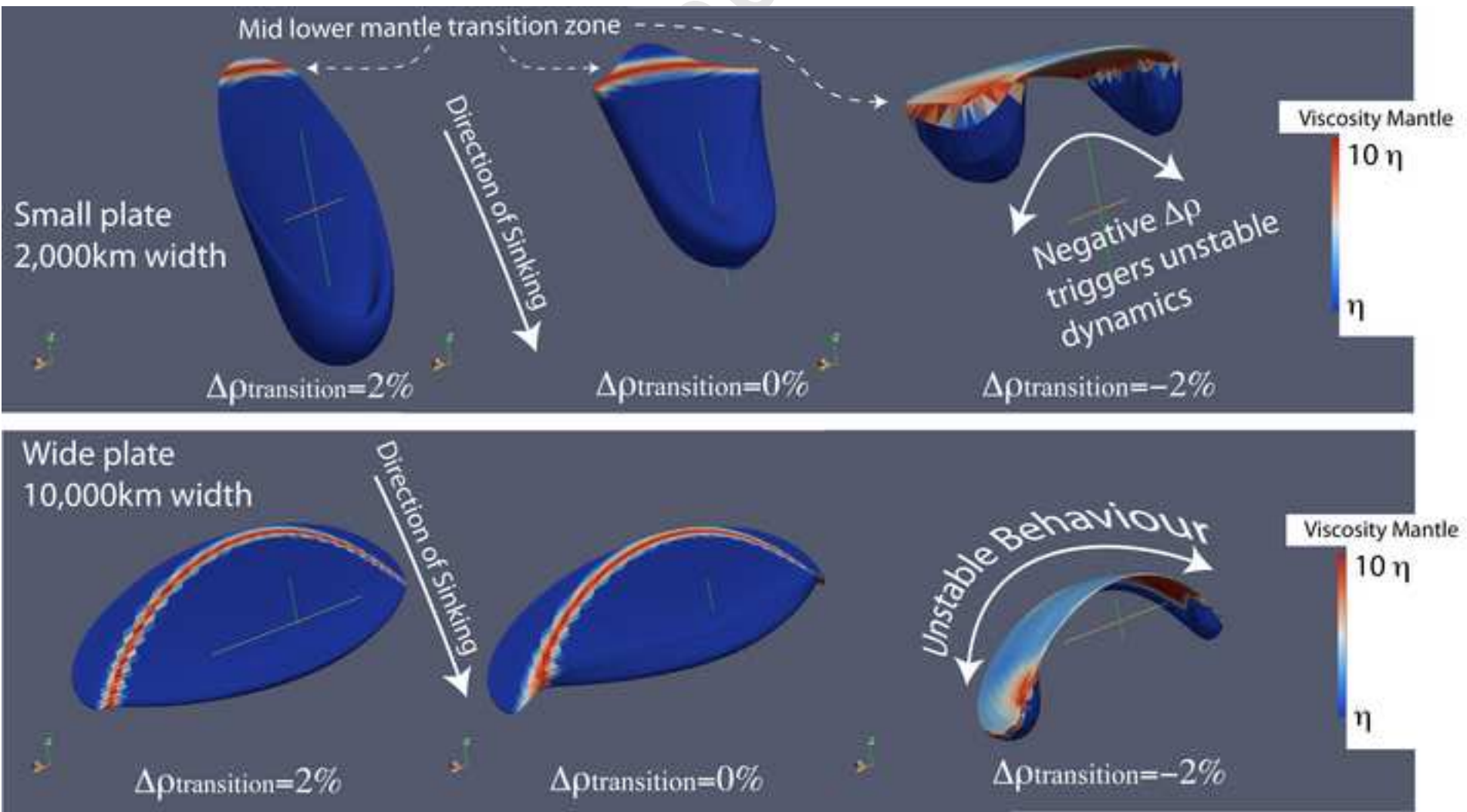
Our lower mantle Viscosity models

Our LM Density-fluctuation ($\delta\rho$) model

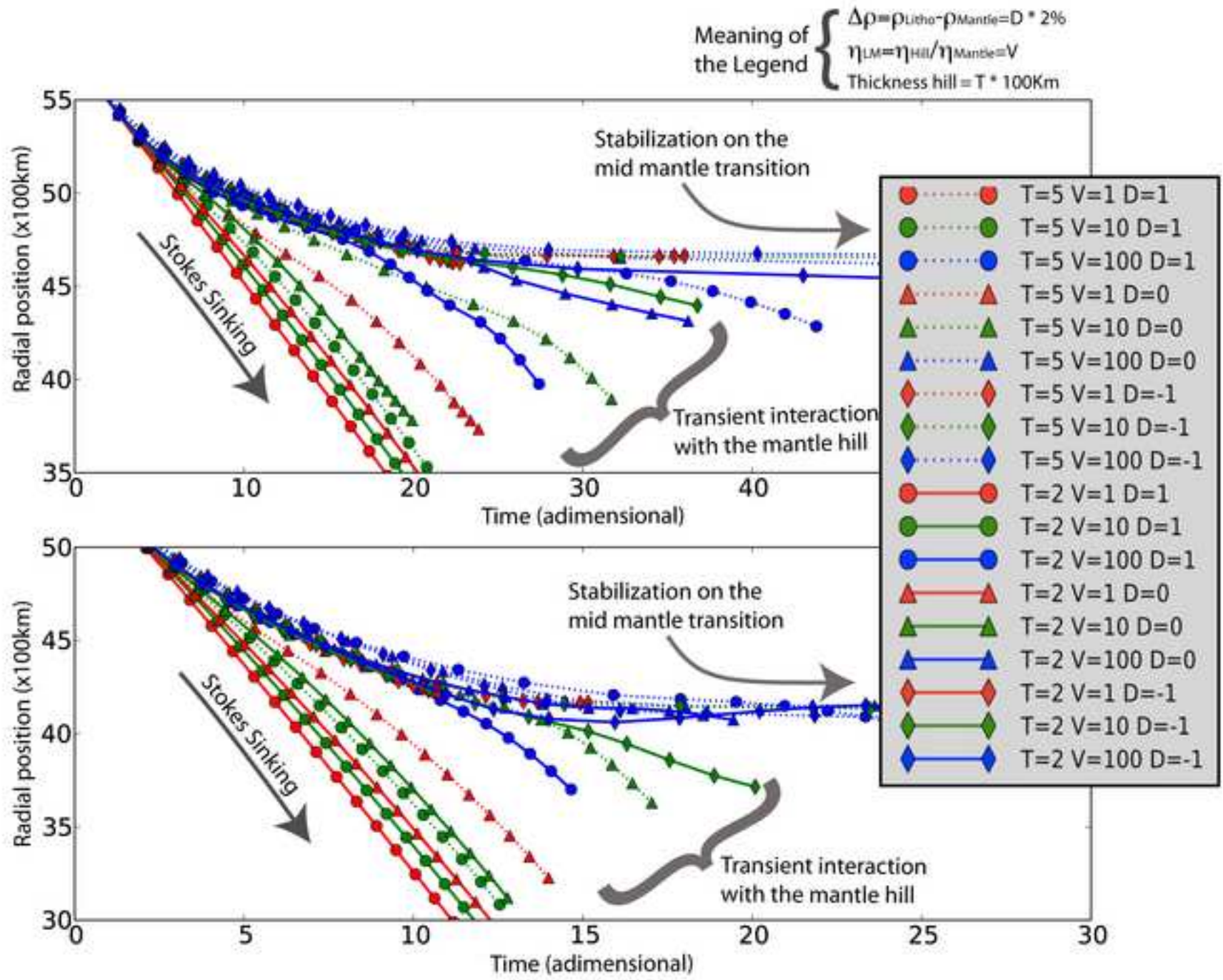
Preprint







Figure(s) 9



Tables

Hill thickness (km)	Viscosity Ratio ($\eta_R = \eta_{\text{Hill}} / \eta_{\text{Mantle}}$)	Differential density ($\Delta\rho = \rho_{\text{Litho}} - \rho_{\text{Hill}}$)	Slab width (km)	Fate of the slab (Stokes, Transient or Stalling)
200	1	-2%	2,000	Transient
200	1	-2%	10,000	Transient
200	1	0%	2,000	Stokes
200	1	0%	10,000	Stokes
200	1	2%	2,000	Stokes
200	1	2%	10,000	Stokes
200	10	-2%	2,000	Transient
200	10	-2%	10,000	Transient
200	10	0%	2,000	Stokes
200	10	0%	10,000	Stokes
200	10	2%	2,000	Stokes
200	10	2%	10,000	Stokes
200	100	-2%	2,000	Stalling
200	100	-2%	10,000	Stalling
200	100	0%	2,000	Transient
200	100	0%	10,000	Stalling
200	100	2%	2,000	Transient
200	100	2%	10,000	Transient
500	1	-2%	2,000	Stalling
500	1	-2%	10,000	Stalling
500	1	0%	2,000	Stokes
500	1	0%	10,000	Stokes
500	1	2%	2,000	Stokes
500	1	2%	10,000	Stokes
500	10	-2%	2,000	Stalling
500	10	-2%	10,000	Stalling
500	10	0%	2,000	Transient
500	10	0%	10,000	Transient
500	10	2%	2,000	Stokes
500	10	2%	10,000	Stokes
500	100	-2%	2,000	Stalling
500	100	-2%	10,000	Stalling
500	100	0%	2,000	Stalling
500	100	0%	10,000	Stalling
500	100	2%	2,000	Transient

500

100

2%

10,000

Stalling

Table 1: Summary of the results of the 36 experiments.

Color version of the table 2:

$\Delta\rho=\rho_{\text{Litho}}-\rho_{\text{Hill}}$ $\eta_{\text{R}}=\eta_{\text{Hill}}/\eta_{\text{Mantle}}$	2%	0%	-2%
1	Stokes	Stokes	Transient (T=200km) Stalling (T=500km)
10	Stokes	Stokes (T=200km) Transient (T=500km)	Transient (T=200km) Stalling (T=500km)
100	Transient	Transient (T=200km) Stalling (T=500km)	Stalling

Table 2: Synthetic summary of the results

BW version of the table 2:

$\Delta\rho=\rho_{\text{Litho}}-\rho_{\text{Hill}}$ $\eta_{\text{R}}=\eta_{\text{Hill}}/\eta_{\text{Mantle}}$	2%	0%	-2%
1	Stokes	Stokes	Transient (T=200km) Stalling (T=500km)
10	Stokes	Stokes (T=200km) Transient (T=500km)	Transient (T=200km) Stalling (T=500km)
100	Transient	Transient (T=200km) Stalling (T=500km)	Stalling

Table 2: Synthetic summary of the results