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Small-scale gravitational instabilities under the oceans: implications for the evolution of oceanic lithosphere and its expression in geophysical observables

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(31 Jan 2008)

Sublithospheric small-scale convection (SSC) is thought to be responsible for the flattening of the seafloor depth and surface heat flow observed in mature plates. Although the existence of SSC is generally accepted, its ability to effectively produce a constant lithospheric thickness (i.e. flattening of observables) is a matter of debate.

Here we study the development and evolution of SSC with a 2D thermomechanical finite-element code. Emphasis is put on i) the influence of various rheological and thermophysical parameters on SSC, and ii) its ability to reproduce geophysical observables (i.e. seafloor depth, surface heat flow, and seismic velocities). We find that shear heating plays no significant role either in the onset of SSC or in reducing the lithospheric thickness. In contrast, radiogenic heat sources and adiabatic heating exert a major control on both the vigour of SSC and the thermal structure of the lithosphere. We find that either dislocation creep, diffusion creep, or a combination of these mechanism, can generate SSC with rheological parameters given by laboratory experiments. However, vigorous SSC and significant lithospheric erosion are only possible for relatively low activation energies. Well-developed SSC occurs only if the first ~ 300 km of the mantle has an average viscosity of $\lesssim 10^{20}$ Pa s; higher values suppress SSC, while lower values generates unrealistic high velocities. Seismic structures predicted by our models resemble closely tomography studies in oceanic mantle. However, the fitting to observed

Seismic structures predicted by our models resemble closely tomography studies in oceanic mantle. However, the fitting to observed seafloor topography and surface heat flow is still unsatisfactory. This puts forward a fundamental dichotomy between the two datasets. This can be reconciled if most of the observed flattening in seafloor topography is influenced by processes other than SSC.

1 Introduction

Oceanic lithosphere is being continuously created at mid-ocean ridges, where adjacent plates move apart from each other in a process called seafloor spreading [1]. As these plates diverge, hot mantle rocks ascend to fill the gap. Upon subsequent conductive cooling, these rocks become rigid and form new oceanic lithosphere. The complementary process of plate consumption occurs along subduction zones, where plates bend and descend into the Earth's mantle. The entire process of creation, lateral displacement, and eventual subduction of oceanic lithosphere can be thought of as a large scale convection cell, where the oceanic lithosphere represents the upper thermal boundary layer.

As oceanic plates move away from ridges, they cool from above, thicken, and become denser by thermal contraction. This cooling is reflected on the dependence of geophysical observables on the age of the plate t [2, 3]. For plates younger than about 70 My, both sea floor topography and surface heat flow (SHF) decrease linearly as \sqrt{t} , consistent with predictions from the *half-space cooling model* [4].

For larger ages, however, this relation breaks down and the two observables decrease less rapidly, reaching almost constant values in ocean basins [3,5,6]. Since these observables reflect the thermal structure of the lithosphere, their flattening implies a similar behaviour for the isotherms within the plate. These features are included in the popular *plate model* [2], which considers the lithosphere as a cooling plate with an isothermal lower boundary. Although this model can explain the observed flattening of both sea floor topography and SHF, it does not propose any particular mechanism by which the horizontal isotherm is maintained at constant depth.

One such mechanism, known as small-scale convection (SSC) [7], involves the generation of thermal instabilities at the lower parts of the lithosphere. Conductive cooling cause the isotherms to migrate downwards until the cold material at the base of the lithosphere becomes gravitationally unstable and

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43 SSC is triggered. This mechanism provides the necessary extra heat to keep a horizontal and isothermal
44 lithospheric base by replacing cold (dense) material with hotter mantle. The resulting thermal erosion
45 would maintain a quasi-constant lithospheric thickness through time.

Small-scale convection has been studied using theoretical (e.g. [7–9]), numerical (e.g. [10–16]) and analogue models (e.g. [17]). These studies have been focused on the general conditions for the existence of SSC, but non attempted a systematic exploration of the effects of all relevant physical parameters on SSC. Moreover, either seismically derived thermal structures or SHF and sea floor topography data have been used to test the reliability of the results, but no study has combined these three observables into a single consistent model. This is of particular relevance, because seismic observations seem to favour half-space cooling models over plate models, while SHF and sea floor topography observations suggest the opposite.

This work presents a systematic study on the influence of several rheological and thermophysical parameters on SSC, its expression in geophysical observables, and its role in determining the thickness of oceanic lithosphere. Predictions of seismic velocities, SHF and sea floor topography are used to ensure compatibility with current observations. In the following sections we first introduce the statement of the problem and the applied numerical methods; we then describe the approach to calculate the relevant geophysical observables; finally, we discuss the results and their implications on the evolution of the oceanic lithosphere.

60 2 Model description

61 2.1 Governing equations

The Earth's mantle behaves as a highly viscous fluid over the long time scale $(t > 10^4 \text{yr})$ (cf. [1, 18]). Since the physical properties of this fluid are strongly dependent on temperature, the physical model involves a mechanical flow problem coupled to a thermal problem. We consider an incompressible fluid in a rectangular domain. Due to the almost infinite Prandtl number of the fluid, inertial terms are neglected and the problem becomes quasi-static. The transient character of the solution is due to the evolution of the temperature field. Under the Boussinesq approximation (i.e. the effects of density variations other than in the body-force terms are neglected), the three unknowns, velocity \mathbf{u} , the pressure P and the temperature T are determined by solving the conservation of momentum (Stokes equation), mass, and energy equations (cf. [1]):

$$\nabla \cdot (\eta \nabla^{\mathbf{s}} \mathbf{u}) + \nabla P = \rho \mathbf{g} \tag{1}$$

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

$$\rho C_p \left(\frac{\partial T}{\partial t} + \mathbf{u} \nabla T\right) = \nabla \cdot (k \nabla T) + \rho f \tag{3}$$

where the operator $\nabla^{\rm s}$ is the symmetrized gradient, namely $1/2(\nabla^{\rm T}+\nabla)$, η is the viscosity, ρ the density, g the gravitational acceleration vector, C_p the isobaric heat capacity, k the thermal conductivity, and f a heat source term which is the sum of: i) a constant term f_r corresponding to the decay of radioactive elements, ii) an adiabatic heating term $f_{ah} \approx T \alpha \rho u_z g_z$, where α denotes the coefficient of thermal expansion and subscript z refers to the vertical component of the vectors, and iii) a shear heating term associated with mechanical heat dissipation. The shear heating is computed from eqs. (1) and (2) as $f_{sh} = \sigma_{ij} \dot{\varepsilon}_{ij}$, where σ and $\dot{\varepsilon} = \nabla^{s} \mathbf{u}$ are the stress and strain rate tensors. As the constitutive equation described in the next section depends on the velocity, the system is highly non-linear.

The mechanical problem (eqs. (1) and (2)) is solved in an Eulerian framework with the finite element method using a mixed formulation in terms of velocity and pressure. We use the well-known triangular mini element $5(P1^+-P1)$ [19], with four nodes for the velocity (three at the vertices with linear shape functions and one at the centre with a cubic bubble function) and three pressure nodes (piecewise linear interpolation). The mini element exhibits linear convergence. It has been reported [19] that the accuracy

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of pressure approximations provided by the mini element is degraded in 3D with respect to the more accurate and computationally demanding Taylor-Hood element (P2-P1). Nevertheless, in the 2D examples presented in this paper the pressure is almost linear and it is fairly approximated using the mini element with a minimum computational cost.

A basic Picard method is used to handle the non-linear character introduced by the constitutive equation. In most cases, this simple method converge to the desired accuracy in a reduced number of iterations. Due to the quasi-static character of the mechanical model, the time evolution is determined solving a series of steady state problems. On the other hand, the thermal model (eq. (3)), has to be integrated along time using a time stepping strategy. This requires a space-time discretization. The space discretization is performed using standard linear finite elements with a Galerkin formulation, while the time discretization uses an explicit fourth-order Padé method [19], which properly accounts for both the advective and the diffusive part of the equation. The explicit nature of the time stepping scheme used for the thermal model together with the quasi-static character of the mechanical problem allow solving the thermo-mechanical coupling using a staggered scheme. The explicit scheme for the thermal problem is conditionally stable. The time step must be sufficiently small to fulfill the stability requirements established by limiting the Courant number. This small time step guarantees the stability of the thermal scheme and, hence, of the fully coupled thermo-mechanical model.

The models we are using consider only one type of material (with characteristics associated with the mantle; the oceanic crust is not included in the modeling). Moreover, all physical properties (viscosity, thermal conductivity, density...) are computed as explicit functions of T, P, \mathbf{u} . Consequently, there is no need of tracking the Lagrangian motion of the material. The reader is referred to [20] for examples of Lagrangian tracking of different materials using a level-set approach.

Non-linearity in eq. (3) arises due to the dependency of density and thermal conductivity on temperature. It is assumed that within a time step the material properties (ρ and k) can be approximated by their values at the beginning of the time step. This simplifies the implementation and reduces the computational cost. A detailed description of the numerical strategy used here to solve the thermo-mechanical coupled problem is given in [20].

111 2.2 Constitutive equation

Convective flow in the Earth's mantle is possibly due to the high-temperature creep of mantle rocks. This solid-state deformation mechanism occurs due to the thermally activated motion of atoms associated with lattice defects such as dislocations and vacancies (cf. [21]). There is general agreement that two main creep mechanism are likely responsible for most of the deformation in the mantle: diffusion creep (Herring-Nabarro and Coble creep) and dislocation creep [21,22]. Although there are significant uncertain-ties associated with the extrapolation of laboratory results (performed at low pressures and high strains rates) to mantle conditions, a comparison of microstructures on experimentally and naturally deformed peridotites indicates that the same deformation mechanisms detected in laboratory take place in the mantle as well [21,23]. Deformation caused by dislocation creep is evidenced in lithospheric mantle samples (e.g. xenoliths, peridotitic massifs) and indirectly inferred in the shallow upper mantle from seismic anisotropy studies (see [24] for a recent review). On the other hand, diffusion creep may be dominant over dislocation creep at depths > 250-300 km, where stresses are low and pressure effects become dominant (i.e. the activation volume of diffusion creep seems to be smaller than that of dislocation creep, [21]). This change in deformation mechanism with depth is consistent with the lack of significant anisotropy at such depths, although not conclusive [25].

Theoretical treatments and experimental observations demonstrate that the macroscopic creep behaviour of rocks is well described using a "power-law" of the form [21, 26, 27]

$$\dot{\varepsilon} = A(\sigma'/\mu)^n (b/d)^m \exp\left(-\frac{E+PV}{RT}\right)$$
(4)

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Table 1.	Phase	transition	parameters.	
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Depth(km)	$slope(MPa K^{-1})$	T(K)	P(MPa)	$\Delta \rho \; (\mathrm{kg \; m^{-3}})$
$410^a \\ 510^b \\ 660^c$	4.0	1600	14200	250
	4.0	1700	17000	0
	-2.5	1873	23100	250

^{*a*} from [29], ^{*b*} from [30], ^{*c*} from [31]

where d is the average grain-size, σ' the deviatoric stress, A the pre-exponential factor, μ the shear modulus, b the length of the Burgers vector, n the stress exponent, m the grain-size exponent, E the activation energy, V the activation volume, and R the gas constant (see Table 4). The combination of eq. (4) with the definition of viscosity ($\eta = \frac{1}{2}\sigma'/\dot{\varepsilon}$), allows isolating an explicit expression for η in terms of T, P and $\dot{\varepsilon}$. This expression is then used to solve eq. (1).

To compute the viscosity we assume a constant material parameter $A_D = \frac{1}{2} A^{-1/n} \mu^{-1} (b/d)^{-m/n}$, including the pre-exponential factor A, the grain-size dependence, and the shear modulus. Although grain-size may change due to grain growth and dynamic recrystallization processes, its dependence on stress is not well known. We simplify the rheology here by assuming a constant grain size.

Diffusion and dislocation creep act simultaneously in the mantle [21]. In order to account for the effect of the two mechanisms, two different viscosities η_{diff} and η_{disl} are computed separately and then combined into an effective viscosity η_{eff} , which is computed as the harmonic mean of η_{diff} and η_{disl} :

$$\frac{1}{\eta_{\text{eff}}} = \left(\frac{1}{\eta_{\text{diff}}} + \frac{1}{\eta_{\text{disl}}}\right). \tag{5}$$

This expression is truncated if the resulting viscosity is either greater or lower than two imposed cutoff values (10¹⁸ to 10²⁴ Pa s). The viscosity η_{diff} is computed using n = 1 while for η_{disl} we use m = 0. The values of the rest of the parameters are described section 3.

144 2.3 Phase transitions and mineral domains

At least four main solid-solid mineral phase transitions occur in the mantle region considered in this study: plagioclase-spinel, spinel-garnet, olivine-wadsleyite, and wadsleyite-ringwoodite. Other phase transitions (e.g. orthoenstatite to clinoenstatite) may occur within the domain, but their effect on the type of gravi-tational instabilities of interest are negligible. Here we consider explicitly only the olivine-wadsleyite, and wadslayite-ringwoodite phase changes, which are the most relevant in terms of density and viscosity con-trasts that may exert a control on the vertical structure of SSC. Each of these phase transitions is character-ized by a particular Clapevron slope, which we approximate as linear functions in the temperature-pressure domain (See Table 1). The olivine-wadsleyite and wadsleite-ringwoodite transitions occur at ~ 410 and 510km depth, respectively, in a pyrolitic adiabatic mantle. In the oceanic mantle, the plagioclase-spinel and spinel-garnet transitions occur at depths of ~ 30 and 50-80 km, respectively [28]. The depth variability in the latter is due to the exothermic nature of the reaction and the rapid horizontal temperature variation in the shallow oceanic upper mantle. In principle, the spinel-garnet phase change occurs deep enough to be affected by SSC, and given its exothermic nature and associated density change (0.8-1.0%, [28]), it could promote SSC through buoyancy enhancement (cf. [1]). However, the present study is focused on the role that other major physical parameters play on the development of SSC, and therefore we leave petrological and phase change effects for a future study (work in progress).

Representative reference properties for the whole mantle are estimated considering the stable phases at different temperatures and pressures. Therefore, the average rock properties are computed as follow: i) the volumetric fractions of the major constituent phases along a 1600 K adiabat are taken from [32] (See Table 2), ii) each phase is identified with only one (the most abundant) end-member (e.g. enstatite for opx, diopside for cpx, pyrope for grt), iii) experimentally derived properties of these end-members are taken from the references listed in Tables 1 to 3, and finally iii) the average rock properties are computed as either the arithmetic mean of the end-members weighted by their respective volumetric proportions or with a

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Table 2. Stable phases at different depths. After [32].

Mineral	200 km	$420~\mathrm{km}$	600 km	$800 \mathrm{km}$
Olivine(Fo)	51.3%	0	0	0
Olivine(Fa)	5.7%	0	0	0
Orthopyroxene(Enstatite)	13.5%	0	0	0
Clinopyroxene(Diopside)	10.0%	0	0	0
Garnet (Pyrope)	19.6%	40.0%	37.5%	0
Wadsleyite	0	60.0%	0	0
Ringwoodite	0	0	60.0%	0
Ferropericlase	0	0	0	16.0%
Mg perovskite	0	0	0	78.0%
Ca perovskite	0	0	2.5%	6.0%

Table 3. Parameters for thermal expansivity α .

Param.	$200 \mathrm{~km}$	420 km	$600 \mathrm{km}$
$ \begin{array}{c} a_0(\times 10^{-5}) \\ a_1(\times 10^{-9}) \\ a_2 \\ a_3 \end{array} $	2.65032	2.43507	2.30551
	9.19917	2.97727	2.79119
	-0.27127735	-0.184206667	-0.17269375
	0	0	0

Parameters for the different phases were taken from: Forsterite, Fayalite, Enstatite, Diopside, Pyrope: [28]; Wadsleyite, Ringwoodite: [35]; Ferropericlase, Ca-Perovskite: [36]; Mg-Perovskite: [37].

Voigt-Reuss-Hill averaging scheme. The latter is used only when calculating the elastic moduli for seismic velocities. We acknowledge that this approach is only valid to the first-order and lack thermodynamic consistency. However, it gives values comparable, within one standard deviation, to those obtained with more sophisticated methods (e.g. [32, 33]) with a minimum of computational time.

¹⁷² Coefficient of thermal expansion. The thermal dependence of the coefficient of thermal expansion at ¹⁷³ pressure P_0 is approximated by a polynomial expression of the form (e.g. [28,34])

$$\alpha(P_0, T) = a_0 + a_1 T + a_2 T^{-2} + a_3 T^4 \tag{6}$$

¹⁷⁴ The averaged coefficients for each stability field are listed in Table 3.

The pressure effect on the coefficient of thermal expansion can be described by the Anderson-Grüneisen parameter δ [38, 39] as

$$\alpha(P,T) = \alpha(P_0,T) \left(\frac{\rho(P,T)}{\rho(P_0,T)}\right)^{\delta\frac{\rho(P_0,T)}{\rho(P,T)}} \approx \alpha(P_0,T) \left(1 + \beta(P - P_0)\right)^{\delta(1 + \beta(P - P_0))}$$
(7)

The approximation in eq. (7) provides an explicit formula to compute the thermal expansion coefficient in terms of pressure and temperature.

Density. Density changes associated with temperature and pressure variations accompanying convection are small compared to the spherically averaged density of the mantle. Therefore, it is appropriate to simplify the density ρ as a linear function of temperature and pressure with respect to a reference value ρ_0 (calculated in a previous step as described above). Our simplified equation of state is of the form

$$\rho(P,T) = \rho_0 \left[1 - \alpha(P,T) \times (T - T_0) \right] \times \left[1 + \beta \times (P - P_0) \right]$$
(8)

where β is the compressibility, and T_0 and P_0 are, respectively, the temperature and pressure at which the reference density ρ_0 is given.

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Table 4. Physical and geometrical model parameters

Symbol	Meaning	Value used	Dimension
Н	model height	660	km
W	model width	7000 or 10000	km
g	gravity acceleration vector	[0, -9.8]	${\rm m~s^{-1}}$
\overline{R}	gas constant	8.314510	$\rm J~mol^{-1}~K^{-1}$
T_0	reference temperature	273	K
p_0	reference pressure	0.1	MPa
ρ_{ol}	reference density	3300	$\rm kg \ m^{-3}$
β	compressibility coefficient	1×10^{-5}	MPa^{-1}
C_p	thermal Capacity	1200	$J \ kg^{-1} \ K^{-1}$
f_r	radiogenic heat production	2×10^{-8}	$W m^{-1} K^{-1}$
μ	shear modulus	80	GPa
b	length of the Burgers vector	0.5	nm
γ	Grüneisen parameter	1.28	
δ	Anderson-Grüneisen parameter	5.5	
K_0	bulk modulus	120	GPa
K'_0	K_0 derivative with respect to pressure	4.5	
$b_0/b_1/b_2/b_3$	radiation polynomial approx.	$0/0/0/8.5 \times 10^{-11}$	

¹⁸⁵ *Thermal conductivity.* The thermal conductivity is calculated using the model of [40], which gives the ¹⁸⁶ relation

$$k(P,T) = k_{298} \left(\frac{298}{T}\right)^a \exp\left[-\left(4\gamma + \frac{1}{3}\right) \int_{298}^T \alpha(\theta) \,\mathrm{d}\theta\right] \left(1 + \frac{K_0'P}{K_0}\right) + k_{\mathrm{rad}} \tag{9}$$

where k_{298} is the thermal conductivity measured at ambient conditions, a is a parameter with a value ~ 0.33 for silicates, γ the averaged thermal Grüneisen parameter, K_0 is the isothermal bulk modulus, and K'_0 its pressure derivative. The last term $k_{\rm rad}$ is the radiative contribution, approximated by the polynomial function

$$k_{\rm rad} = b_0 + b_1 T + b_2 T^2 + b_3 T^3 \tag{10}$$

where temperature T is in kelvin. Refer to Table 4 for a list of representative parameters.

192 2.4 Geophysical constraints to mantle dynamics

Geophysical observables are commonly used to infer the physical state of the Earth's interior. Available data sets that help to constrain, to different extents, mantle dynamics include ocean floor topography, surface heat flux, seismic velocities, and gravity. As a post-process of our simulations, we estimate these observables. [41]

¹⁹⁷ Sea floor topography. Sea floor topography is estimated assuming local isostasy (i.e. mass per unit area ¹⁹⁸ of a vertical column is compared with respect to a reference value taken at the ridge). Following [42], we ¹⁹⁹ define the isostatic topography w_{iso} as

$$w_{\rm iso} = \int_0^{d_{\rm com}} \left(\rho - \rho_{\rm ref}\right) \,\mathrm{d}z \tag{11}$$

where $d_{\rm com}$ is a compensation depth, and $\rho_{\rm ref}$ the density of a reference column. Although the choice of $d_{\rm com}$ is somewhat arbitrary, it should be taken close to the depth of the deepest isotherm with a dominant conductive component. Isotherms significantly deflected by convection are associated with dynamic loads that are not isostatically compensated (see dynamic topography below).

Dynamic topography. Vertical components of mantle flow may result in a modification of the surface topography. The resulting topography arising from this mechanism is known as dynamic topography, to

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able 5. Solution notation and formulae			
Symbol	Meaning	Value used	
Ol Opx Sp Cpx	olivine orthopyroxene spinel clinopyroxene	$\begin{bmatrix} Mg_x Fe_{1-x} \\ 2SiO_4 \\ Mg_x Fe_{1-x} \\ 2_{-y} Al_{2y} Si_{2-y} O_6 \\ Mg_x Fe_{1-x} Al_2 O_4 \\ Ca_{1-y} Mg_x Fe_{1-x} \\ 1_{+y} Si_2 O_6 \end{bmatrix}$	
Gt	Garnet	Fe ₃ x Ca ₃ y Mg ₃ $(1-x+y+z/3)$ Al ₂ $-2z$ Si _{3+z} O ₁₂ ; $x + y + 4z/3 \le 1$	
C2/c Aki Pv Ppv Ring Wad Wus	pyroxene akimotoite perovskite post-perovskite ringwoodite waddsleyite magnesiowuestite	$ [Mg_xFe_{1-x}]_4\overline{Si}_4O_{12} Mg_xFe_{1-x-y}Al_2ySi_{1-y}O_3, x+y \le 1 Mg_xFe_{1-x-y}Al_2ySi_{1-y}O_3, x+y \le 1 Mg_xFe_{1-x-y}Al_2ySi_{1-y}O_3, x+y \le 1 [Mg_xFe_{1-x}]_2SiO_4 [Mg_xFe_{1-x}]_2SiO_4 Mg_xFe_{1-x}O$	

Unless otherwise noted, the compositional variables x, y, and z may vary between zero and unity and are determined as a function of the computational variables by free-energy minimization.

distinguish it from that part of the topography resulting from the isostatic compensation of *static* loads (see above). Following [43], we estimate the dynamic component of topography w_{dyn} as

$$w_{\rm dyn} = \frac{\sigma_{zz}}{\rho g} \tag{12}$$

where σ_{zz} is the vertical stress component acting on the surface, ρ the density, and g the vertical component of the gravity acceleration. Convective shear stresses acting at the base of the lithosphere would also have an effect in the dynamic topography. However, they typically represent less than 5% of the dynamic topography generated by vertical stresses and therefore they can be neglected ([44]). In all our simulations the dynamic topography associated with SSC never exceeds \pm 150 m. This number would be reduced by as much as 75 % if we included the elastic strength of the plate ([44]).

Seismic velocities. The calculation of seismic velocities $[V_p^2 \rho = K_S + 4/3G \text{ and } V_s^2 \rho = G]$ requires knowing the elastic moduli of each stable phase, the density of the bulk rock at the pressures and temperatures of interest, and estimations of anelastic attenuation. Here we compute these properties by a free energy minimization procedure (see details in [45]) within the system CFMAS (CaO-FeO-MgO-Al₂O₃-SiO₂). These five major oxides make up more than 98% of the Earth's mantle, and therefore they constitute an excellent representation of mantle's composition. The minimization program (PerpleX) requires a thermo-dynamic database for pure end-members and solution models to compute the properties of stable phases (usually solid solutions of two or more end-members). The thermodynamic database used in the energy-minimization is that of [32] with solution models as listed in Table 5.

The thermal and pressure fields necessary to calculate the seismic velocities are obtained from the thermomechanical simulation. This generates an unavoidable inconsistency between the density values used to calculate buoyancy forces in the thermo-mechanical problem and those used to calculate seismic properties in the energy-minimization scheme (densities from the energy-minimization are systematically greater than those from the thermo-mechanical simulation). Parallel computations indicate that this inconsistency translates into errors of ≤ 1.1 % in our calculated *absolute* seismic velocities. However, the seismic structure (i.e. spatial velocity distribution) generated by our models is not significantly affected.

Anelastic effects are computed as a function of grain size (d), oscillation period (T_o) , T, P, and empirical parameters A, E, and α as [41, 46]

$$V_{\theta} = V_{\theta_o}(P, T) \left[1 - \zeta \cot\left(\frac{\pi\alpha}{2}\right) Q_s^{-1}(T_o, T, P, d) \right]$$
(13)

⁵⁷_{58 232} where $V_{\theta_o}(P,T)$ is the unrelaxed high frequency wave velocities at a given temperature and pressure (i.e. ²³³ including anharmonic effects) and θ stands for either P-wave or S-wave velocities. The term ζ takes the ⁶⁰



Figure 1. Boundary (a) and initial (b) conditions for SSC models. The gray areas in panel (a) are the regions were viscosity is decreased. The dashed lines delimit areas where shear heating is neglected.

values 2/9 and 1/2 for P-waves and S-waves, respectively ([41]). The quality factor is represented as

$$Q_s^{-1}(T_o, T, P, d) = A \left[T_o d^{-1} \exp\left(\frac{-E + VP}{RT}\right) \right]^{\alpha}$$
(14)

with $A = 750 \text{ s}^{-\alpha} \mu \text{m}^{\alpha}$, $\alpha = 0.26$, $E = 424 \text{ kJ mol}^{-1}$, $V = 1.2 \cdot 1.4 \times 10^{-5} \text{ m}^3 \text{ mol}^{-1}$, and R the universal gas constant [47,48].

237 2.5 Model setup and boundary conditions

The simulation domain is a rectangular box representing a vertical plane parallel to the plate motion (Fig. 1). The box is divided into approximately 13000 triangular elements representing a vertical thickness of 660 km in nature. The horizontal dimension is 7000 or 15000 km wide, depending on the particular model. Since the oceanic crust does not play any significant role on the dynamics of SSC, it is neglected in our model. Temperature boundary conditions assume constant temperatures at the surface and at the bottom of the simulation domain. The initial internal temperature distribution follows the half-space cooling model, which is calculated in terms of the surface temperature $T_{\rm surf}$, temperature at the bottom of the lithosphere T_{lith} , and thermal diffusivity κ as ([1])

$$T(t,z) = (T_{\rm surf} - T_{\rm lith}) \,\, \text{erfc} \left(\frac{z}{2\sqrt{\kappa t}}\right) + T_{\rm lith} \tag{15}$$

where t is the age of the plate, z is depth, and erfc is the complementary error function. The age t is directly related to the horizontal space dimension through the plate velocity. This model gives "conductive" temperatures above a specific isotherm T_{lith} , which represents the base of the lithosphere. For temperatures below this isotherm, a linear interpolation is done between T_{lith} (here chosen = 1603 K) and the temperature at the bottom of the box, T_{bot} . The latter is chosen to be 1880 K, in accordance with results from highpressure and high-temperature experiments on mineral phase equilibria (e.g. [29]). On the laterals the normal flow is set to zero.

The initial velocity field is set to zero in the entire domain. A constant horizontal velocity is imposed at the top of the model, everywhere but near the corners (see Figure 1a). The vertical velocity at the top of the box is zero. Imposing a constant velocity at the top along the entire domain length generates singularities in the upper corners (and nearby regions) where the strain rate reaches unrealistic high values and consequently high shear heating. We emphasize that these extremely high strain rate values are numerical artifacts produced by the boundary conditions, and consequently they do not represent any relevant physical process. To avoid these undesired effects we use a free-slip condition in regions near both corners, which leads to the generation of a smoother corner flow. For similar reasons, we further introduce two small weak zones near the upper corners (see Figure 1a), as commonly done in similar studies [12].

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The viscosity in these regions is divided by 10, 100 or 1000 depending on the model. In the other three sides of the domain free slip boundary conditions are imposed. To avoid excessive heating generation at the corners of the model, shear heating is turned off within two rectangular areas at the ends of the domain (see Figure 1a). Each rectangle represents 10% of the total domain length. Since we are interested in the generation and evolution of SSC regions well outside these regions, the neglect of shear heating within them does not affect our results and conclusions.

No initial condition needs to be specified for the pressure. This is a consequence of the fact that no time derivative of pressure appears in the governing equations (eqs. 1-3). When velocity is imposed everywhere on the boundaries, only pressure gradients appear in eq. (1), and total pressure can be determined by assuming an arbitrary constant (usually P=0) at some "quiescent" point at the surface.

3 Results

We organize this section in two parts. In the first part we described the main features of SSC and its effect on the thermal structure of both lithosphere and sublithospheric mantle. In the second part we analyze systematically the influence of key physical parameters on the generation and evolution of SSC.

276 3.1 General features of small-scale convection

In this section we present an illustrative model in which SSC is fully developed. The imposed upper velocity is 3.5 cm yr^{-1} , comparable to absolute velocities reported for oceanic plates [49]. The only internal heating term included in the energy equation is the adiabatic heating (i.e. shear heating and radiogenic heat production are set = 0). The model assumes a Newtonian rheology with the following parameters: activation energy $E = 120 \text{ kJ mol}^{-1}$, activation volume $V = 4 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1}$, and pre-exponential factor $A_D = 7.6 \times 10^{-16} \text{ Pa}^{-n} \text{ s}^{-1}$. Similar values have been extensively used in earlier studies on SSC (e.g. [11–14]), allowing qualitative comparisons between these and our models. We emphasize, however, that these activation energy and pre-exponential factor values are too low to be consistent with currently available laboratory experiments on diffusion creep (e.g. [23, 26]). A complete discussion on the effects of these parameters on the development and evolution of SSC is provided in the next section.

Figure 2a shows the resulting thermal structure after 83 My of simulation time (simulation time is the total number of time-steps in My and should not be confused with plate age, t, which is related to the horizontal dimension D and the imposed velocity v as t=D/v). At this time the model is already in a "dynamic steady-state". During this stage, the onset of SSC occurs at ~ 2100 km from the ridge (dotted line in Fig. 2), or what is the same, when the lithosphere is ~ 60 My old. We note that neither lateral boundary conditions nor the downstream developed at the rightmost part of the model (not shown in Fig. 2) influence these SSC instabilities. At shorter distances (younger lithosphere), the isotherms follows closely the initial thermal structure predicted by the half-space cooling model. The wavelength of the instabilities is of the order of 150-200 km throughout the entire model. Some isotherms (in Kelvin) are plotted to show the perturbing effect of SSC. The 1603 K isotherm, which is typically assumed to represent the base of the lithosphere, is completely distorted due to SSC (advection-dominated). Even the 1473 K isotherm shows an advective component, although considerably less than hotter isotherms.

The resulting viscosity structure is plotted in Fig. 2b. In order for SSC to develop, we find that the viscosity of the upper 300 km of mantle needs to remain lower than 10^{20} Pa s. Higher values suppress SSC, while lower values generates unrealistic high velocities. We will discuss further the effects of viscosity in the next section. Here we only note that the above value is similar to those previously reported by different authors ([12,13,50]). Figure 2c shows the vertical component of the velocity vector. Small-scale convection produces a significant vertical flow. Maximum velocities reach values of about 6 cm yr^{-1} , which is a factor of two greater than the imposed surface velocity. Strain rates values are within the range of 10^{-14} - 10^{-15} s⁻¹ between 100 and 500 km depth; the greatest values in the entire domain are always associated with SSC cells.

59 308 As expected, SSC slows down the conductive cooling of the lithosphere by replacing cold mantle with





Figure 2. Typical temperature (a), viscosity (b) and vertical velocity (c) when small scale convection develops. See text for details.

hotter mantle, effectively reducing its thermal thickness when compared with predictions from the half-space cooling model. Likewise, the underlying sublithospheric mantle is cooled by the cold downwellings. We illustrate this effect in Fig. 3. This figure compares temperature profiles across the mantle at different simulation times, starting from the initial temperature distribution given by the half-space cooling model (red line in Fig. 3). The profiles are located at 3000 km away from the ridge, where the plate is 85.7 My old. To remove short-wavelength temperature anomalies, the temperature is horizontally averaged in a 400 km region (from x = 2800 to 3200 km). The profiles in Fig. 3a clearly show that the lithosphere reaches a "steady" thermal thickness after ~ 60 My of simulation time. However, a closer inspection (Fig. 3c) reveals that this steady thickness is reached after only ~ 30 My of simulation time. During this time, about 50 km of unstable lithospheric material is removed, thinning the thermal thickness plate by an equal amount. This indicates that the temperature structure predicted by the half-space cooling model for plates older than 65 My is extremely unstable in a convecting mantle characterized by the present physical parameters.

At the base of the lithosphere, temperature increases by about 200 K with respect to predictions from the half-space cooling model (see Fig. 3b). At depths between 200 and 400 km, the temperature variation is $< \pm 25$ K with respect to the initial "adiabatic" profile, but this difference increases to > 75 K (i.e. colder) in the transition zone. However, the latter value needs to be taken with caution, since our energy equation does not include the effect of latent heat of phase transformations (olivine-wadsleyite and wadsleyite-ringwoodite). It has been shown that the temperature increase across the transition zone along an adiabat can be as much as 100 K ([29]). Therefore, if we take into account this effect, the temperature difference between our initial "adiabatic" and the final profiles is reduced to $\leq 10-20$ K. This supports our choice for the temperature at the bottom of the simulation box.

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Figure 3. Temporal evolution of temperature for model 16 at 3000 km away from ridge. Panel (a) and (c): temperature profiles at different times. Panel (b) temperature variation between 1 and 180 My.



Figure 4. Lithospheric base defined by the averaged 1603 K isotherm.

Table 6.	Models.	
Model	$_{\rm SH}$	RHP (W m^{-3})
09	no	0
14	yes	0
15	no	2.0×10^{-8}
16	yes	2.0×10^{-8}

3.2 Influence of key physical parameters

Effect of shear heating and radiogenic heat production. In the following set of numerical experiments, the effects of shear heating and radioactive heat production (RHP) are tested for a model with identical rheological parameters as in the previous section. Table 6 list the models and whether they include or not shear heating and RHP. The adopted RHP rate per unit mass is at the high end of estimated values for the mantle ([21,51]). Model 09 was described in the previous section and is shown in Figure 2. We note that SSC is active in all four models.

57 330 billet SSC is derive in an rour inducits.
58 337 In order to make a meaningful comparison between our results and those from conductive plate models,
59 338 we calculate the averaged depth of the 1603 K isotherm by applying a moving-average filter to the depths
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Figure 5. (a) Temperature in K, (b) sea floor topography, and (c) surface heat flow. Temperature in panel (a) corresponds to model 69 after 120 My of simulation time. Horizontal distance [km] is measured from the ridge, and the equivalent plate age [My] is included at the top.

of the isotherm. Since the wavelength of SSC is 150-200 km, a window size of 250 km removes the intrinsic local variability caused by SSC without adding significant diffusion.

The averaged 1603 K isotherm, defining the bottom of the lithosphere, for the four models is shown in Figure 4. We find that values of shear heating associated with SSC are of the same order as the RHP $(10^{-8} \text{ W m}^{-3})$, although some punctual values can become one order of magnitude greater. Nevertheless, as shown in Fig. 4 the shear heating has little influence on the final lithospheric thickness. This is because the mean shear heating remains much lower ($\sim 10 - 15\%$) than the RHP. The strain rate, and consequently the shear heating, may be underestimated in the sublithospheric mantle where the SSC cells develop. Thermally controlled shear zones may have a spatial length scale of about 1 km [52] while the size of the elements in our model is several times greater. Nonetheless, reducing the spatial discretization will not drastically change the viscous dissipated heat. Models 09 and 14 exhibit an average lithospheric thickness of ~ 200 km in plates 100 My old. This value is well outside the ranges of all existing plate models (cf. [1]), and predicts patterns of sea floor topographies and SHF that do not fit observed data satisfactory (Fig. 5). On the other hand, the inclusion of RHP brings the average thickness of old oceanic lithosphere to values around 125 km, closer to results from theoretical plate models ([3, 53]) and combined geophysical-petrological models ([28]). This was also found by [12].

Influence of adiabatic heating. In the absence of other sources of heat, a fluid element would undergo changes in temperature due to variations in pressure. In a convecting mantle, major changes in pressure are related to vertical motion of material. Without adiabatic heating, the density difference between downwellings and upwellings increases, enhancing SSC. We illustrate this by setting to zero the adiabatic heating term in one of our previous models (model 95 in Fig. 4). In this case, the vigour and eroding capacity of the convective cells is dramatically increased. Average lithospheric thicknesses are now 25 to 50 km thinner than in the case with adiabatic heating. As a result, sea floor topography and SHF data are more closely reproduced, although a continuous decrease with time is still seen in both observables (i.e. no flattening, Fig. 5).

Adiabatic heating has been neglected in most studies addressing SSC (e.g. [11, 13, 14, 16, 54]). [12] state that adiabatic and shear heating was explicitly included in their calculations, but no assessments of their influence on SSC or on the thermal structure of the plate were provided. Our results indicate that adiabatic heating exerts a major control on the final structure of the oceanic lithosphere (Fig. 4). Consequently, any attempt to explain its thermal structure and associated geophysical observables using thermo-mechanical

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Figure 6. Models varying the activation energy E and the material parameter A_D . Models marked with cross do not initiate SSC, models marked with tildes do initiate SSC. A linear relation between E and A_D defines the region where SSC is active. The gray region indicates ranges in agreement with laboratory experiments. Numbers at the side of the markers are the run number.

models should consider this effect. In principle, this conclusion also applies to other gravitational instabil-ities such as lithospheric delamination and unroofing, although further studies are needed. In this context, we emphasize that it is widely accepted that the temperature increase within most of the Earth's mantle follows closely the condition of uniform entropy (i.e. adiabatic profile, cf. [1]).

Influence of plate motion. A positive relation between the onset time of SSC and plate velocity has been reported in several studies ([11,14,55]). Using isoviscous convection models, [55] found a clear delay in the SSC onset when increasing plate velocity. This author suggested that sufficiently high plate motions may even prohibit the development of gravitational instabilities. [11] described a similar behaviour in their 2D Newtonian simulations. On the other hand, [16] found no significant correlation between the onset time of the first dripping instability and plate velocity.

We have run five Newtonian models to test the role of plate velocity on the development of SSC. The parameters used in these models are those of model 16 in Table 6. We found a positive correlation between plate velocity and SSC onset time. For velocities of 2, 4, 6, 8, and 10 cm yr⁻¹, the first drippings appear when the plate is 28, 43, 70, 72, and 80 My old, respectively, in agreement with the observations reported in [11] and [55].

Influence of rheological parameters. We run fourteen Newtonian models varying the pre-exponential factor A_D and the activation energy E to test the sensitivity of SSC to these parameters. Figure 6 shows which models developed vigorous SSC (in green) and which ones failed to do so (red). The results suggest a quasi-linear relation between E and A_D that defines the region where SSC can be active. Moreover, the range of values that define this boundary seems to be rather limited (see e.g. experiments 69-64 and 9-48), indicating that the development of SSC is extremely sensitive to these parameters, and hence to viscosity. In fact, the straight line in Fig. 6 separates the models with average upper mantle viscosities \lesssim (green) and > (red) 10²⁰ Pa s. The latter value is found to be a limit above which vigorous SSC cannot develop. When the average viscosity of the upper mantle is above this threshold, the cooling of the oceanic lithosphere follows closely the HSC model. Although some small instabilities may develop at the base of the lithosphere, they cannot evolve into well defined drippings due to the high viscosity of the mantle below. This in turn precludes any significant thermal erosion of the lithosphere. A similar result was also found by [14]. On the other hand, if the average viscosity of the upper mantle is below $\sim 2 \times 10^{19}$ Pa s, velocities associated with SSC reach unrealistic values (> 10 m yr⁻¹), the lithosphere is strongly eroded, and observables depart significantly from the expected values.

The shaded box in Fig. 6 represents the range of values for E and A_D from laboratory experiments [21, 23, 26]. Three of our models (labeled as 63, 69, and 73) that developed SSC are well within this range, although their vigour is moderate. Note also that the parameters used in our previous Newtonian examples (taken from previous studies) are far from experimental results. In particular, a low value for

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E is commonly applied to mimic the contribution of dislocation creep [11–14], which otherwise would underestimate the erosion of the lithosphere by SSC. We anticipate, however, that the water content of residual peridotites in oceanic lithosphere, as well as their intrinsic compositional buoyancy, may be more important in controlling the thermal erosion of the lithosphere. A detailed assessment of these two effects is under way and will be published elsewhere.

Fig. 5 shows the observables predicted by model 69. In this model, both elevation and SHF increase monotonically with age, in contrast to what is observed. However, its seismic structure (Fig. 7) resembles closely those recently obtained from tomography studies in the Pacific as well as in other oceans (compare with Fig. 10 in [56] and Fig. 11 in [57]). Other similar models (not shown here) give identical results. This discrepancy between what is inferred from seismic data and from other observables (i.e. SHF, bathymetry) is a fundamental, yet not solved, problem in geodynamics.

4 Discussion

It is worth noting that [13] recently concluded that dislocation creep (i.e. non-Newtonian rheology) is the main deformation mechanism in the upper mantle, based on a comparison between their numerical simulations and the tomography of Ritzwoller [57] in the Pacific. The lithospheric thermal structure derived by [57] shows a distinctive flattening of the isotherms (strictly, isovelocity contour lines) between 70 and 100 My, which these authors associate with a period of lithospheric reheating. [13] showed that this reheating could be modelled only with non-Newtonian models, if the adopted rheological parameters are taken to be consistent with laboratory experiments. However, the unrealistic thermal structure used in the numerical simulations (see their Fig. 3), as well as the neglect of activation volumes and internal heating, make their conclusions ambiguous. Moreover, [56] has recently presented a tomography for the Pacific in which no flattening of the isovelocity contour is observed. These authors pointed out that the flattening observed by [57] may be an artifact due to insufficient path coverage. If this is true, and in light of our results with Newtonian models, the argument used by [13] to favour dislocation over diffusion creep in the upper mantle becomes invalid.

There is, however, abundant independent evidence that point to dislocation creep as the main defor-mation mechanism in the upper mantle [21, 23, 26]. We have run experiments with combined dislocation-diffusion creep rheologies, always restricting rheological parameters to realistic ranges (i.e. within the shaded box in Fig. 6 for diffusion creep, and n = 3.5, E = 500, V = 17 and $A_D = 2.42 \times 10^{-14}$ for dislocation creep), to test the potential eroding effect of dislocation creep. In our simulations, dislocation creep becomes dominant in the first 250-300 km (i.e. $\eta_{disl} < \eta_{diff}$) and generates an extra 20-50 km of erosion at the bottom of the lithosphere (Fig. 4). Due to the greater activation volume of dislocation creep, and the small convective stresses generated below ~ 250 km depth (≤ 0.1 MPa), diffusion creep becomes dominant below this depth. Interestingly, the resulting sea floor topography and SHF exhibit wide regions

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where the signals become flat (Fig. 5), but the overall fitting to observed data is still poor.

In a recent study, [58] argued that the "normal" evolution of sea floor topography may be explained by the half-space cooling models. These authors conclude that most of the observed bathymetric signal is influenced by anomalous crust and hot-spots, and therefore they do not necessarily reflect the normal evolution of the oceanic lithosphere. This conclusion is in agreement with the seismic tomography of [56] and the results reported in this study. If true, it could offer (at least partially) a solution to the mentioned dichotomy between surface observables and seismic observations.

As a final caveat, we note that our 2D results may represent lower bounds in terms of SSC vigour and 444 445 eroding capacity. Results from 3D numerical [11] and laboratory experiments on SSC [59, 60] indicate that longitudinal rolls (LR; aligned with plate motion) tend to be favoured over transverse rolls (TR; 446 perpendicular to plate motion). More importantly, there is evidence that LR are more vigorous than TR, 447 and that plate motion enhances SSC in 3D models, contrary to what is predicted by 2D simulations [11]. If 448 these observations are correct, 3D SSC could lead to thinner plates than those obtained in 2D simulations, 449 at least for fast-moving plates. Whether or not this would result in a significant modification of the thermal 450 structure of plates or in a better fitting of the observables remains to be tested. 451

Further studies on i) the viscosity structure of the oceanic upper mantle, ii) influence of shallow phase changes and melt dehydration on SSC, and iii) and high-resolution seismic experiments will provide crucial insights into the evolution of the oceanic lithosphere.

455 **5** Conclusions

We have studied the development and evolution of small-scale gravitational instabilities under the Earth's oceanic lithosphere using numerical simulations. Our results can be summarized as follow:

- (i) The influence of three different heating terms was tested. Shear heating has negligible influence on the
 overall temperature of the model as well as on SSC. In contrast, heat from radiogenic sources increases
 mantle temperatures, favour the development of SSC, and reduce the average thermal thickness of
 the lithosphere. The inclusion of adiabatic heating reduces the temperature contrasts between ambient
 mantle and downwellings/upwellings. This in turn diminish the vigour of SSC and lithospheric erosion.
- mantle and downweilings/upweilings. This in turn diminish the vigour of SSC and lithospheric erosion. **35** 463 (ii) A low viscosity region below the lithosphere is needed to develop and maintain SSC. The height of this **36** 464 zone has to be similar to the horizontal wavelength of instabilities (≥ 200 km). For realistic rheological **37** 465 parameters, the average viscosity in this region cannot be higher than $\sim 10^{20}$ Pa s.
- 38 Small-scale convection can be generated using experimentally derived rheological parameters cells. 466 (iii) 39 Nevertheless, the activation of SSC does not always result in significant lithospheric erosion. To reduce 467 40 the lithospheric thickness considerably, Newtonian models need activation energy and pre-exponential 468 41 factor values that are too low in comparison with those reported in laboratory studies. In models 469 42 where both deformation mechanisms (i.e. diffusion and dislocation) are present, the low viscosity zone 470 43 is dominated by dislocation and considerable lithospheric erosion occurs. 471 44
- 45 472 (iv) A positive relation is observed between plate velocity and SSC onset time. This corroborates results
 46 473 from previous studies.
- 47 (v) Although our synthetic seismic structures resemble closely results from tomography studies in oceanic mantle, sea floor topography and SHF hardly fits the observed trends in old lithospheres. Further numerical and seismological studies are needed to explain this discrepancy.

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References

- G. Schubert, D.L. Turcotte, and P. Olson Mantle Convection in Earth and Planets, Cambridge University Press, UK, 2001.
- D.P. McKenzie, Some remarks on heat flow and gravity anomalies, J. Geophys. Res. 72 (1967), pp. 6261–6273.
- B. Parsons and J.G. Sclater, An analysis of the variation of ocean floor bathymetry and heat flow with age, J. Geophys. Res. 82 [3] (1977), pp. 803-827.
 - D.L. Turcotte and E.R. Oxburgh, Finite amplitude convective cells and continental drift, J.Fluid Mech, 28 (1967), pp. 29-42.
- W. Schroeder, The empirical age-depth relation and depth anomalies in the Pacific ocean basin, J. Geophys. Res. 89 (1984), pp. |5|9873 - 9883.
- C.A. Stein and S. Stein, A model for the global variation in oceanic depth and heat flow with lithospheric age, Nature 129 (1992), [6] pp. 123–129.
- B. Parsons and D. McKenzie, Mantle convection and thermal structure of the plates, J. Geophys. Res. 83 (1978), pp. 4485–4496.
- D.A. Yuen, W.R. Peltier, and G. Schubert, On the existence of a second scale of convection in the upper mantle, Geophys. J. Roy. Astron. Soc. 65 (1981), pp. 171–190.
- C. Jaupart and B. Parsons, Convective instabilities in a variable viscosity fluid cooled from above, Phys. Earth Planet. Inter. 39 [9] (1985), pp. 14-32.
- D.A. Yuen and L. Fleitout, Thinning of the lithosphere by small-scale convective destabilization, Nature 313 (1985), pp. 125-128. [10]J. Hunenvan, J. Huang, and S. Zhong, The effect of shearing on the onset and vigor of small-scale convection in a Newtonian [11]
- rheology, Geophys. Res. Lett. 30 (2003), p. doi:10.1029/2003GL018101. J. Huang and S. Zhong, Sublithospheric small-scale convection and its implications for the residual topography at old ocean basins [12]
 - and the plate model, J. Geophys. Res. 110 (2005) doi:10.1029/2004JB003153. [13] J. Hunenvan et al., New evidence for dislocation creep from 3-D geodynamic modeling of the Pacific upper mantle structure, Earth Planet. Sci. Lett. 238 (2005), pp. 146–155.
 - [14] J. Huang, S. Zhong, and J. Hunenvan, Controls on sublithospheric small-scale convection, J. Geophys. Res. 108 (2003) doi:10.1029/2003JB002456.
 - [15] J. Korenaga and T. Jordan, Physics of multiscale convection in Earth's mantle: evolution of sublithospheric convection, J. Geophys. Res. 109 (2004), p. doi:10.1029/2003JB002464.
 - [16] C. Dumoulin, M. Doin, and L. Fleitout, Numerical simulations of the cooling of an oceanic lithosphere above a convective mantle, Phys. Earth Planet. Int. 125 (2001), pp. 45–64.
 - N. Curlet, Experimental and numerical modeling of three dimansional natural convection in an enclosure, Phys. Earth Planet. Int. |17|125 (2001), pp. 45-64.
- [18] F.H. Busse, Fundamentals of thermal convection, in Mantle convection. Plate tectonics and global dynamics, W.R. Peltier ed., The fluid mechanics of astrophysics and geophysics Gordon and breach science publishers, Montreux, Switzerland, 1989, volume 2, pp. 23 - 95
 - [19] J. Donea and A. Huerta Finite Element Methods for Flow Problems, Wiley, Chichester, West Sussex PO19 8SQ, England, 2002.
 - [20] S. Zlotnik et al., Numerical modelling of tectonic plates subduction using X-FEM, Comput. Methods Appl. Mech. Engrg. 196 (2007), pp. 4283-4293.
 - G. Ranalli Rheology of the Earth, second Chapman and Hall, 2–6 Boundary Row, London, 1995.
 - [22]S.H. Kirby, Rheology of the lithosphere, Rev. Geophys 21 (1983), pp. 1458–1487.
 - G. Hirth and D. Kohlstedt, Rheology of the Upper Mantle and the Mantle Wedge: A View ¿From the Experimentalists, in Inside [23]the subduction factory, Geophys. Monograph 138, AGU, J. Eiler ed., 2003, pp. 83-105.
 - [24] M. Nettles and A. Dziewonski, Radially anisotropic shear-velocity structure of the upper mantle globally and beneath North America, J. Geophys. Res. (2008), p. in press.
 - [25] T.A.C.H.C.P.F.D. Mainprice D., TPressure sensitivity of olivine slip systems and seismic anisotropy of Earth's upper mantle, Nature 433 (2005), pp. 731–733.
 - [26] S.I. Karato and P. Wu, Rheology of the upper mantle: a synthesis, Rev. Science 260 (1993), pp. 771–778.
 - K. Regenauer-Lieb et al., From point defects to plate tectonic faults, Philosophical Magazine 86 (2006), pp. 3373 – 3392.
 - [28] J.C. Afonso, G. Ranalli, and M. Fernàndez, Density structure and buoyancy of the oceanic lithosphere revisited, Geophys. Res. Lett. 34 (2007), pp. L10302, doi:10.1029/2007GL029515.
 - $[29] T. Katsura et al., Olivine-wadsleyite transition in the system (Mg, Fe)_2 SiO_4, J. Geophys. Res. 109 (2004) doi:10.1029/2003JB002438.$ A. Guest, G. Schubert, and C.W. Gable, Stresses along the metastable wedge of olivine in a subducting slab: possible explanation [30]for the Tonga double seismic layer, Phys. Earth Planet. Inter. 149 (2004), pp. 187–200.
 - [31] T. Katsura and E. Ito, The system Mg_2SiO_4 -Fe₂SiO₄ at high pressures and temperatures: Precise determination of stabilities of olivine, modified spinel, and spinel, J. Geophys. Res. 94 (1989), pp. 15663-15670.
 - [32] L. Stixrude and C. Lithgow-Bertelloni, Influence of phase transformations on lateral heterogeneity and dynamics in Earth's mantle, Earth Planet. Sci. Lett. 263 (2007), pp. 45-55.
 - [33] L. Stixrude and C. Lithgow-Bertelloni, Mineralogy and elasticity of the oceanic upper mantle: Origin of the low-velocity zone, J. Geophys. Res. 110 (2005), pp. B03204, doi:10.1029/2004JB002965.
 - [34] D.L. Schutt and C.È. Lesher, Effects of melt depletion on the density and seismic velocity of garnet and spinel lherzolite, J. Geophys. Res. 111 (2006) doi:10.1029/2003JB002950.
 - [35] M.E. Akaogi, E. Ito, and A. Navrotsky, Olivine-modified spinel transitions in system Mg_2SiO_4 - Fe_2SiO_4 : calorimetric measurements, thermomechanical calculations, and geophysical implications, J. Geophys. Res. 94 (1989), pp. 15671–15685.
 - [36] Y. Fei, Thermal expansion, in Rock Physics and Phase Relations, T. Ahren ed., third ed., , AGU Reference Shelf AGU, Washington DC, 1995, part 3, pp. 29–44. A.R. Oganov, J.P. Brodholt, and G.D. Price, Ab initio elasticity and thermal equation of state of MgSiO₃ perovskite, Earth Planet.
 - Sci. Lett. 184 (2001), pp. 555-560.
 - [38] A. Chopelas, Thermal expansivity of mantle relevant magnesium silicates derived from vibrational spectroscopy at high pressure, Am. Miner. 85 (2000), pp. 270–278.
 - [39] J.C. Afonso, G. Ranalli, and M. Fernández, Thermal expansivity and elastic properties of the lithospheric mantle: results from mineral physics of composites, Phys. Earth Planet. Inter. 149 (2005), pp. 279-306.
 - [40] A.M. Hofmeister, Mantle values of thermal conductivity and the geotherm from phonon lifetimes, Science 283 (1999), pp. 1969–1706. [41]J.C. Afonso et al., Integrated geophysical-petrological modelling of the lithospheric-sublithospheric upper mantle: methodology and applications, Geochem. Geophys. Geosyst. - (2008), p. in press.
- [42] G.T. Jarvis and W.R. Peltier, Mantle convection as a boundary layer phenomenon, Geophys. J. R. Astron. Soc. 68 (1982), pp. 389 - 427
- [43] D.P. McKenzie, Surface deformation, gravity anomalies and convection, Geophys. J. Roy. Astron. Soc. 48 (1977), pp. 211–238.

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- [44] G. Marquart and H. Schmeling, Topography and geoid undulations caused by small-scale convection beneath continental lithosphere of variable elastic thickness, Geophys. J. 97 (1989), pp. 511–527.
 - [45] J.A.D. Connolly, Computation of phase equilibria by linear programming: A tool for geodynamic modeling and its application to subduction zone decarbonation, Earth Planet. Sci. Lett. 236 (2005), pp. 524-541.
 - S.I. Karato, Importance of anelasticity in the interpretation of seismic tomography, Geophys. Res. Lett. 20 (1993), pp. 1623–1626.
 - I. Jackson et al., Grain-size-sensitive seismic wave attenuation in polycrystalline olivine, J Geophys. Res. 107 (2002), pp. [47] $NO.B12,\!1360,\!doi:\!10.1029/2001JB001225.$
 - [48] U.H. Faul and I. Jackson, The seismological signature of temperature and grain size variations in the upper mantle, Earth Planet. Sci. Lett. 234 (2005), pp. 119–134.
 - A.E. Gripp and R.G. Gordon, Young tracks of hotspots and current plate velocities, Geophys. J. Int. 150 (2002), pp. 321-361. [49]
 - [50] J. Korenaga, On 'steady-state' heat flow and the rheology of oceanic mantle, Geophys. Res. Lett. 29 (2002) doi:10.1029/2002GL016085.
 - [51] S. Labrosse and C. Jaupart, Thermal evolution of the Earth: Secular changes and fluctuations of plate characteristics, Earth Planet. Sci. Lett. 260 (2007), pp. 465–481.
 - [52] K. Regenauer-Lieb and D.A. Yuen, Modeling shear zones in geological and planetary sciences: solid- and fluid-thermal-mechanical approaches, Earth Sciences Reviews 63 (2003), pp. 295-349.
 - [53] D. McKenzie, J. Jakson, and K. Priestely, Thermal structure of oceanic and continental lithosphere, Earth Planet. Sci. Lett. 233 (2005), pp. 337–349.
 - [54] S. Zaranek and E. Parmentier, Convective cooling of an initially stably stratified fluid with temperature-dependent viscosity: impli-cations for the role of solid-state convection in planetary evolution, J. Geophys. Res. 109 (2004), p. doi:10.1029/2003JB002462.
 - G. Houseman, Large aspect ratio convection cells in the upper mantle, Geophys. J. R. Astron. Soc. 75 (1983), pp. 309-334.
 - A. Maggi et al., Multimode surface waveform tomography of the Pacific Ocean: a closer look at the lithospheric cooling signature, Geophys. J. Int. 166 (2006), pp. 1384–1397.
 - [57] M. Ritzwoller, N. Shapiro, and S.J. Zhong, Cooling history of the Pacific lithosphere, Earth Planet. Sci. Lett. 226 (2004), pp. 69-84. [58]T. Korenaga and J. Korenaga, Subsidence of normal oceanic lithosphere, apparent thermal expansivity, and seafloor flattening, Earth Planet. Sci. Lett. (2008) doi: 10.1016/j.epsl.2007.12.022.
 - [59] F.M. Richter, Convection and the large-scale circulation of the mantle, J. Geophys. Res. 78 (1973), pp. 8735–8745.
- [60] F.M. Richter and B. Parsons, On the interaction of two scales of convection in the mantle, J. Geophys. Res. 80 (1975), pp. 2529–2541.









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