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HAL Authorization

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2

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12

13 **Abstract:**

14 Formation of a hydrated superficial layer early in the history of the Earth and Earth-like
15 planets may occur by extensive serpentinization of mantle rocks by water, similar to that still
16 active at present-day in slow-spreading oceanic plates. We investigated the lifetime of this
17 buoyant and weak hydrated "crust" with 2D numerical simulations of mantle convection.
18 Three regimes of preservation of the hydrated crust exist as a function of its thickness and
19 buoyancy, both of which depend on the intensity of hydration. Firstly, an unlikely regime for
20 the Earth is that where a thick buoyant crust (>100 km) can form a convective lid above the
21 mantle. Secondly, a regime exists where weakly hydrated crust disappears through convection
22 at a rate similar to the reference null hydration model. The lag time before which convection
23 starts decreases with increasing thickness, decreasing the potential survival of the hydrated
24 crust. Finally, a regime is found where rafts of highly hydrated crust of intermediate thickness
25 are preserved at the surface. This regime is possible on the early Earth, and the survival time
26 of the hydrated crust is proportional to its thickness. The present models indicate that a 50
27 km-thick fully serpentinized layer may have formed and survived over about 1 Gyr at the
28 surface the early Earth, and could have favored the formation of the first differentiated felsic
29 crust by partial melting.

30

31 **Keywords:** hydrated crust, serpentine, early Earth

32

33 **1. Introduction**

34 Hydration of the ultramafic to mafic rocks results in the formation of hydrous
35 phyllosilicates like serpentine and chlorite. They play a major role in tectonic activity (Guillot
36 et al., 2015) and subduction style (Gerya et al., 2008) because of their low yield strengths
37 (Amiguet et al., 2012; Amiguet et al., 2014). A fully serpentinized crust is nearly as buoyant
38 as continental crust but is reversibly transformed to anhydrous mantle rocks by dehydration in
39 subduction (Ulmer and Trommsdorff, 1995), hence it has a limited lifetime once plate
40 tectonics is active. Before the onset of plate tectonics and extensive formation of continental
41 crust, an early Hadean hydrated ultramafic crust may have formed in a context of low-tectonic
42 mobility (Boyet et al., 2003) and in the presence of a thick hydrosphere (Korenaga, 2007).

43 The early formation of a serpentinized crust in the Earth's history requires several
44 conditions. A first condition is that water was available at the surface of the planet to form the
45 serpentinized crust together with ultramafic to mafic rocks. The end of Earth accretion and
46 differentiation is deemed to occur after the Moon-forming impact circa 4.4-4.5 Ga (Jacobson
47 et al., 2014; Touboul et al., 2007), concomitant to major differentiation in the silicate Earth at
48 4.4 Ga as traced by the ^{146}Sm - ^{142}Nd system (Carlson et al., 2019). Zircons from Jack Hills
49 (Australia) ranging in age from 4.3 to 3.9 Ga were sourced from melts of weathered material
50 that formed in the presence of water (Mojzsis et al., 2001; Wilde et al., 2001), suggesting
51 water was present at the Earth's surface soon after the Moon-forming impact.

52 A second condition is that serpentine is stable, which imposes requirements on the
53 superficial temperature gradient. At present time, serpentine is preserved at depths only in
54 subduction zones (Reynard, 2013), as it destabilizes back to olivine and pyroxene at
55 temperatures of about 900 K (Ulmer and Trommsdorff, 1995). It is marginally stable at the
56 top of the mantle below oceanic or continental crust, except in slow-spreading oceans with
57 core-complex structures in the crust (MacLeod et al., 2002). Archean plate tectonics in a

58 globally warmer Earth (van Hunen and van den Berg, 2008) favors thicker oceanic crust
59 rather than serpentinization. The modes of cooling that are favorable for serpentinization over
60 great depth assumed here are the heat-pipe (Moore and Webb, 2013) and squishy lid regimes
61 (Lourenço et al., 2020) that were proposed to account for Hadean/Archean Earth conditions.
62 Mass and heat flow through localized hot volcanic conduits and plutonic sills, while a slowly
63 sinking lithosphere is maintained at temperatures below 900 K down to depth of up to ~200
64 km (Moore and Webb, 2013), allowing potential serpentinization and serpentine preservation
65 down to these depths.

66 We explored through numerical models of thermo-chemical convection how such
67 hydrated crust may have persisted at the surface of the early Earth in a context of convection
68 onset.

69

70 **2. Methods**

71 We studied the stability of a hydrated layer (or primitive serpentinized crust) at the
72 surface of a convective system. The hypotheses are that the hydrated layer has a lower
73 density, lower viscosity and lower pseudo-plastic yield stress than mantle rocks, and that
74 mantle dynamics is dominated by solid-state convection. Therefore we build a convection
75 model with two distinct materials, being the mantle and hydrated rocks. The initial setup
76 places the hydrated layer at the top of the system. We defined a depth of 300 km at which
77 complete dehydration takes place, switching hydrated rocks to ambient mantle. Therefore
78 sinking of hydrated rocks would lead to progressive vanishing of the initial layer. The purpose
79 is to identify different regimes of stability by extracting time scales for survival/entrainment
80 of this layer.

81 We model convection with the code StagYY in 2D spherical annulus geometry
82 (Hernlund and Tackley, 2008), solving the following coupled non-dimensional equations:

83
$$\nabla \cdot v = 0 \quad (1)$$

84
$$\nabla \cdot \eta[\nabla v + (\nabla v)^T] - \nabla p = Ra(T + B_c C)e_r \quad (2)$$

85
$$\partial_t T + v \cdot \nabla T = \nabla^2 T + H \quad (3)$$

86
$$\partial_t C + v \cdot \nabla C = 0 \quad (4)$$

87 where v is the velocity, p the dynamic pressure, T the temperature, C the fraction of hydrated
 88 rocks (0 being only ambient mantle rocks and 1 being only hydrated rocks). The non-
 89 dimensional viscosity is η , H is the nondimensional heat production rate and Ra is the
 90 Rayleigh number based on the temperature drop between the base and the top of the model
 91 and B_c the buoyancy ratio defined below. We employ pseudo-plasticity where the viscosity is
 92 defined by:

93
$$\eta(T, p) = \eta_A(p) \exp\left(\frac{E_a + pV_a}{T}\right) \quad (5)$$

94
$$\sigma_Y = \sigma_0 + d\sigma'_Y \quad (6)$$

95
$$\eta_Y = \frac{\sigma_Y}{2\dot{\epsilon}_{II}} \quad (7)$$

96 where η_A is the non-dimensional reference viscosity at pressure p , including a and step-wise
 97 30-fold jump at 660 km, relevant to Earth (Ricard et al., 1993). E_a is the nondimensional
 98 activation energy, V_a is the nondimensional activation volume, σ_Y is the nondimensional yield
 99 stress (σ_0 being the surface value and σ'_Y its dependence on depth d), η_Y is the non-
 100 dimensional viscosity of yielded rocks and $\dot{\epsilon}_{II}$ the second invariant of the strain rate.

101 The tracer ratio method was used for composition (Tackley and King, 2003). We
 102 monitor the evolution of the composition to evaluate the survival timescale of the hydrated
 103 layer. A stable average composition means no recycling of the hydrated layer, and a
 104 composition of 0 everywhere means complete recycling and vanishing of the hydrated layer.

105 We use a reference parameterization for all convection computations (Table 1), varying
 106 only parameters of the hydrated layer. The reference setting is defined with a Rayleigh

107 number of 10^6 and a non-dimensional surface yield stress of 10^4 . To obtain present-day Earth
108 diagnostics given our formulation and choice of parameters, we would have to increase the
109 Rayleigh number by one order of magnitude (Arnould et al., 2018). Working at lower Ra is
110 easier to explore the parameter space. To transpose our results to dimensional values and
111 Earth’s convective vigor, we scale time so it can be compared to present-day Earth time. We
112 assume that the non-dimensional time-averaged root-mean-square velocity at the surface of
113 the reference model (no hydrated layer) at statistical steady state corresponds to that of the
114 present-day Earth, being 3.9 cm y^{-1} . We call the transit time the time it takes a particle to
115 travel a distance equal to the depth of the mantle at the rms surface velocity (with the value
116 above for the Earth, it corresponds to 75 Myr). This classical choice of scaling gives a
117 reference, but is by no means a precise extrapolation because for a younger Earth, the surface
118 velocity is unknown.

119 The initial thermal condition corresponds to a 1D geotherm built from running the
120 convection setup without the hydrated layer, where plate-like tectonic regime emerges and is
121 sustained when the thermal statistically steady state is reached. The initial condition is
122 therefore gravitationally unstable and destabilizes by recycling the surface boundary layer. A
123 lateral harmonic perturbation is introduced to initiate the process.

124 Over this initial purely thermal setting we superimpose a hydrated layer. We study the
125 efficiency of its recycling as a function of its density and thickness. We explore the role of
126 material properties of hydrated rocks, having lower density and softer rheology than the
127 ambient mantle. For the hydrated layer, we use a reference η_A and σ_Y 10 times lower than
128 those of ambient mantle (Fig. 1). The hydrated layer is viscous close to the surface because it
129 is cold, except in areas where the maximum stress is reached and shear zones may develop.
130 The buoyancy number of the hydrated crust B_c is defined as $\Delta\rho_c/\Delta\rho_T$, where $\Delta\rho_c$ and $\Delta\rho_T$ are
131 the density contrasts between hydrated layer and ambient mantle, and between the hottest and

132 coldest ambient mantle, respectively. B_c was set to 0, -0.034, -0.17, -1.7 and -3.4,
133 corresponding for Earth to dimensional values of the density of the hydrated "crust" of 4400,
134 4355, 4310, 3510 and 2620 kg m⁻³. The latter value is the nominal value of serpentine density
135 in the high temperature antigorite form (Hilaireret et al., 2006), for which the density jump with
136 the incompressible mantle of average density of 4400 kg m⁻³ is 1780 kg m⁻³, about twice the
137 actual density jump between serpentine and dry peridotite. Models with this density jump are
138 therefore likely unrealistic but were nonetheless performed to explore the stability of the
139 convection regimes.

140 "Hydration rate" is defined as 100% for this extreme case, and would take values of 50,
141 5 and 1% for decreasing values of B_c . Models with half this density jump ($B_c = -1.7$)
142 correspond to realistic values for a fully serpentinized upper layer. Models with $B_c = -0.034$,
143 and -0.17 would represent a density jump if the crust is partially serpentinized to 2 and 10 %,
144 respectively, for examples along deep discrete fractures such as those generated by thermal
145 cracking (Korenaga, 2007). Thus serpentinization is more useful for the discussion, and is
146 twice the "hydration rate" defined above. Serpentine contains about 12 wt% H₂O, from which
147 the average water concentration in the simulation is obtained.

148 The thickness of the hydrated layer was set to 10, 50, 100 and 200 km (0.0034, 0.017,
149 0.034 and 0.068 in non-dimensional values). Calculations are named by the series of numbers
150 indicating starting dimensional parameters as thickness (km)-hydration rate-simulation
151 number (Fig. 2, Table 2).

152

153 **3. Results**

154 We monitored the decrease of the fraction of hydrated material in the system. Two
155 reference solutions of entrainment are calculated before exploring the role of hydrated rocks
156 material properties. These solutions correspond to the entrainment of a layer having the same

157 material properties as the ambient mantle, solutions Ref-1 and Ref-2 having 50 and a 200 km-
158 thick layers, respectively (Fig. 2a). The flow is identical in the two reference solutions, and
159 the top layer is quickly recycled by cold downwellings. The first subduction-like downwelling
160 (*i.e.* asymmetric to one-sided, continuous and stiff) initiates after about three transit times.
161 The 50 km-thick layer disappears within less than the duration of one transit time. In Ref-1,
162 small-scale convection or hot flow near the upper thermal boundary erodes slightly the deeper
163 "crust" before full model convection initiation (Fig. 2a). Recycling follows an exponential
164 decay with characteristic recycling half-life τ of less than 10 Myr, about one order of
165 magnitude lower than the transit time (Table 2).

166 Numerical solutions with a hydrated layer display three distinct regimes of hydrated
167 crust preservation depending on its initial thickness and density defined by the degree of
168 serpentinization (Figs. 2 and 3).

169 Regime I is observed for small density jumps between crust and mantle (2-10 %
170 serpentinization), where numerical solutions show a similar behavior to the reference
171 solutions (Fig. 2b). Temperature and viscosity fields show hot upwellings reaching the
172 surface, and cold downwellings penetrating the mantle that set up the global scale convection.
173 In that regime, the hydrated layer is entrained as a passive layer in the global flow pattern
174 because the integrated buoyancy of the top boundary layer is negative at all times. Once
175 downwelling initiates, the exponential decay time ($\tau \sim 10-15$ Myr) is similar to that of the
176 reference models, one order of magnitude lower than the transit time. For small thickness
177 (< 50 km) of the hydrated crust, downwelling initiation occurs after about three transit times
178 (~ 200 Myr). Above 50 km thickness, downwelling initiates even faster because the hydrated
179 layer is weaker than the ambient mantle, favoring yielding and forming weaker slabs that sink
180 with less resistance.

181 Regime II is a domain corresponding to large hydration degree (50-100%) where the
182 hydrated crust is slowly recycled (Table 2, Fig. 2c). The time of total disappearance of the
183 hydrated layer is longer than the overturn time in reference calculations. Some calculations do
184 not show complete recycling of the layer even after statistical steady state is reached. This
185 dynamic regime is sustained because the hydrated layer brings the surface boundary layer
186 close to neutral buoyancy. Temperature and viscosity fields show hot upwellings reaching the
187 surface and cold downwellings penetrating the mantle (Fig. 2c). Where hot upwellings reach
188 the surface, the hydrated layer yields, thins and stretches, ultimately forming buoyant rafts
189 that drift. Above downwellings, the hydrated layer thickens and its base flows down into the
190 mantle dragged by temporary flat and cold asymmetric slabs. Regime II corresponds to fully
191 serpentinized crust of small to moderate thickness (Fig. 3), buoyant enough to cause recycling
192 rate to be close to linear rather than exponential behavior of regime I, progressing at slower
193 and regular pace. The constant flux of recycling does not depend on the volume of hydrated
194 material that remains at the surface, and the entrainment characteristic time of hydrated crust
195 τ is proportional to the percentage of hydration (Fig. 4). It means that in this regime, flat and
196 cold downwellings erode regularly the hydrated material but do not entrain the full layer in
197 subduction. Numerical solutions in which the rheology of the hydrated layer is identical to
198 that of ambient mantle present the same characteristics as the solution with a viscosity
199 contrast of 0.1, showing that the occurrence of regime II is controlled to first order by
200 buoyancy, not by rheological contrast.

201 Two solutions fall in a third regime that displays no recycling even after statistical
202 steady state is reached (Table 2, Fig. 2d). The corresponding parameter subspace is that of
203 thick (≥ 100 km) and extremely buoyant hydrated layer (100% hydration), imposing a
204 stagnant lid tectonic regime with convective deformation within the hydrated layer decoupled
205 from the convection within the ambient mantle. The transition between the mixing regimes

206 (Regime I and II) and the stable layer regime for a light surficial layer (Regime III) is
207 symmetrical to the one observed in convection with a dense basal layer (Le Bars and Davaille,
208 2004). Regime III is unlikely on any terrestrial planet because of the excessive density
209 contrast between hydrated crust and mantle.

210

211 **4. Discussion and conclusions**

212 Mantle convection models including a serpentinized primordial crust show the existence
213 of 3 regimes of mixing/stability of such a light layer: Regime I being rapid exponential
214 vanishing for low hydration degree, Regime II slow linear vanishing for efficient hydration
215 over a moderate thickness, and Regime III survival for excessively light and thick crust.

216 Current estimates of the water content of the Earth (Marty, 2012; Piani et al., 2020) lie
217 between 1 and 10 oceanic masses, corresponding to 350-3500 ppm in the bulk silicate Earth
218 and to a 6.5-55 km-thick serpentine layer (12 wt% H₂O). Thus a primordial fully
219 serpentinized crust may have reached about 50 km on the early Earth, and regime II was
220 possible. A 50-km thick serpentine crust may have survived over several overturns
221 (corresponding here to ~1 Gyr, Fig. 4). Such a thick serpentinized crust provides the ideal
222 context of formation of a differentiated felsic crust at shallow depths (Borisova et al., 2021)
223 over the first few hundred millions of the Earth history. Coexistence of TTG-type felsic crust
224 and serpentinized ultramafic rocks would account for geochemical evidence for both 4.3 Ga
225 differentiated crust recycling in Hadean Jack Hills zircons (Blichert-Toft and Albarède,
226 2008), and survival time of 1 Gy of a 50-km thick serpentinized crust for direct formation of
227 the 4 Ga Acasta gneiss complex rocks from near surface ultramafic-mafic rocks (Reimink et
228 al., 2016).

229 The present models could be improved since the setup was simplified in order to
230 pinpoint physical behavior and for computational cost considerations. Firstly, the models are

231 2D, not accounting for 3D features like toroidal flow or cylindrical plumes. However the
232 defined regimes are based on average properties that are usually similar in 2D and 3D
233 convective systems. The Boussinesq approximation neglects the compressibility of rocks and
234 the effects of phase changes. Since the explored regimes of recycling occur mostly in the
235 upper mantle, the results should arguably not be strongly affected. Due to calculation time,
236 the convective vigor of our models is lower than what is expected for the Earth. Scaling was
237 thus based on surface velocities and characteristic transit/overturn times. Using the scaling of
238 convective velocities with $Ra^{2/3}$, our results could be extended to cover ranges of young
239 Earth-like values, potentially reducing the stability dimensional times by a factor of 3-10.
240 However, models of a hot planet with large magma extraction show that the scaling of
241 convection velocities with Ra does not hold, velocities remaining stable instead (Xie and
242 Tackley, 2004). Exploring hydrated crust evolution with melt abundant mantle convection
243 model and continent formation would be the next step to make.

244

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250

251

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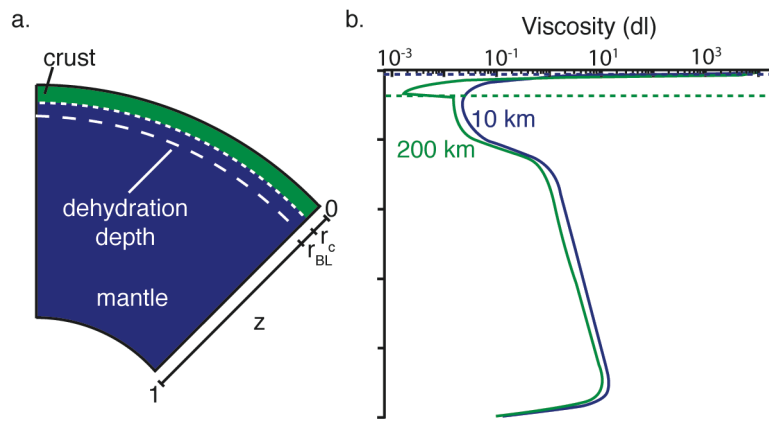
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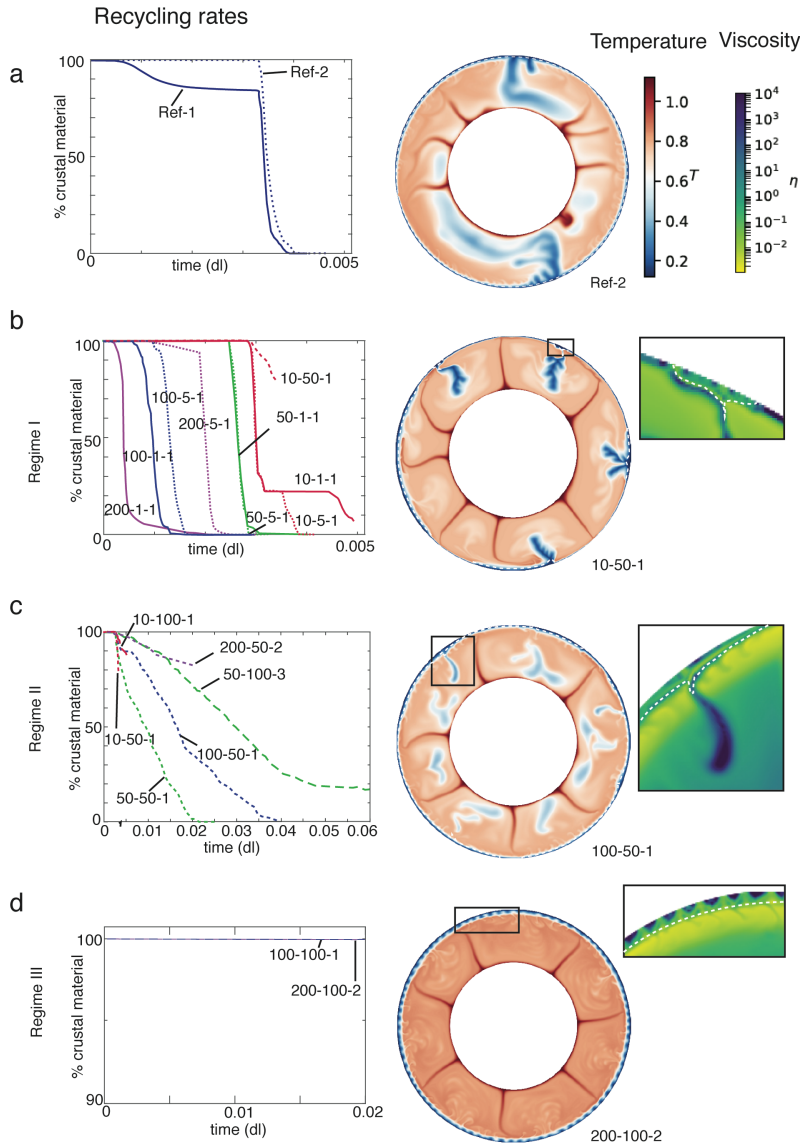


335

336 Figure 1: Initial set up. a. Schematic representation of the spherical annulus geometry with
 337 two layers and definition of the different thicknesses. z is the depth of the mantle; r_c is the
 338 hydrated layer thickness; r_{BL} is the thickness of the top thermal boundary layer (BL). b.
 339 Viscosity profile for a fully hydrated crust thickness of 200 km and 10 km are compared. dl:
 340 dimensionless.

341

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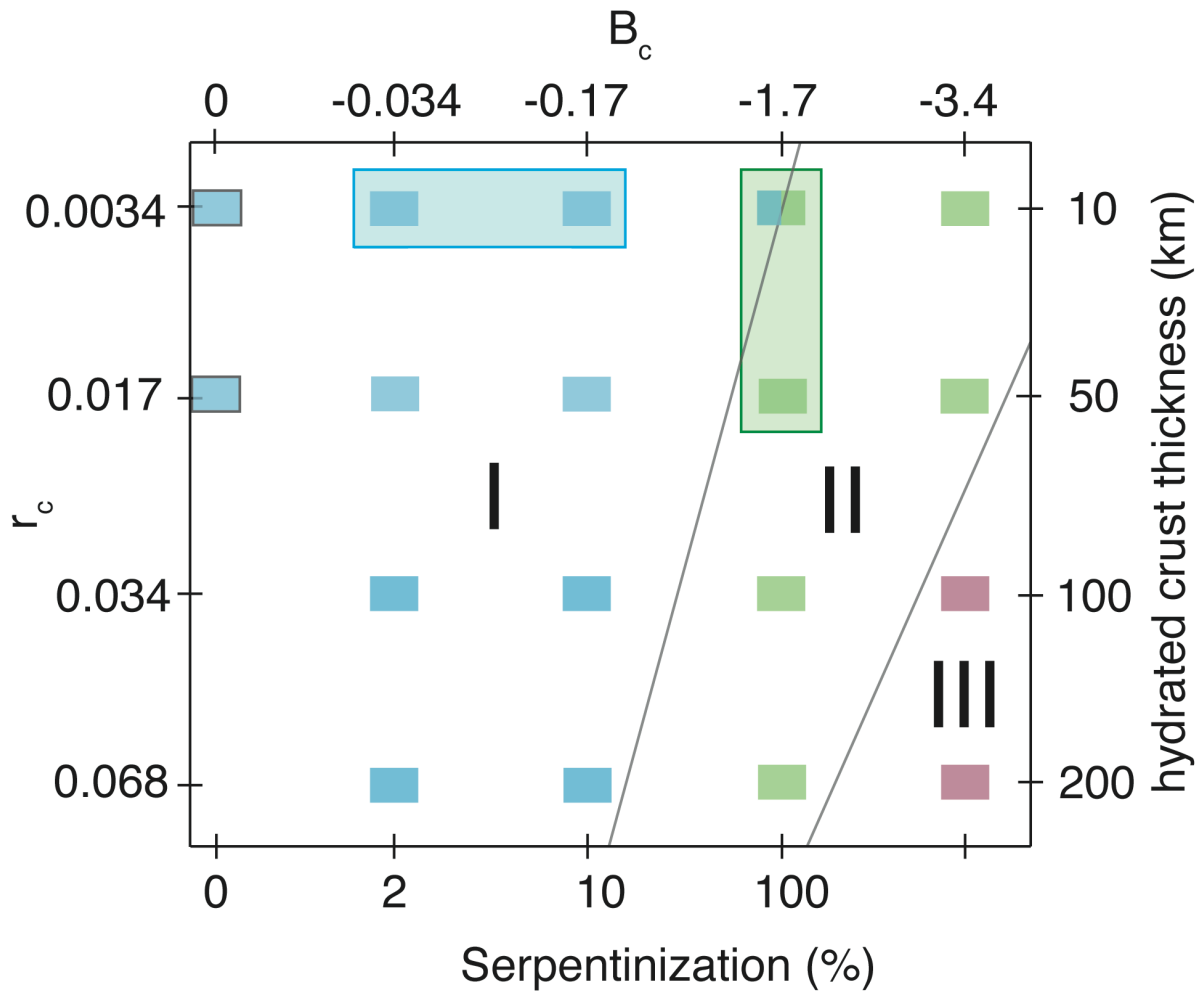
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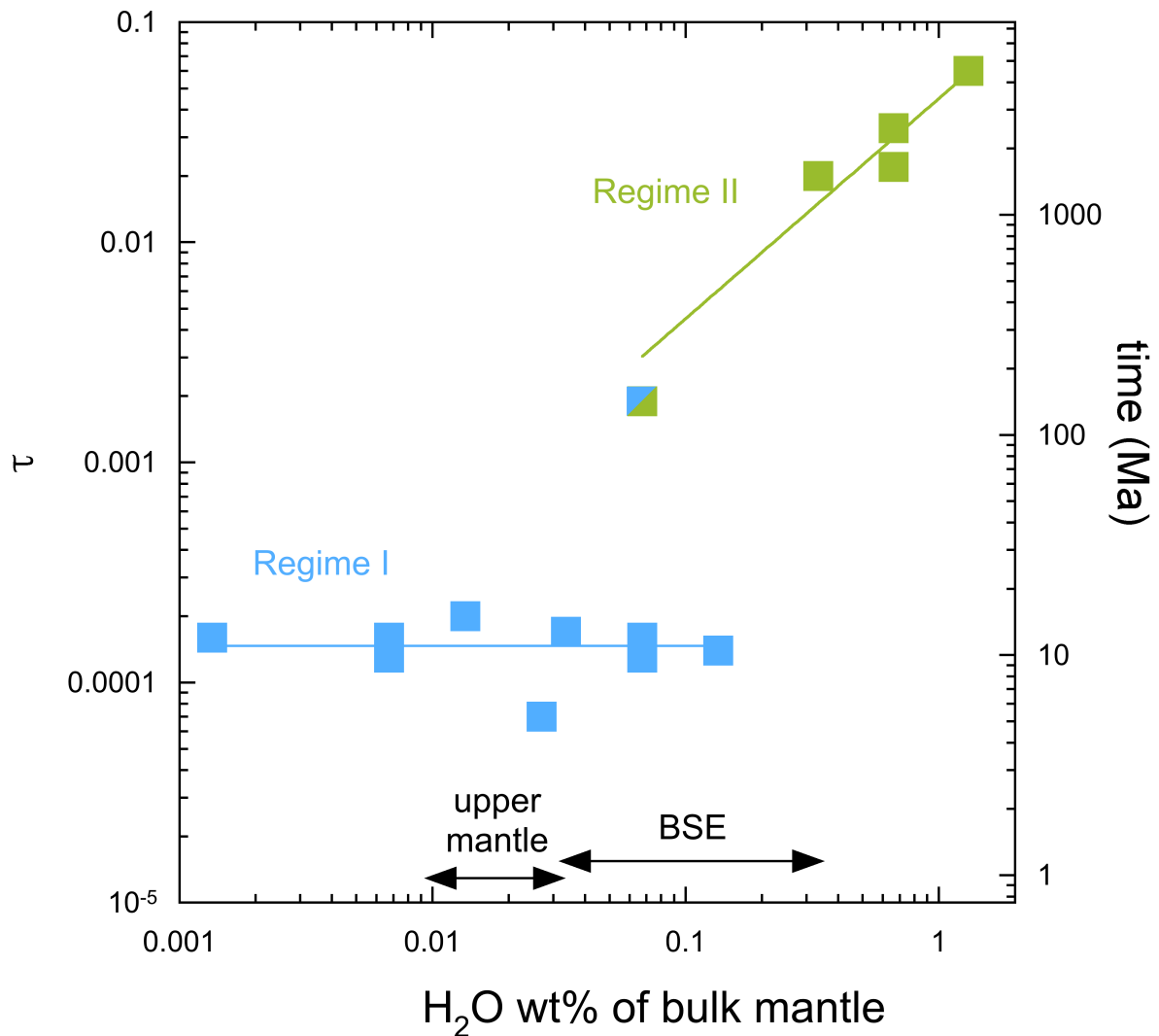
Figure 2: Characteristic recycling rates and flow patterns of the reference solutions (a) and of the three regimes of recycling/preservation of the hydrated layer (b,c,d). Surface recycling is similar in both simulations where the passive markers of “crustal” material have the same properties as the mantle. A dimensionless time of 0.001 corresponds to ~ 75 Myr for present Earth conditions, and the characteristic overturn time is ~ 300 Myr. Temperature snapshots are shown after downwelling/subduction initiation. Viscosity at the top of the mantle is shown for selected areas to illustrate tectonic features of the three regimes with hydrated crust. Temperature fields were obtained for two reference models with markers implemented over the first 200 km (Ref-1) and over the first 50 km (Ref-2).



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354 Figure 3: Crust recycling regimes as a function of the crust thickness and buoyancy contrast
 355 (r_c : dimensionless crust thickness. B_c : dimensionless buoyancy number of the crust). Regime I
 356 (blue) groups numerical simulations that share the same characteristics as the reference
 357 simulations with fast disappearance of the crust due to direct entrainment by downwelling,
 358 regime II (green) simulations where hydrate crust is eroded by downwelling and its
 359 disappearance time depends on crust thickness, and regime III (pink) where the hydrated crust
 360 is preserved. Typical thickness and degree of serpentinization are reported for present-day
 361 Earth (blue box) and Early Earth (green box). Note that the density contrast for the last
 362 column is not compatible with known degrees of serpentinization (see text).

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365 Figure 4. Geodynamical models are similar to present-day Earth regime (I) at low water
 366 content with dimensionless recycling time τ of the hydrated crust similar to that of the
 367 reference models ($\tau \sim 0.00015$). Models in regime II at high water content show a linear
 368 relationship between the time of recycling and the thickness of the hydrated crust converted as
 369 a percentage of H_2O in the mantle. The thick and buoyant serpentized crust can survive as
 370 long as one Gyr for the early Earth for the highest values of geochemical estimates of water
 371 content for the bulk silicate Earth (BSE). Scaled times are shown on the right vertical axis.

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373 Table 1. Model parameters.

Parameter	Symbol	Non-dimensional value	Dimensional value
Rayleigh number	Ra	10^6	
Initial quantity of heat	Rh	20	$3.7 \times 10^{-12} \text{ W kg}^{-1}$
Basal temperature of the upper boundary layer (temps=0)	T_{BL}	0.72	1900 K
Temperature difference (Bottom –Top)	ΔT	1	2639 K
Reference density	ρ_m	1	4400 kg.m^{-3}
Thermal expansivity	α	1	$4.5 \times 10^{-5} \text{ K}^{-1}$
Thermal diffusivity	κ	1	$10^{-6} \text{ m}^2.\text{s}^{-1}$
Thermal conductivity		1	$4 \text{ W.m}^{-1}.\text{K}^{-1}$
Reference viscosity	η_m	1	10^{23} Pa.s
Viscosity jump factor at 660 km		30	
Activation energy for viscosity	Ea	8	142 kJ.mol^{-1}
Yield stress at the surface		10^4	118 MPa
Hydrated crust viscosity and yielding factor		0.1	
Thickness of the model	r	1	2900 km
Thickness of the upper boundary layer	r_{BL}	0.07 – 0.09	203 – 261 km
Thickness of the crust	r_c	0 – 0.068	0 – 200 km
Buoyancy number for the crust	B_c	-3.4 – 0	

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378 Table 2: Inputs and outputs of the models. r_{BL} and r_c : thickness of the final top boundary layer
 379 and of the original crust, respectively. B_c and B_{bl} : initial buoyancy number of the crust. τ :
 380 characteristic recycling time.

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Name	Buoyant crust	r_{BL}	r_c	B_c	τ
Ref-1	No	0.09	0.068	0	0.00010
Ref-2	No	0.09	0.017	0	0.00012
200-100-2	Yes	0.08	0.068	-3.4	∞
200-50-2	Yes	0.08	0.068	-1.7	0.06
200-5-1	Yes	0.08	0.068	-0.17	0.00014
200-1-1	Yes	0.08	0.068	-0.03	0.00007
100-100-1	Yes	0.09	0.03	-3.4	∞
100-50-1	Yes	0.09	0.03	-1.7	0.033
100-50-2*	Yes	0.09	0.03	-1.7	0.022
100-5-1	Yes	0.09	0.03	-0.17	0.00016
100-5-2*	Yes	0.09	0.03	-0.17	0.00013
100-1-1	Yes	0.09	0.03	-0.03	0.0002
50-100-3	Yes	0.09	0.017	-3.4	0.054
50-50-1	Yes	0.09	0.017	-1.7	0.02
50-5-1	Yes	0.09	0.017	-0.17	0.00017
50-1-1	Yes	0.09	0.017	-0.03	0.00016
10-100-2	Yes	0.07	0.003	-3.4	0.022
10-50-1 [§]	Yes	0.07	0.003	-1.7	0.0019
10-5-1	Yes	0.07	0.003	-0.17	0.00013
10-1-1	Yes	0.07	0.003	-0.03	0.00016

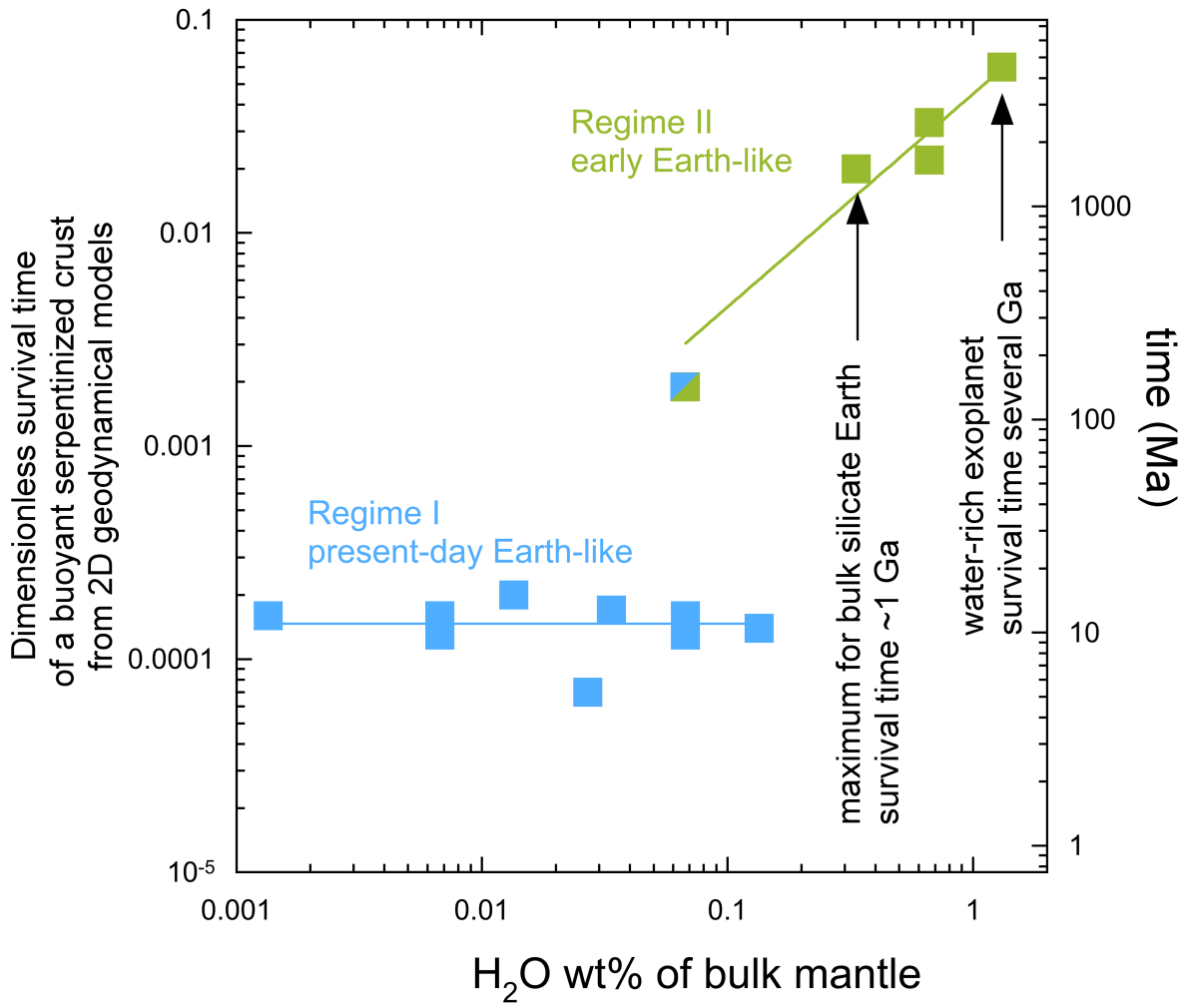
382 * simulations with no rheological contrast between the hydrated crust and mantle.

383 [§] calculation results with mixed first and second regime character.

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Graphical abstract



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