

The segregation of recycled basaltic material within mantle plumes explains the detection of the X-Discontinuity beneath hotspots: 2D geodynamic simulations

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Abstract

Mantle plumes are thought to recycle material from the Earth's deep interior. One constraint on the nature and quantity of this recycled material comes from the observation of seismic discontinuities. The detection of the X-discontinuity beneath Hawaii, interpreted as the coesite-stishovite transition, requires the presence of at least 40% basalt. However, previous geodynamic models have predicted that the percentage of high-density basaltic material that mantle plumes can carry to the surface is no higher than 15–20%. We propose this contradiction can be resolved by taking into account the length scale of chemical heterogeneities. While previous modeling studies assumed mechanical mixing on length scales smaller than the model resolution, we here model basaltic heterogeneities with length scales of 30–40 km, allowing for their segregation relative to the pyrolitic background plume material. Our models show that larger basalt fractions than previously thought possible—exceeding 40%—can accumulate within plumes at the depth of the X-discontinuity. Two key mechanisms facilitate this process: (1) The random distribution of basaltic heterogeneities induces large temporal variations in the basalt fraction with cyclical highs and lows. (2) The high density contrast between basalt and pyrolite below the coesite-stishovite transition causes ponding and accumulation of basalt at that depth, an effect that only occurs for intermediate viscosities of pyrolite. These results further constrain the chemical composition of the Hawaiian plume. Beyond that, they provide a geodynamic mechanism that explains the seismologic detection of the X-discontinuity and highlights how recycled material is carried towards the surface.

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8 **Key Points:**

- 9 • Plumes can carry enough basaltic material to explain observations of the X-
10 discontinuity at 300 km depth
11 • Accounting for the dynamics of individual chemical heterogeneities allows the
12 accumulation of more basalt than previously thought possible
13 • Near phase transitions, the local basalt fraction of plumes carrying 10-20% of
14 basalt on average can temporarily reach more than 40%

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Abstract

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We propose this contradiction can be resolved by taking into account the length scale of chemical heterogeneities. While previous modeling studies assumed mechanical mixing on length scales smaller than the model resolution, we here model basaltic heterogeneities with length scales of 30–40 km, allowing for their segregation relative to the pyrolitic background plume material. Our models show that larger basalt fractions than previously thought possible—exceeding 40%—can accumulate within plumes at the depth of the X-discontinuity. Two key mechanisms facilitate this process: (1) The random distribution of basaltic heterogeneities induces large temporal variations in the basalt fraction with cyclical highs and lows. (2) The high density contrast between basalt and pyrolite below the coesite-stishovite transition causes ponding and accumulation of basalt at that depth, an effect that only occurs for intermediate viscosities of pyrolite.

These results further constrain the chemical composition of the Hawaiian plume. Beyond that, they provide a geodynamic mechanism that explains the seismologic detection of the X-discontinuity and highlights how recycled material is carried towards the surface.

Plain Language Summary

Mantle plumes are thought to cause hotspot magmatism. Around 300 km beneath the Hawaiian hotspot, seismologic studies identified a jump in seismic velocities named the ‘X-discontinuity’. This feature is only observed at specific locations and is interpreted to be the result of a transformation in the quartz minerals in basaltic rocks.

For the X-discontinuity to be seismically visible, basalt fractions of 40% or more are required around 300 km depth. However, previous studies found that plumes can not carry more than 15–20% of heavier basaltic material. To overcome this contradiction, we create a series of models featuring a section of the plume and the basaltic material carried within it. In contrast to existing studies, we model the basaltic heterogeneities as individual inclusions that can move upwards or downwards with respect to the plume, rather than assuming that the basalt and the plume material are well-mixed.

We find that depending on the plume viscosity, basaltic material can pond and accumulate above and around 410 km depth, reaching peak fractions of more than 40%. Our models show how larger fractions of basaltic material than previously thought can accumulate within plumes in the upper mantle, thus reconciling the observations of the X-discontinuity.

1 Introduction

Geodynamic studies have long acknowledged the thermochemical nature of mantle plumes (Ballmer et al., 2013; Dannberg & Sobolev, 2015; Jellinek & Manga, 2004; Lin & van Keken, 2006; Lin & Van Keken, 2006). Several lines of evidence (e.g. A. V. Sobolev et al., 2005, 2007; S. V. Sobolev et al., 2011) indicate that recycled sub-

63 ducted eclogitic material is entrained (Christensen & Hofmann, 1994; Tackley, 2007)
64 and eventually brought to the surface by rising plumes.

65 Eclogite is denser than the surrounding mantle. Therefore, its entrainment criti-
66 cally influences plume dynamics in a number of ways that include: slowing down
67 or even halting plume ascent (Dannberg & Sobolev, 2015; Lin & van Keken, 2006;
68 Samuel & Bercovici, 2006), increasing the thickness of the plume conduit (Dannberg
69 & Sobolev, 2015; Kumagai et al., 2008), and facilitating an asymmetric plume shape
70 deviating from the classical head–tail structure (Ballmer et al., 2013; Farnetani &
71 Samuel, 2005). However, since only a fraction of the recycled material makes it to
72 the surface, very few constraints exist on the amount of denser recycled material that
73 remains in the mid-mantle.

74 Some of these important constraints come from seismology. The presence of ad-
75 ditional mineral phases has been linked to the local appearance of additional seismic
76 discontinuities in the mantle. One such discontinuity is the X-discontinuity, which
77 has been observed in the mid-upper mantle (~ 300 km depth) across a variety of tec-
78 tonic settings, including subduction zones (Revenaugh & Jordan, 1991; Schmerr et al.,
79 2013), hotspots (Kemp et al., 2019; Pugh et al., 2021), and mid-ocean ridges (Schmerr
80 et al., 2013). The X-discontinuity is seismically < 5 km thick (Revenaugh & Jordan,
81 1991; Bagley & Revenaugh, 2008; Pugh et al., 2021), with an impedance contrast of
82 3–8% (Schmerr et al., 2013; Chen et al., 2017; Bagley & Revenaugh, 2008). Its enig-
83 matic occurrence is commonly attributed to two mineral phase transitions. One such
84 phase transition occurs in the crystal structure of (Mg,Fe)SiO₃ pyroxene (Woodland
85 & Angel, 1997; Woodland, 1998; Jacobsen et al., 2010), which transforms from or-
86 thoenstatite to high-pressure clinoenstatite. The other involves the transition from
87 coesite to stishovite occurring in silica minerals (Deuss & Woodhouse, 2002; Williams
88 & Revenaugh, 2005; Bagley & Revenaugh, 2008; Schmerr et al., 2013). The latter pro-
89 duces an impedance contrast of 2–5% (Williams & Revenaugh, 2005) due to its large
90 change in seismic velocities (Chen et al., 2017; Faccenda & Dal Zilio, 2017; Williams
91 & Revenaugh, 2005). In order for the Co–St phase transition to occur, enough free
92 silica needs to be available. Eclogite, which is the stable mineral assemblage of basaltic
93 material under the pressure and temperature conditions around 300 km depth in the
94 mantle (Aoki & Takahashi, 2004; Faccenda & Dal Zilio, 2017), is often assumed to be
95 this source of silica.

96 Mineral physics studies show that in models where the mantle composition is
97 modeled as a mechanical mixture of basalt and harzburgite, stishovite is present re-
98 gardless of the basalt fraction in the mantle, and the X-discontinuity appears (Xu et
99 al., 2008). Conversely, stishovite is not present in an equilibrium assemblage where
100 the mantle is assumed to be pyrolitic in composition, unless basalt fractions are very
101 large (around or above 70%) (Xu et al., 2008). From a geodynamics perspective,
102 these findings mean we can use the occurrence of the X-discontinuity to constrain the
103 amount of basalt that is present as a separate chemical phase in mantle plumes.

104 The Hawaiian hotspot is a prominent example of mantle plume activity, where
105 both the seismic properties of the mantle and the geochemistry of the magmatic prod-
106 ucts have been extensively investigated. Geochemical constraints (Eiler et al., 1996;
107 Hauri, 1996; Hofmann & White, 1982; A. V. Sobolev et al., 2005, 2007) indicate the
108 presence of a substantial component of recycled basaltic crust in the plume. Sev-
109 eral seismic studies (Courtier et al., 2007; Kemp et al., 2019) have detected the X-
110 discontinuity beneath Hawaii using receiver functions and ScS reverberations. Kemp
111 et al. (2019) observed the X-discontinuity beneath Big Island at around 300 km depth,
112 finding a deeper, stronger signal beneath the eastern part of the island. In the same
113 area, their results show a very weak signal from the 410 km discontinuity. They sug-
114 gest that percentages of basaltic material between 40–50% are required to explain the

115 occurrence of the X-discontinuity, and up to 60-70% of eclogite is required to cause
 116 the disappearance of the 410 km discontinuity.

117 However, geodynamic studies (Dannberg & Sobolev, 2015) predict that a plume
 118 can carry no more than 15–20% dense eclogitic material to the surface. Models show
 119 that higher fractions of recycled basalt would render the plume negatively buoyant,
 120 causing it to stall in the deep mantle or transition zone and thus to never reach the
 121 surface.

122 To resolve the discrepancy between these different lines of evidence, we model the
 123 accumulation of eclogitic material within a mantle plume conduit. In particular, we
 124 focus on the depth range between 300 and 410 km, where the density difference between
 125 eclogite and the average mantle is especially large (Aoki & Takahashi, 2004; Stixrude &
 126 Lithgow-Bertelloni, 2011). Consequently, plumes carrying eclogite might slow down or
 127 even halt their ascent in this depth range (Ballmer et al., 2013; Dannberg & Sobolev,
 128 2015), but the high density contrast can also facilitate ponding of eclogitic material in
 129 a so-called Deep Eclogitic Pool (Ballmer et al., 2013). Here, we investigate whether
 130 this increased density contrast can provide a mechanism for the plume material to
 131 accumulate up to $\approx 40\%$ eclogite — the fraction needed to explain the appearance of
 132 the X-discontinuity in seismic studies.

133 We set up a series of geodynamic models featuring a pyrolitic background — rep-
 134 resenting a section of a plume conduit — and higher density chemical heterogeneities
 135 flowing in, representing the eclogitic recycled material. To do so, we adopt a new
 136 approach of modeling the motion of chemical heterogeneities. Most existing stud-
 137 ies assume that mechanical mixing of the different chemical components occurs on a
 138 length scale below the model resolution and that accordingly, the components in the
 139 mixture all move at the same velocity. Instead, we model the recycled material as
 140 having a larger length scale than the model resolution. This means that the basaltic
 141 heterogeneities can move upwards or downwards within the plume conduit at a dif-
 142 ferent velocity to the plume as a whole. Because the density contrast between the
 143 different chemical components varies with depth, this can allow for basaltic material
 144 to segregate and to accumulate to higher fractions than carried by the plume on av-
 145 erage. We perform 230 model runs, employing three different density profiles, a range
 146 of background viscosities between 10^{18} – 10^{21} Pa s, and five different inflow percentages
 147 of recycled material.

148 Our models predict the conditions under which chemical heterogeneities can pond
 149 in plumes and provide an upper limit for the eclogite fractions that can accumulate.
 150 These results link plume heterogeneity to the local occurrence of the X-discontinuity
 151 beneath some hotspots, shedding light on the underlying mechanisms behind its ap-
 152 pearance and allowing us to better constrain the Hawaiian plume composition.

153 2 Model setup

154 To model the plume conduit, we developed geodynamic models employing the
 155 mantle convection code ASPECT (Kronbichler et al., 2012; Heister et al., 2017; Bangert
 156 et al., 2021a, 2021b). ASPECT is a parallel, modular, open source, finite-element code
 157 written in C++. It is based on three libraries: *deal.II*, a general-purpose finite ele-
 158 ment library (Arndt et al., 2021), *Trilinos*, which performs linear algebra computations
 159 (Trilinos Project Team, n.d.; Kronbichler et al., 2012), and *p4est*, which handles the
 160 adaptive meshes (Burstedde et al., 2011).

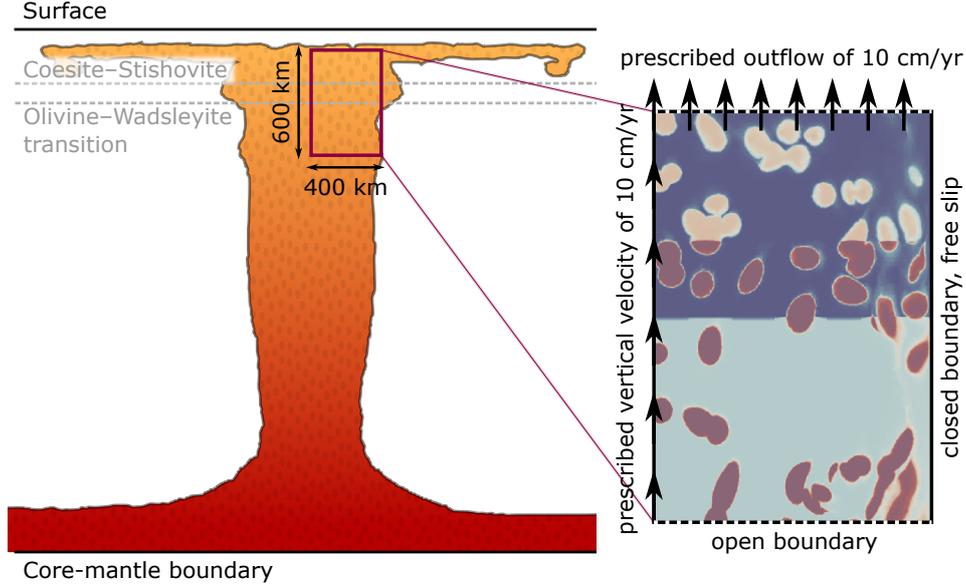


Figure 1. Model setup and boundary conditions. We model a section of the plume conduit in the upper mantle, including the Co–St transition at 300 km and the Ol–Wd transition at 410 km depth. Red and orange blobs represent chemical heterogeneities.

161 **2.1 Basic equations**

Over geologic timescales, the Earth’s mantle behaves like a viscous, slow-moving fluid (Schubert et al., 2001) whose convection can be described using the Stokes equations that state the conservation of mass (Equation 2) and momentum (Equation 1). We formulate these equations using the Boussinesq approximation, assuming that the material is incompressible and neglecting all density variations except for the buoyancy term $\rho\mathbf{g}$ in the momentum equation (Schubert et al., 2001; van Zelst et al., 2021).

$$-\nabla \cdot [2\eta\dot{\boldsymbol{\varepsilon}}'] + \nabla p = \rho\mathbf{g} \tag{1}$$

$$\nabla \cdot \mathbf{u} = 0 \tag{2}$$

162 Here, \mathbf{u} is the velocity, η the viscosity, $\dot{\boldsymbol{\varepsilon}}'$ the deviatoric strain rate tensor defined as
 163 $\dot{\boldsymbol{\varepsilon}}' = \frac{1}{2}(\nabla\mathbf{u} + (\nabla\mathbf{u})^T)$, p the pressure, ρ the density, and \mathbf{g} the gravity.

Since we do not take into account the effects of temperature, we do not solve the energy conservation equation. On the other hand, our models include chemical heterogeneities in the form of spherical basaltic inclusions within a pyrolitic matrix. The transport of the basaltic material can be described by the following advection equation:

$$\frac{\partial C}{\partial t} + \mathbf{u} \cdot \nabla C = 0, \tag{3}$$

164 with t being the time and C the composition, which in our case represents the basalt
 165 (eclogite) fraction. The fraction of the background pyrolite is then obtained as $1 - C$.
 166 We solve this advection problem adopting the particle-in-cell method (Gassmüller et
 167 al., 2018, 2019). Particles are generated at random locations at the start of the model
 168 run, and in cells where material enters the model. The initial composition of each
 169 particle is an input parameter, which we specify according to the initial conditions
 170 or boundary conditions (see Section 2.4). To interpolate composition from the parti-
 171 cle locations to the grid points, we employ a cell-wise constant averaging operation.

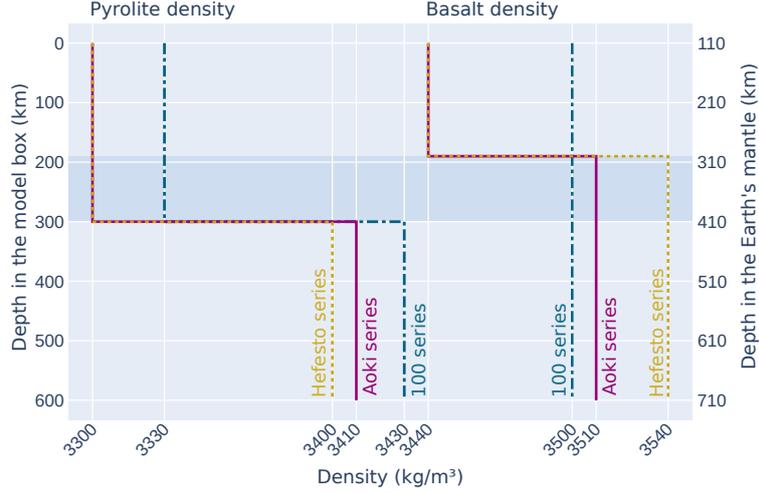


Figure 2. Density profiles for the three series. The left y-axis shows the depth in the model box, while the right y-axis refers to the real mantle depth. The curves on the left show the density distribution for the background pyrolite, while the right-hand side shows the values for the basaltic chemical heterogeneities. The sharp jumps in pyrolite density correspond to the Ol–Wd phase transition at 410 km depth. The density increase in the basaltic material is associated with the Co–St phase transition at 300 km depth (only included in the Aoki and Hefesto series).

172 Particle motion is computed using a second-order Runge-Kutta advection scheme. In
 173 order to make sure there are always enough particles for an accurate solution with-
 174 out unreasonably increasing the computational time, we impose a minimum of 10 and
 175 maximum number of 15 particles per cell.

176 **2.2 Model geometry**

177 The model setup consists of a 2D box, 400 km in width and 600 km in depth,
 178 representing a section of the plume conduit (Figure 1). The top of the box is located at
 179 110 km depth in the Earth’s mantle, since the model does not include the lithosphere.

180 To resolve the chemical heterogeneities, which have a diameter between 30 and
 181 40 km, we use an adaptive mesh to discretize the domain (Bangerth et al., 2021b),
 182 varying the cell size from 1.56–25 km through global and adaptive refinements. This
 183 strategy allows us to selectively resolve the smallest features in the model, as the mesh
 184 is refined only at locations where additional precision is needed (Gasmöller et al.,
 185 2018; Kronbichler et al., 2012). Specifically, we use the highest resolution where we
 186 estimate the error of the composition C to be large (based on the Kelly error estimator,
 187 Kronbichler et al., 2012) and at the bottom boundary where material flows in. This
 188 is required to make sure that the chemical heterogeneities are resolved as they enter
 189 the box.

190 **2.3 Density profiles**

191 We test three different density profiles in our models. All profiles include the
 192 olivine–wadsleyite (Ol–Wd) phase transition, which occurs at 410 km depth in the
 193 mantle. Relative to our model box, this depth corresponds to 300 km (as we do not
 194 include the lithosphere), which allows us to model two equal-sized halves with different
 195 densities (see Figure 2, left-hand curves).

196 The three profiles differ in terms of the basalt density distribution (Figure 2,
 197 right-hand curves). We include one profile (the 100 series) that features a uniform
 198 basalt density of 3500 kg/m^3 and a pyrolite density jump of 100 kg/m^3 at the Ol–Wd
 199 transition. In this series, the density difference between basalt and pyrolite increases
 200 from 70 kg/m^3 below the 410 km depth phase transition to 170 kg/m^3 above the phase
 201 transition. The 100 series is our simplest case, and we use it to analyze the general
 202 behavior of the heterogeneities when interacting with phase transitions before testing
 203 more complex scenarios.

204 The Aoki and Hefesto series adopt density values from Aoki and Takahashi (2004)
 205 and Stixrude and Lithgow-Bertelloni (2011), respectively. In addition to the Ol–Wd
 206 transition, they feature the coesite–stishovite (Co–St) phase transition in the basaltic
 207 material at 300 km depth, such that basalt has two different density values above and
 208 below (see Figure 2). Below the transition, the Hefesto series features the highest
 209 densities of basalt and the lowest densities of pyrolite, while the Aoki series has in-
 210 termediate values of pyrolite and basalt densities compared to the other two series.
 211 Given that the Co–St transition occurs at a shallower depth than the Ol–Wd transi-
 212 tion, the density contrast between basalt and pyrolite reaches its largest values in the
 213 depth range between 300 and 410 km depth in the mantle (190 to 300 km depth in
 214 our model box).

215 2.4 Boundary and initial conditions

216 Our model represents a section of an existing plume conduit. Therefore we
 217 assume that the material moves upwards within the plume, and that it can move faster
 218 in the hot center of the plume (left model boundary) compared to the colder plume
 219 margin. Accordingly, we prescribe the following boundary conditions (Figure 1): At
 220 the top and left sides, we impose a boundary velocity of 10 cm/yr in vertical direction,
 221 and zero in horizontal direction. The right side of the model—representing the plume
 222 margin—has an unrestrained velocity tangential to the boundary and a zero velocity
 223 normal to the boundary (free slip conditions). The bottom boundary is open and
 224 stress-free, so that material can flow in according to the forces acting in the model.

225 Consequently, our models are driven by the boundary conditions. We do not
 226 model the buoyancy forces of the plume itself, but instead assume that the plume is
 227 moving upwards with a fixed speed of 10 cm/yr and then investigate how chemical
 228 heterogeneities within the plume conduit affect its internal convection.

229 At the bottom boundary, where material flows into the model, the composition
 230 has to be prescribed. Specifically, we assume that the basaltic material is distributed
 231 within the plume in the form of circular inclusions at random locations, each with a
 232 diameter between 30 and 40 km. The number of basaltic inclusions flowing into the
 233 box within a given time is a model parameter that allows us to regulate the average
 234 fraction of basalt within the plume (see Section 2.6). No heterogeneities are present
 235 in the model at the start time. We let all models evolve for 100 million years, which
 236 is long enough for them to reach a steady state.

237 2.5 Model parameters

In order to analyze our models, we identify which parameters have the strongest
 influence on the motion of the chemical heterogeneities in the plume using a scaling
 analysis. This analysis allows us to predict the general physical behaviour of basaltic
 material in a plume and to select a reasonable parameter range. We assume that the
 behavior of the heterogeneities with respect to the background pyrolite is governed by
 three forces: (1) the drag force F_D (equation 4), which can be expressed by the Stokes
 law for a falling sphere in a laminar flow, (2) the gravitational force F_G (equation 5),

and (3) the buoyancy force F_B (equation 6):

$$F_D = 6\pi\eta rv, \quad (4)$$

$$F_G = \frac{4}{3}\pi r^3 \rho_s g, \quad (5)$$

$$F_B = \frac{4}{3}\pi r^3 \rho_b g, \quad (6)$$

238 with r being the radius of the spherical heterogeneity, v the relative velocity of that
 239 sphere with respect to the fluid it moves in, η the viscosity and ρ_b the density of that
 240 fluid, ρ_s the sphere density, and g the acceleration due to gravity.

Let us assume that the sphere accelerates until the drag force reaches a point where the net force on the sphere is zero. From that point onward, the sphere moves with a constant velocity, the terminal velocity v_T :

$$v_T = \frac{2r^2(\rho_s - \rho_b)g}{9\eta} \quad (7)$$

241 v_T is directly proportional to the square of the sphere radius, the density difference
 242 between the sphere and the background, and inversely proportional to the background
 243 viscosity. This means that in our models, the motion of the heterogeneities is governed
 244 by three parameters: the pyrolite viscosity, the density difference between basalt and
 245 pyrolite, and the diameter of the heterogeneities.

246 Given that the basaltic heterogeneities are denser than the pyrolitic background,
 247 v_T points downwards relative to the upwards flow within the plume. Consequently,
 248 we expect heterogeneities to sink, rise or hover, depending on our model parameters.
 249 Because there is a trade-off between these parameters, we can achieve the same regime
 250 by either changing the pyrolite viscosity, the density distribution, or the diameter of the
 251 spheres. Since the length scale of chemical heterogeneities in the mantle is unknown,
 252 we here choose a fixed diameter for the spheres of 30 – 40 km, which is enough for us to
 253 resolve them individually. This also allows us to observe the three different behaviors
 254 described above across a range of pyrolite viscosities that can be reasonably expected
 255 within a plume.

256 Because the density difference between pyrolite and basalt varies with depth, it
 257 is possible for the heterogeneities to show a combination of the different behaviors
 258 described above. For example, heterogeneities might rise upwards with the plume
 259 below 410 km depth, but then sink above 410 km depth, providing a mechanism for
 260 basalt to accumulate. To estimate at which conditions the spheres hover above the
 261 Ol-Wd phase transition, we can calculate the viscosity where v_T equals 10 cm/yr, the
 262 plume velocity we impose. Using the average value of 17.5 km for the sphere radius,
 263 a density contrast of 170 kg/m³ or 70 kg/m³ (as in the 100 series), and $g = 10$ m/s²,
 264 the resulting background viscosity range where we expect accumulations of basaltic
 265 material is $\eta = 1.5...3.7 \times 10^{19}$ Pa s. Note that this is only a rough estimate, since
 266 our models are two- instead of three-dimensional and they allow several spheres to
 267 interact.

268 2.6 Model Runs

269 In addition to using different density profiles (Section 2.3), we also vary the quantity
 270 of chemical heterogeneities flowing into the box by setting five influx percentages:
 271 10%, 15%, 20%, 30%, and 40%. These influx percentages correspond to the fraction
 272 of recycled material entrained by the plume, and they are implemented by placing
 273 spheres at random locations at the inflow boundary with a given probability (see Section
 274 2.4). As a result, some heterogeneities overlap with one another, and the actual
 275 influxes are lower than the influx percentages given above, specifically: 9.21%, 14.12%,

276 17.81%, 25.91%, and 32.86%. In the following, we will therefore refer to the different
 277 series by their rounded percentages: 10%, 15%, 20%, 25%, and 35%.

278 Previous studies (Dannberg & Sobolev, 2015, and references therein) have shown
 279 that plumes likely can not carry more than about 15-20% of dense eclogitic material
 280 towards the surface, as higher fractions would make them too dense to rise. Conse-
 281 quently, we use the higher basalt influx percentages (25% and 35%) to explore the
 282 parameter space, although they are likely not applicable to plumes in the Earth's
 283 mantle.

284 For all three density profiles, we run models with the five influx percentages above
 285 and across a range of pyrolite viscosities from 10^{18} – 10^{21} Pa s. The viscosity of the
 286 basaltic material is always two orders of magnitude larger than the pyrolite viscosity.
 287 The relative strength between pyrolite and eclogite in the Earth is uncertain and is
 288 expected to vary with temperature and pressure, depending on the stable mineral
 289 phases (Farla et al., 2017; Jin et al., 2001). But since eclogite bodies with a lower
 290 viscosity would have easily been stirred into the mantle, we here assume that eclogite
 291 has the higher viscosity. This prevents excessive deformation of the heterogeneities,
 292 especially in models where the pyrolite viscosity is low (Manga, 1996).

293 3 Results

294 We analyze the model runs based on the amount of basalt that accumulates
 295 around the two phase transitions and test whether basalt fractions as high as 40%, as
 296 indicated by seismological observations, are reached around 310 km depth. For the
 297 100 series, which only features the Ol–Wd phase transition, we compute the average
 298 basalt fraction between 310 and 510 km depth. For the Aoki and Hefesto series, we
 299 take into account the Co–St transition as well, and average the basalt fraction between
 300 260 and 360 km depth. We use these averages to classify the behavior of each model
 301 based on the maximum basalt fraction and the maximum root mean square velocity,
 302 and additionally analyze the evolution of the basalt fraction over time compared to
 303 the basalt influx, highlighting the mechanisms of basalt accumulation. However, we
 304 note that using these different depth ranges for computing basalt accumulation in the
 305 models affects their statistical properties (see Section 3.2).

306 3.1 Influence of the viscosity

307 The behavior of the chemical heterogeneities is first and foremost determined by
 308 the pyrolite viscosity. Depending on its value, we identify three different regimes: (1)
 309 *sinking*, (2) *mixed*, and (3) *rising*.

310 The *sinking* regime occurs for the lowest viscosity values and is highlighted in
 311 purple in Figures 4 and A1. In this regime, the gravity forces acting on the dense
 312 chemical heterogeneities exceed the frictional forces that would allow them to be carried
 313 upwards with the plume (see Section 2.5). Therefore, most heterogeneities sink shortly
 314 after entering the box. Only a few sparse and elongated strands of basalt rise above
 315 the Ol–Wd phase transition at 410 km depth, after undergoing extreme deformation
 316 (see Figure 3, top left). This strong deformation, enabled by the low viscosity, is
 317 also reflected by the temporally variable and generally very high root mean square
 318 velocities. They show values from 0.185 m/yr up to 0.796 m/yr, substantially deviating
 319 from the imposed boundary velocity of 0.1 m/yr.

320 We define a model as belonging to the sinking regime if the maximum basalt
 321 percentage is substantially lower or equal to the prescribed inflow percentage. The
 322 mean basalt percentages in the sinking regime are: 3.86% for the 100 series, 2.56% for

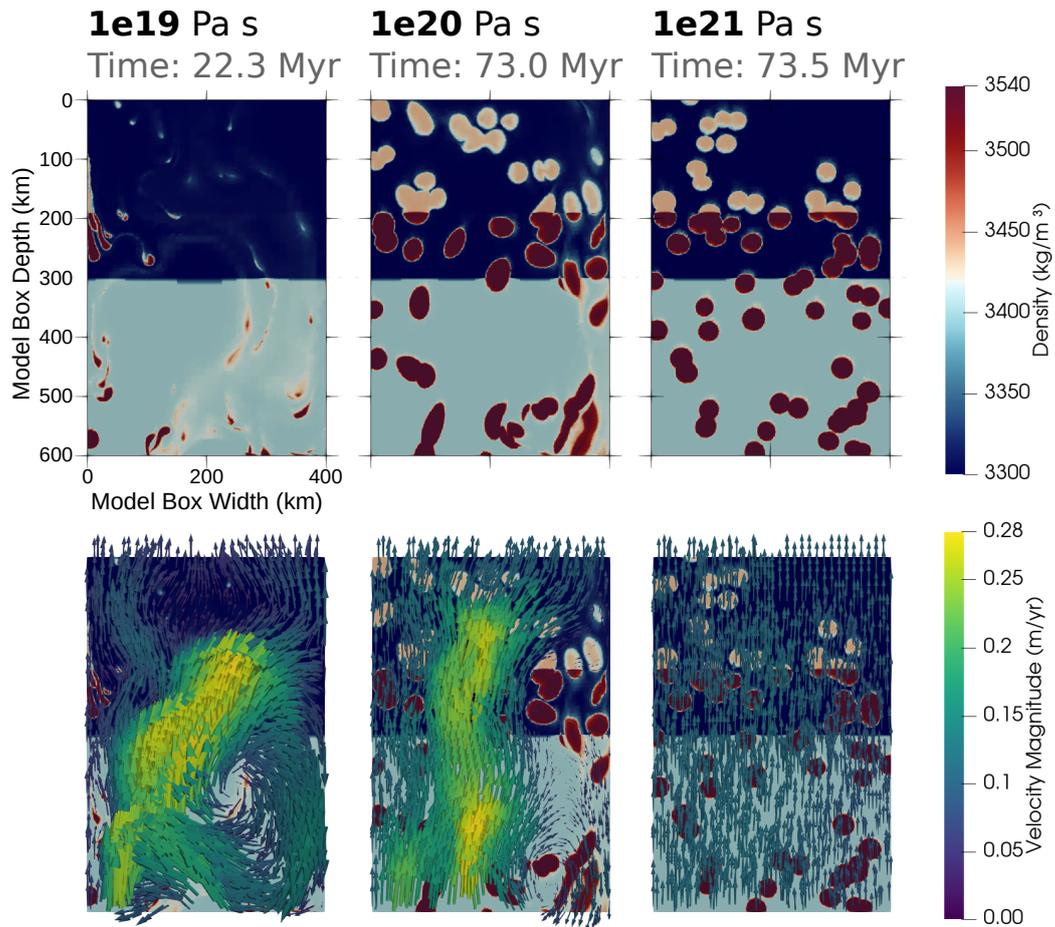


Figure 3. Distribution of the basaltic heterogeneities, as illustrated by the density (top row), and velocities in the model (bottom row) for the three different regimes, which are, left to right: sinking, mixed, and rising. The models shown are from the Hefesto series.

the Aoki series, and 2.11% for the Hefesto series, including the 25% and 35% influx runs.

At the opposite end of the spectrum is the *rising* regime, colored in green in Figures 4 and A1. For this regime, the pyrolite viscosity is so high that all chemical heterogeneities rise with the same velocity as the background (right panel in Figure 3), indicating the absence of deformation. We consequently define this regime based on the root mean square velocity: Whenever its maximum value is below 0.12 m/yr in a model throughout its whole evolution, we define the model as belonging to the rising regime. Regardless of the model series, pyrolite viscosities between 1 and 2×10^{20} Pa s mark the onset of the rising regime. The models belonging to this regime show high maximum basalt fractions, in many cases twice as high as the basalt influx.

We obtain the most noteworthy results for the *mixed* regime, which occurs at intermediate pyrolite viscosities and is highlighted in yellow in Figures 4 and A1. In this regime, the basaltic heterogeneities show a combination of rising and sinking motions. Consequently, heterogeneities cyclically pond around the Ol–Wd transition. This ponding reflects the density contrast between pyrolite and basalt and the associated downwards gravitational force, which is largest between the Co–St and the Ol–Wd transitions (see Section 3.3). Because the friction force stays the same, the heterogeneities can be carried upwards more easily below 410 km and above 300 km depth, but are more likely to sink in between. This makes their ascent path irregular, as they get entrained by the background flow to a variable degree. Therefore, depending on their position with respect to the model boundaries, the phase transition, and the degree of clustering of individual inclusions, they either sink, rise or pond (see Figure 3, center column).

Because basalt cyclically accumulates between 300 and 410 km depth, the fraction of basalt at this depth at times substantially exceeds the average fraction of basalt flowing in at the bottom of the model (Figure 5). The maximum percentages of basalt in the *dynamically consistent* models with basalt inflows up to 20% are between 16% and 43%.

The mixed regime consists of all the models that are neither in the rising nor in the sinking regime. Its lower boundary is determined by the basalt percentages, i.e. when the maximum basalt fraction is higher than the prescribed inflow. The upper boundary instead depends on the velocity, such that all models in the mixed regime reach a root mean square velocity of 0.12 m/yr or more averaged over the depth range of interest at some point throughout the model evolution.

The regime diagrams also show that the boundaries between regimes are not always horizontal lines. This is a direct consequence of the differences caused by the different basalt influx percentages. The chemical heterogeneities modeled here are always two orders of magnitude more viscous than the background. Therefore, the higher the fraction of basaltic material in the box, the higher the average viscosity. Note that basalt inflow fractions of more than 20% are unlikely to be present in plumes in the Earth, which is why we focus our analysis on the three lowest influx percentages only.

3.2 Evolution over time

In addition to analyzing the maximum basalt fraction in each model, it is also useful to look at how the basalt fraction around the phase transitions evolves over time. Figure 5 shows this evolution for three selected models of the Aoki series. The basalt fraction remains low in the sinking regime (blue line), while it varies dramatically over time in the mixed and rising regimes. In these two latter cases (purple and yellow lines), the evolution shows an initial increase in basalt fraction up to 20% as material flows

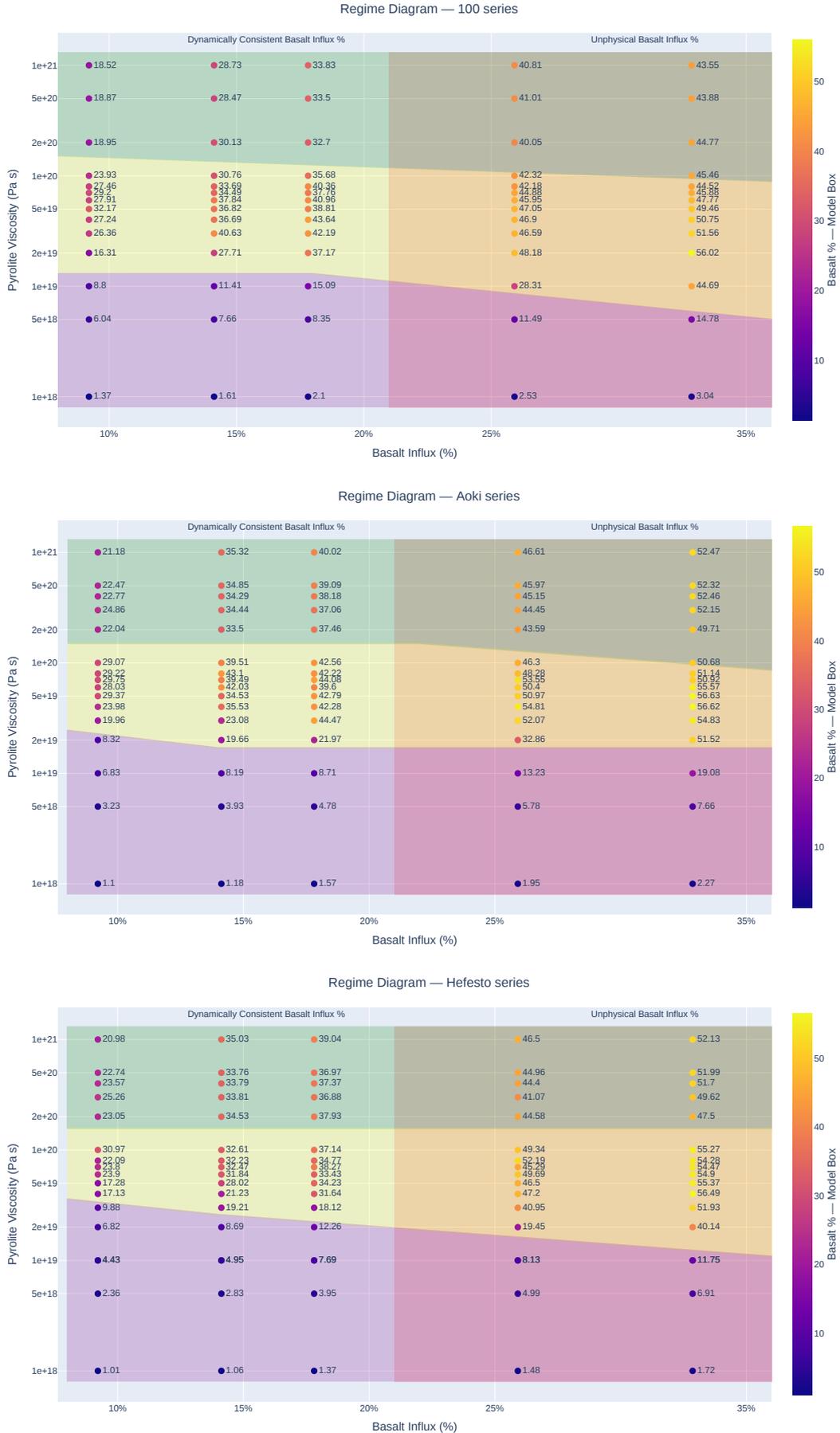


Figure 4. Regime Diagrams for the three series. The shading colors indicate, from bottom to top: the sinking regime (purple), the mixed regime (yellow), and the rising regime (green). The colors of the data points indicate the maximum percentage of basalt attained in the depth layer between 300 and 410 km depth.

373 in from below. Models in the mixed regime (purple line) then gradually accumulate
 374 more basalt, regularly reaching peaks above 30%, and sometimes even 40% or more.

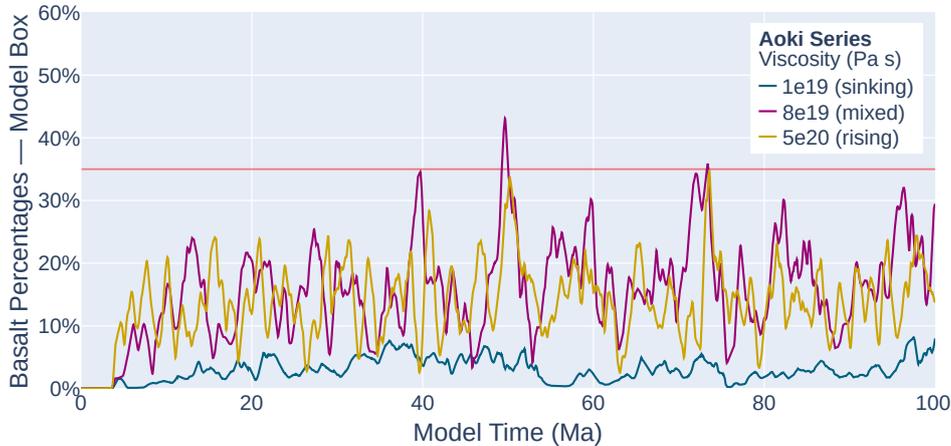


Figure 5. Evolution of the basalt fraction for three selected models in the Aoki series, all with a basalt inflow percentage of 15%. The model with a pyrolite viscosity of 8×10^{19} Pa s (mixed regime) shows the highest peak in the basalt fraction, reaching 40% at 50 Myr. The 5×10^{20} Pa s model (rising regime) displays a similar pattern, but has lower peaks of basalt fraction (up to approximately 32%), while the 10^{19} Pa s model (sinking regime) shows the lowest basalt percentage.

375 We note that basalt fractions comparable to the maximum values indicated by
 376 the regime diagrams (Figure 4) are only reached over short time intervals. Peaks in
 377 accumulation are followed by troughs, and throughout the majority of the model time,
 378 the basalt fractions remain consistently lower than these peaks, closer to the prescribed
 379 influx percentage.

380 To provide more context to our results in relation to the observations of Kemp
 381 et al. (2019), we further analyzed this evolution of the dynamically consistent models
 382 in the mixed regime. By integrating the time the basalt fraction exceeded 35%, we
 383 obtained the statistical probability of accumulating enough basaltic material for the
 384 X-discontinuity to be seismically visible at any given point in time (Figure 6). The
 385 threshold was chosen to be 35% as this is sufficiently close to the 40–50% of basalt esti-
 386 mated by Kemp et al. (2019) for the Hawaiian hotspot. This accounts for uncertainties
 387 in the seismic response to mineral phase transitions as evidenced by the discrepancy
 388 in amplitudes between receiver functions of different frequencies (Kemp et al., 2019;
 389 Pugh et al., 2021). Furthermore, 35% allows us to take into account a more substantial
 390 number of models in our analysis, with at least one model for each series.

391 The center panel of Figure 6 shows the results for a basalt influx of 20%. For the
 392 Hefesto series, the basalt fraction exceeds 35% only briefly – the probabilities are 1.19
 393 and 0.55%, respectively. In the Aoki series, the lower density contrast between basalt
 394 and pyrolite greatly increases the probabilities. We find the highest probability ($\sim 5\%$)
 395 of large basalt fractions in the middle of the mixed regime, for a pyrolite viscosity of
 396 6×10^{19} Pa s. The probability decreases to approximately 3% for lower and higher
 397 viscosity values, and sinks to 1% for a pyrolite viscosity of 3×10^{19} Pa s.

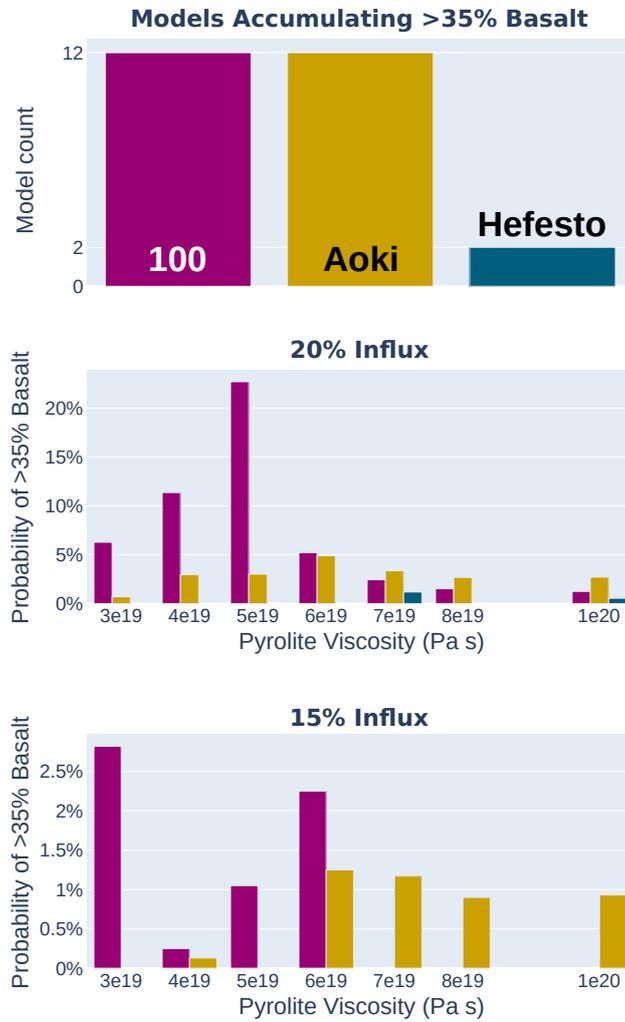


Figure 6. Statistical probability of dynamically consistent models to accumulate more than 35% basalt at any given point in time. The top panel shows the number of models reaching this basalt fraction. The middle and bottom panel display the probability in individual models with 20% and 15% basalt influx, respectively.

398 The 100 series displays ever higher probabilities. Basalt accumulation is most
 399 consistent for pyrolite viscosities of 3×10^{19} and 4×10^{19} Pa s. In these two cases, the
 400 probability of reaching more than 35% is the highest out of all models, and amounts
 401 respectively to 11.35 and 22.68%. The probability then gradually declines towards
 402 the edges of the mixed regime. These results highlight that basalt accumulates more
 403 consistently at the Ol–Wd transition compared to the coesite-stishovite transition (see
 404 also Sections 3.3 and 4).

405 The models with influx percentages of 15% show similar trends (Figure 6, right
 406 panel). In this case, only models from the 100 and Aoki series reach 35% of basalt,
 407 and the probabilities are about an order of magnitude lower than for the basalt inflow
 408 of 20%. Regardless of the series, none of the models featuring a 10% basalt inflow
 409 accumulate more than 35%. However, many models come close to peaks of 30%.

410 3.3 Influence of the density

411 The second relevant factor determining the model evolution is the density con-
 412 trast between the basaltic material and the background (Section 2.3). To illustrate
 413 its influence, we analyze the variability of the basalt fraction over time for a range of
 414 models (see Figure 7).

415 As expected, the basalt percentages reaching the phase transition are lowest for
 416 the lowest viscosities (sinking regime, top row of Figure 7). For viscosities of 10^{18} Pa s
 417 and 5×10^{18} Pa s (latter not shown), values never exceed 2%. This effect is mostly
 418 independent of the density.

419 The models in the mixed regime (yellow background in Figure 7) show the
 420 strongest differences between the three series, in particular for pyrolite viscosities of
 421 3×10^{19} (second row from the top). Again, the 100 series reaches the highest fractions
 422 of basalt, the Hefesto series features the lowest fraction, and the Aoki series falls be-
 423 tween the other two. The relative percentage difference between the 100 and Hefesto
 424 series, for the same basalt influx and pyrolite viscosity, reaches values as high as 77%,
 425 which is a direct consequence of the basalt density. The 100 series features a density
 426 difference of 70 kg/m^3 between pyrolite and the denser heterogeneities in the bottom
 427 half of the box, while the difference in the Hefesto series is twice as high (140 kg/m^3).
 428 The Aoki series shows an intermediate behavior, as the density difference between
 429 basalt and pyrolite is 100 kg/m^3 .

430 Consequently, the Hefesto series also reaches its maximum basalt concentrations
 431 at higher viscosities compared to the other two series, especially the 100 series, and
 432 systematically attains basalt percentages which are on average 5 to 7% lower compared
 433 to similar runs in the other series. All models in the mixed regime also show punctual
 434 peaks in the basalt fraction, exceeding 40% for some models with basalt influx of 20%
 435 and 15%.

436 For increasingly high pyrolite viscosities in the mixed regime (rows 3 and 4 of
 437 Figure 7), the over time trends gradually become more similar between the series,
 438 such that there is a significant amount of overlap between the three curves. The 100
 439 series shows slight decreases in the basalt quantities at viscosities higher than 8×10^{19} ,
 440 while the Aoki and Hefesto series reach the highest basalt fractions around pyrolite
 441 viscosities of 8×10^{19} and 10^{20} Pa s.

442 Rows 5 and 6 of Figure 7 show two models in the rising regime. In these models,
 443 the differences between the series have disappeared completely, to the point that three
 444 curves are not individually distinguishable. These observations reflect the inherent
 445 characteristics of the rising regime. As the velocity of the heterogeneities is very close
 446 to the value prescribed as boundary condition, and the pyrolite viscosity is high enough

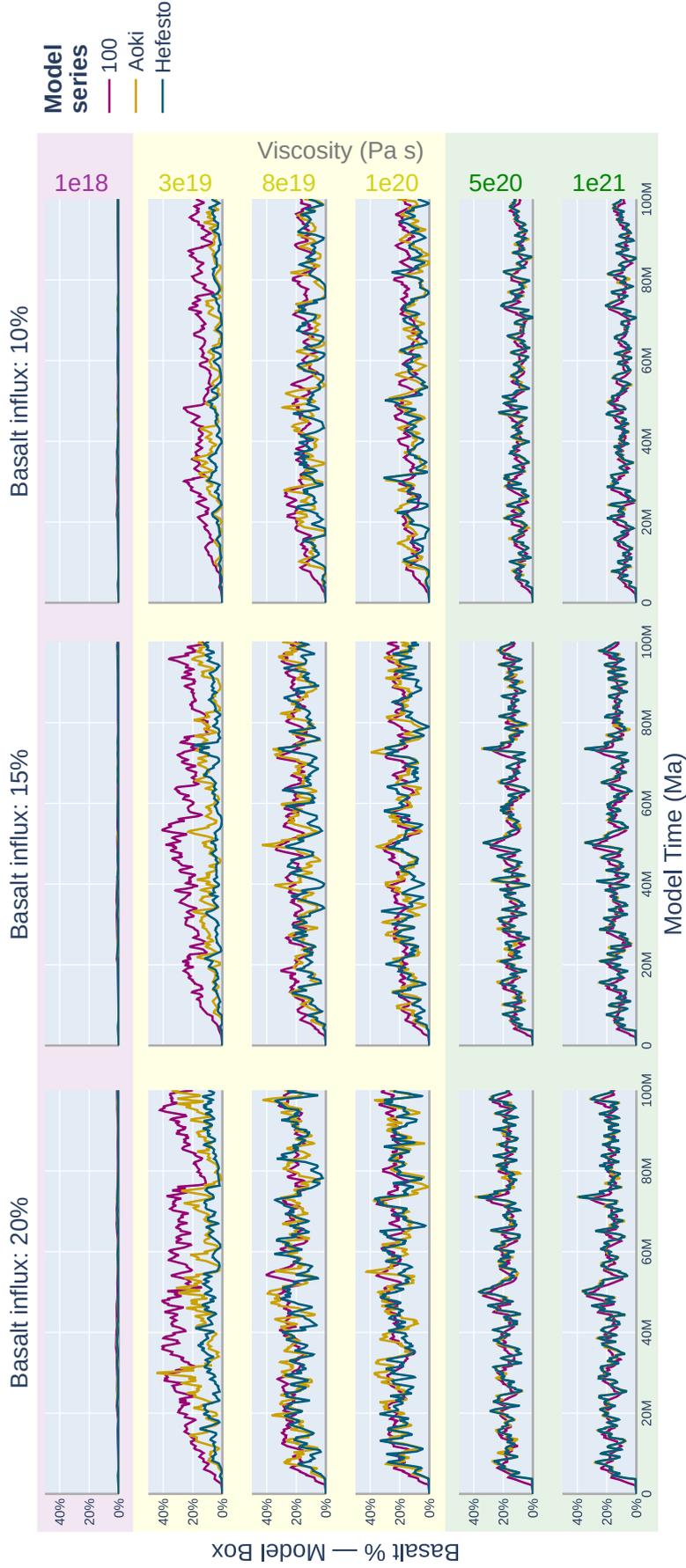


Figure 7. Evolution of the basalt fraction over time for selected pyrolite viscosity values (row numbers, right hand side) and basalt influx (columns). The background color scheme is the same as in Figure 4, with purple identifying the sinking regime, yellow the mixed regime, and green the rising regime. Only the three dynamically consistent influxes are shown (from left to right: 20%, 15%, and 10%).

447 to carry the denser material, the heterogeneities are transported with the same speed
 448 in all models and very little differences exist between them. At the highest viscosity
 449 (row 6), differences of the 100 series compared to the other models are mainly caused
 450 by the different depth ranges we use for averaging the basalt fraction. The basalt
 451 fractions in the Aoki and Hefesto series evolve almost identically.

452 4 Discussion

453 While previous studies indicated that plumes can carry no more than 15-20%
 454 denser material on their way through the mantle, our results show that there is a
 455 range of conditions under which much larger fractions of basalt can accumulate within
 456 a plume conduit that on average carries about 15-20% of basalt. The highest amount of
 457 basalt in the depth range between 310 and 510 km for the 100 series, and between 260
 458 and 360 km depth for the Aoki and Hefesto series, is attained for intermediate viscosity
 459 values that define the mixed regime. Specifically, assuming a radius of 17.5 km for the
 460 heterogeneities, pyrolite viscosities of at least $2 - 3 \times 10^{19}$ Pa s, but not more than
 461 $1 - 2 \times 10^{20}$ Pa s are required for the material to pond between the Ol–Wd and Co–St
 462 phase transitions. Slightly lower maximum basalt fractions are attained for higher
 463 viscosities in the rising regime.

464 Although our models feature high percentages of basalt near the Co–St transition
 465 both in the mixed and the rising regime, the underlying mechanisms are fundamentally
 466 different. Notably, the amount of basalt does not remain constant throughout the
 467 model evolution (Figure 5). Instead, it varies statistically around the influx percentage
 468 and displays highs and lows. This variation is caused by two concurrent mechanisms.

469 In all models, the statistical variations in the distribution of the basaltic heterogeneities—
 470 which flow into the model at random locations—causes cyclical highs and lows in the
 471 amount of basalt at a given depth. On the length scale of the seismologic resolution,
 472 these variations can lead to maximum basalt fractions in the plume of more than dou-
 473 ble of the influx percentage of basalt. Consequently, even if heterogeneities do not
 474 pond and accumulate, such as in the rising regime, the absolute percentages of basalt
 475 fraction at a given point in time might still exceed the average amount of basalt within
 476 the plume.

477 In the mixed regime, there is an additional mechanism that leads to increased
 478 basalt fractions: Basaltic material ponds in the depth range between the two dis-
 479 continuities. Accordingly, the residence time of the heterogeneities within the plume
 480 conduit increases and they can accumulate. This effect is not as strong as the statisti-
 481 cal variation, however, it increases not just the peaks, but also the temporal average
 482 of the basalt fraction at a given depth. The combination of the two effects leads to
 483 overall higher percentages of basalt of 35-40% or more even for basalt influxes of 20%
 484 or less. Consequently, the distribution of basalt in the mixed regime can best explain
 485 the seismological observations.

486 The effect of the two competing mechanisms is illustrated in Figure 8, which
 487 shows different models of the mixed regime. While the basalt fraction varies over
 488 time, its time average is also slightly higher than the basalt influx, indicating the
 489 accumulation of basalt around the phase transition. This effect is most pronounced
 490 in the 100 series (left column), where the time-averaged basalt fraction can be 5-10%
 491 larger than the amount of basalt flowing in. The Aoki series (center column) shows
 492 this effect to a lesser extent, and in the Hefesto series (right column) the average and
 493 influx basalt fractions are comparable. In the rising regime (not shown) this effect is
 494 absent and the model average is equal to the basalt influx.

495 With regards to the modeled phase transitions, these results highlight the im-
 496 portance of the Ol–Wd transition for the accumulation of basalt. At this transition,

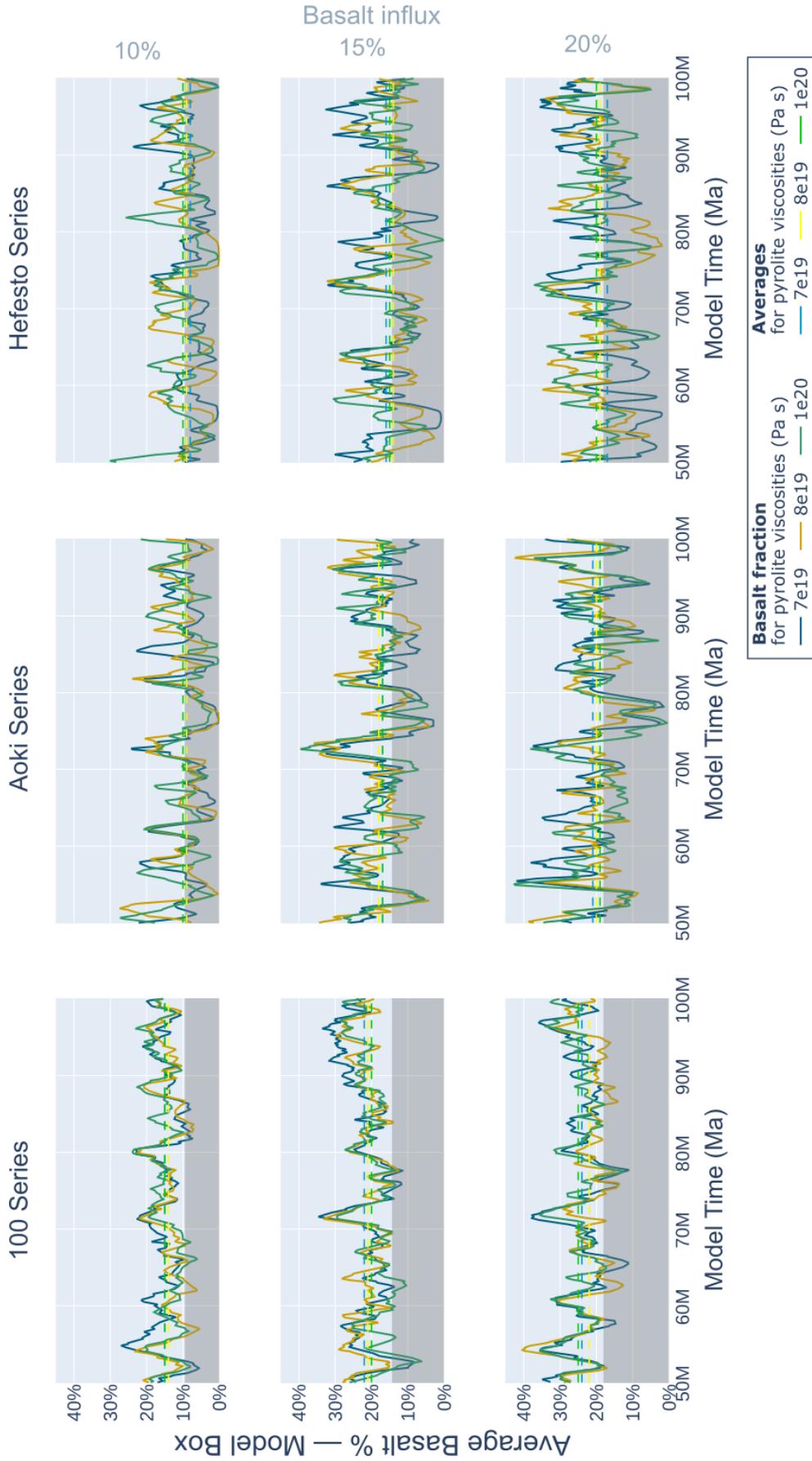


Figure 8. Evolution of the basalt fraction over time (as in Figure 7, solid lines) and average basalt fraction (dashed horizontal lines) for three selected pyrolyte viscosities (darker colors in the legend). The gray boxes indicate the basalt influx percentages, as indicated to the right of each row. The different columns refer to the three different model series. The 100 series shows the most marked basalt accumulation, while the Hefesto series displays the least average basalt with respect to the influx.

497 the heterogeneities slow down as a consequence of the increased density contrast above
 498 410 km depth, and our models predict the strongest effect of hovering. This explains
 499 why the effect is strongest in the 100 series, where we average in a depth range of 310
 500 to 510 km, around the Ol–Wd transition. In contrast, the average basalt fractions in
 501 the Aoki and Hefesto series (depth range of 260 to 360 km, around the X-discontinuity)
 502 are only slightly higher than the influx. The Co–St transition, on the other hand, does
 503 not play a primary role for the model dynamics. Even the models of the 100 series
 504 that do not include this additional transition show substantial accumulation of basalt.

505 4.1 Implications for the Hawaiian mantle plume

506 Our observations only partially corroborate the findings in Kemp et al. (2019).
 507 The authors tentatively explain the suppression of the 410 km depth discontinuity
 508 beneath the Big Island of Hawaii with the occurrence of extremely high quantities
 509 of basaltic material, up to 60-70%. While we observe the strongest effect of ponding
 510 directly above 410 km depth, we never observe more than 45% basaltic material, as-
 511 suming physically feasible influx percentages. It may however be possible that locally,
 512 more than 45% basalt accumulates within the mantle plume. Our models average the
 513 fraction of basalt over the whole cross section of the plume conduit, whereas the re-
 514 ceiver functions beneath the Big Island of Hawaii indicate the vanishing of the 410 km
 515 discontinuity only for a part of the plume. Consequently, a large accumulation of
 516 basaltic heterogeneities on the length scale of the resolution of the study of Kemp et
 517 al. (2019) (~ 100 km) might explain this observation.

518 It remains challenging to reconcile the observation of the X-discontinuity in mul-
 519 tiple plume locations (Pugh et al., 2021). With the time-varying nature of the basalt
 520 fraction in our models, it is implausible that so many plumes have concentrations of
 521 basalt above the 40% threshold required for seismic observation. However, several seis-
 522 mic studies attribute their observations to two, or more, causal mechanisms (Bagley
 523 & Revenaugh, 2008; Pugh et al., 2021).

524 4.2 Model limitations

525 In order to analyze and observe the behavior of the basaltic heterogeneities, we
 526 have made several simplifying assumptions.

527 First, the dynamics of our models does not take into account the influence of
 528 temperature or compressibility. In the Earth’s mantle, a plume would rise with a
 529 variable velocity, depending both on its temperature compared to the surrounding
 530 material and its chemical composition. Phase transitions would occur at a variable
 531 depth depending on their Clapeyron slope and the temperature distribution within the
 532 plume conduit, introducing additional variations in buoyancy. In contrast, our models
 533 prescribe a constant rising velocity for the modeled section of the plume, only taking
 534 into account compositional density variations. This means that the flow in our modeled
 535 plume conduit is still dominantly upwards, even if the negative buoyancy introduced by
 536 the dense basaltic heterogeneities exceeds the positive buoyancy due to the high plume
 537 temperatures. Consequently, we rely on existing studies (e.g., Dannberg & Sobolev,
 538 2015) to constrain the maximum amount of basalt that plumes can carry towards the
 539 surface and to identify which of our models are applicable to plumes in the Earth’s
 540 mantle (see *dynamically consistent* models, Figure 4).

541 Another simplification is the model geometry. Because our side boundaries are
 542 closed, plume material is constrained in its horizontal motion, fixing the shape of the
 543 plume conduit. In the Earth, one would expect that plume material that is cyclically
 544 rising and sinking at 300-400 km depth would move laterally, away from the center of
 545 the plume conduit. This may lead to the accumulation of a large amount of basaltic

546 material around the plume conduit (similar to a deep eclogitic pool Ballmer et al.,
 547 2013). On the other hand, the basaltic material is likely to sink downwards as soon as
 548 it has moved laterally and is no longer supported by the upwards flow within the plume
 549 conduit. Plume material would also be deflected to the side when reaching the base
 550 of the lithosphere, which we did not include in our models. How exactly the results
 551 would change remains an open question to be answered in future studies. However, we
 552 believe our main conclusion—that locally, more than 35-40% of basalt can accumulate
 553 in plume conduits despite its large density—remains unaffected by this assumption.

554 Uncertainty in our models also comes from the mantle composition. We here
 555 assume that the background mantle material has a pyrolitic composition. However,
 556 seismological studies (Cammarano et al., 2009; Ritsema et al., 2009; Xu et al., 2008)
 557 show that in the mantle transition zone, mineral physics predictions fit observed seismic
 558 velocities overall better when taking into account a mechanical mixture of basalt and
 559 harzburgite compared to a pyrolitic equilibrium assemblage.

560 Depending on the length scale of heterogeneities in the mechanical mixture, the
 561 background mantle composition within the plume could also be harzburgitic, or it
 562 could be a mechanical mixture, with basalt and harzburgite being mixed on a much
 563 smaller length scale than the diameter of 30 – 40 km we assume for our basaltic
 564 heterogeneities. Both possibilities would affect the model predictions. The small-scale
 565 mechanical mixture would have a similar density profile as the one we use for pyrolite,
 566 so the effect on the model dynamics is likely to be small. However, it would contain
 567 some free SiO₂, increasing the effective total amount of coesite/stishovite in the plume
 568 and making the detection of the X-discontinuity more likely, even with basalt fractions
 569 slightly less than 40%. If the background composition would be harzburgitic, that
 570 would slightly increase the density contrast between the basaltic heterogeneities and
 571 the background, shifting all regimes to slightly higher viscosities. But it would also
 572 make the plume overall less dense, allowing it to carry slightly more recycled basaltic
 573 material while still being positively buoyant, and therefore allowing larger basalt influx
 574 percentages.

575 5 Conclusions

576 We simulate the ascent of dense chemical heterogeneities within a plume con-
 577 duit to determine the maximum percentages of basaltic material that can accumulate
 578 within a rising mantle plume in the depth around the Ol–Wd and Co–St phase tran-
 579 sitions. Our results show that, even for basalt influx percentages lower than 20%—
 580 consistent with constraints from earlier geodynamic studies—the basalt fraction in a
 581 mantle plume can reach peaks of 40% or more. Our study demonstrates that basalt can
 582 accumulate within a plume conduit to much higher fractions than previously thought.
 583 It also provides a viable mechanism explaining the underlying dynamics that lead to
 584 the seismic observations of the X-discontinuity in the mid-upper mantle beneath the
 585 Hawaiian hotspot.

586 We also explored the conditions under which denser material can accumulate.
 587 Our models show that there are two fundamentally different underlying mechanisms
 588 that can lead to large fractions of basalt within the plume conduit. On the one
 589 hand, basalt ponds above the Ol–Wd transition. This is an effect of the interplay
 590 between downwards gravity forces—which are enhanced at this depth because the
 591 density contrast between basalt and pyrolite is largest—and frictional forces due the
 592 upwards motion within the plume conduit. On the other hand, statistical variations
 593 due to the random distribution of the basaltic heterogeneities cause cyclical highs and
 594 lows in the amount of basalt at any given depth. The latter effect is stronger in our
 595 models. On the length scale of the seismologic resolution, statistical variations can
 596 lead to maximum basalt fractions of more than double of the average amount of basalt

597 in the plume. Conversely, ponding of heterogeneities only increases the basalt fraction
 598 by about one third to half of the amount of basalt being carried upwards from the
 599 lower mantle, and is strongest around the depth of the Ol–Wd transition.

600 For a diameter of the basaltic heterogeneities of 30–40 km, the strongest accu-
 601 mulation of denser material occurs when the pyrolite viscosity is between 3×10^{19}
 602 and 1×10^{20} Pa s, a range that encompasses our mixed regime. At lower values,
 603 only little basalt is entrained into the upwards flow. Higher values do not allow for
 604 the heterogeneities to move relative to the rest of the plume. Therefore, basalt can
 605 not accumulate and high percentages are only due to statistical variations. Since the
 606 statistical effect of basalt accumulation is dominant, high values of 35–40% or more
 607 cannot be sustained for a prolonged amount of time. Under the most favorable cir-
 608 cumstances (a basalt influx of 20%, a viscosity of 6×10^{19} , and using densities from
 609 Aoki and Takahashi (2004)), the basalt fraction at ~ 300 km depth exceeds 35% for
 610 about 5% of the time.

611 Further research is needed to investigate if lateral spreading of the material pond-
 612 ing above the Ol–Wd transition could increase the accumulation effect and how plume
 613 dynamics would be affected by the large basalt fractions.

614 Appendix A Velocity Diagrams

615 Figure A1 shows the maximum value of root mean square of the velocity, averaged
 616 over the depth range of interest, attained in each model. As outlined in Section 3.1, we
 617 used these values to define the boundary between the mixed and rising regimes. The
 618 prescribed boundary velocity is 0.1 m/yr, a value that corresponds to the absence of
 619 deformation. Since deformation is minimal in the rising regime, models in this regime
 620 have a root mean square velocity that is close to 0.1 m/yr. We therefore classify a
 621 model as being in the mixed regime (or sinking regime) if its average velocity in the
 622 depth range of interest for each of the three series exceeds 0.12 m/yr at any point
 623 in time. Mixed and sinking regime are distinguished based on their basalt fraction
 624 (Section 3.1).

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634 All software and data used to generate the results and figures in this manuscript
 635 are freely available. The specific files and instructions are published as a Zenodo
 636 data package (<http://doi.org/10.5281/zenodo.6687407>) and include the source code
 637 and links to the GitHub versions of ASPECT as well as the input files to compute the
 638 models and plotting scripts to generate the figures.

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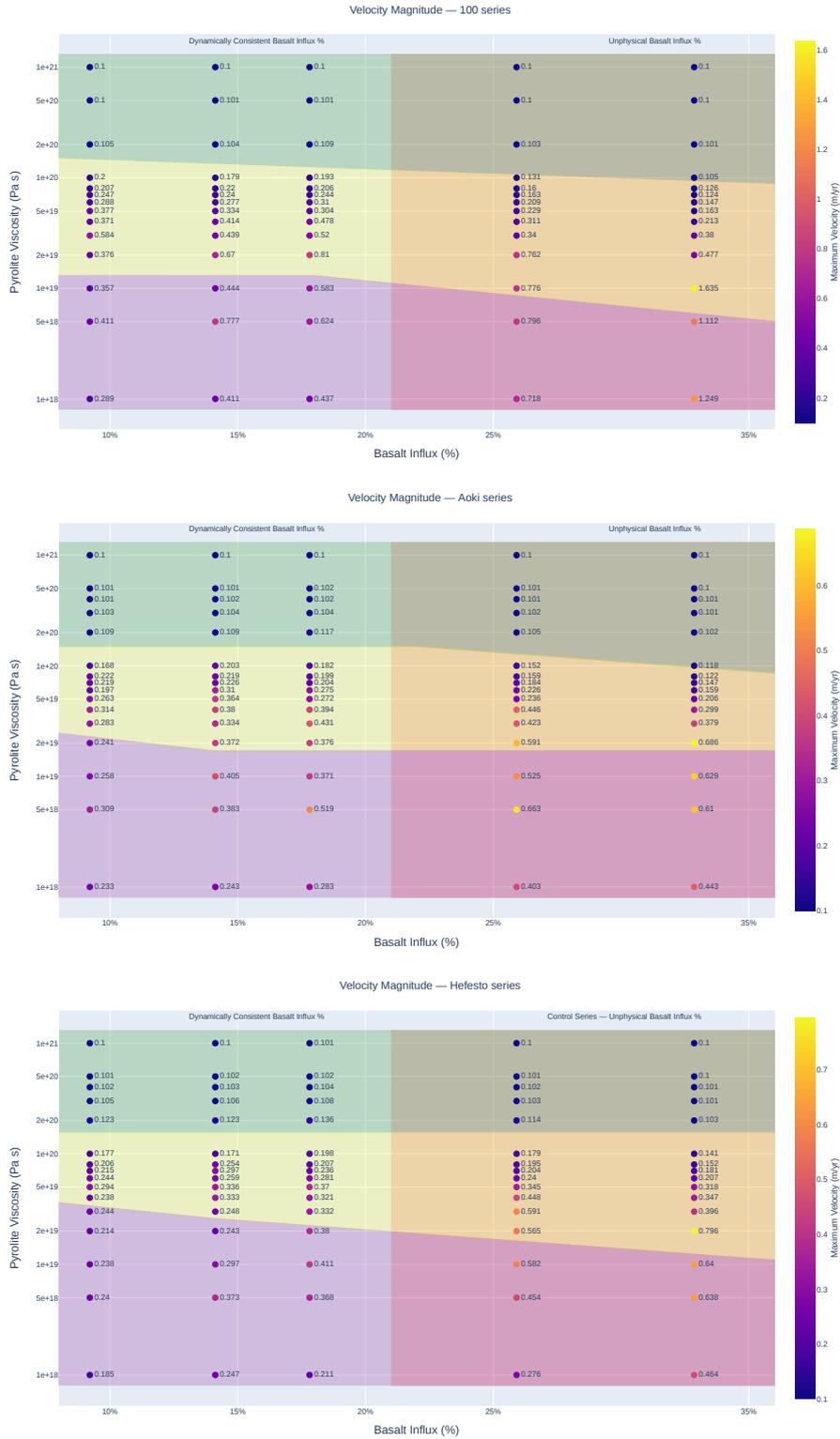


Figure A1. Velocity diagrams for the three series. The color of each point and the value next to it indicate the root mean square velocity in the model, averaged in the depth range of interest (between 260 and 360 km mantle depth in the Aoki and Hefesto series, and between 310 and 510 km depth in the 100 series). The background colors are the same as in the respective regime diagrams (Figure 3), and indicate, from bottom to top: the sinking regime (purple), the mixed regime (yellow), and the rising regime (green).

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