Comparing global mantle models: from mapping to hypotheses testing

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We present a comprehensive quantitative analysis of recent p- and s-wave tomography and compare seismological with geodynamical models of mantle structure. By focusing on similarities and differences between these models as a function of spatial wavelength and depth we can distinguish stable features from structures that are dependent on the data selection and the inversion technique. Our approach should help in moving from the mapping phase of global tomography to the testing of geologically relevant hypotheses since we must understand uncertainties in the input models to proceed with geodynamical interpretations.

We confirm previous results such as the presence of slab-like features in the mid-mantle in all tomography models. Radial correlation functions, furthermore, do not show strong layering at any depth, favoring whole mantle style convection. However, geodynamical models that are based on whole mantle flow, subduction and plate motion histories do not correlate well with tomography on a global level. This testifies to our incomplete understanding of what the mantle is "supposed" to look like.

Focusing on a particular geodynamics application, we proceed to explore consequences of model discrepancies when tomography is used as a proxy for density distribution in the mantle. Geodynamical models that are driven by such buoyancy anomalies are seriously affected by the choice of model. We demonstrate this finding by looking at inferred plate driving forces where we observe that a number of model combinations can explain plate motions, leading to different conclusions about the relative importance of plate tectonic forces such as ridge-push and slab-pull. Since plate velocities appear to be poorly suited to help selecting viscosity profiles or density models, we propose the use of stress and strain fields in the plates instead. Using an improved description of the faulted lithosphere, we explore how well observations such as the global stress or strain maps can be matched by geodynamical models.

A re-assessment of the Present day Heat flux due to Mantle Secular Cooling

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The reduced mantle heat flux, the global surface heat flux of the Earth minus the crustal and lithospheric radioactive heat production, reflects contributions from internal heating by 1) radioactive decay within the mantle, 2) secular cooling within the mantle, and 3) basal heating from the core. An updated estimate of the present day heat flux due to secular cooling of the whole mantle ($Q_{M\text{secular}}$) is calculated here using the thermal properties (T_a, α) of a pyrolite mineral physics model and an independent estimate of the present day cooling rate of the mantle. The mantle temperature structure is used to calculate a refined estimate of the total heat capacity (Φ_M) via Stacey and Loper (1984):

$$\Phi_M = \int_{r_{cmb}}^{r_{surface}} (\rho C_P T_a / T_{cmb} + \alpha P)\, 4\pi r^2 dr$$

where r_{cmb} is the outer core radius, and an updated estimate of the average cooling rate of the mantle (assuming whole mantle convection) calculated via (Turcotte & Schubert, 1982):

$$\left(\frac{\partial T_M}{\partial t}\right)_{\text{NOW}} = -\frac{\frac{3}{4}\lambda\left(\bar{T}_{\text{NOW}} - T_{surface}\right)}{1 + \frac{E_a}{4R\bar{T}^2_{\text{NOW}}}\left(\bar{T}_{\text{NOW}} - T_{surface}\right)}$$

where $\bar{T}_{NOW}$ is the present day volume averaged mantle temperature, λ is the average decay constant for the radiogenic isotopes in the mantle, and $(E_a/R\bar{T})$ is a parameter that may be evaluated from empirical mineral flow laws where R is the gas constant and E_a is the activation energy for viscous creep. From a pyrolite mineral physics adiabatic geotherm that is anchored to the average MORB potential temperature, I calculate a mean mantle temperature of 2117 K and adopt l = 2.77 10^{-1} yr^{-1} and $E_a/R\bar{T}$= 30. These values result in Φ_M = 6.52 10^{27} J/K and $\left(\frac{\partial \bar{T}_M}{\partial t}\right)_{\text{NOW}}$ = -48 K/Gyr and are higher and lower, respectively, than typical values adopted and estimated in parameterized cooling models of the Earth's thermal history. When these updated values for these parameters are combined, they result in a mantle secular cooling heat flux of = 10.0 10^{12} W.

Together with the global surface heat flux (42.5 10^{12} W) and independent estimates of the radioactive heat production in the crust (~9.13 10^{12} W), lithosphere

(~0.38 10^{12} W) and depleted mantle (2.00 10^{12} W), the updated value determined here for constrains the calculated net core heat flux via:

$$Q_{\text{global}} = \left(Q_{\text{crust}} + Q_{\text{lithosphere}} + Q_{\text{mantle}}\right)_{\text{radioactive}} + Q_{M\text{secular}} + Q_{\text{net core}} \cdot$$

This gives =20.96 10^{12} W.

Mantle convection: what about the experimental approach ?

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Mantle convection depends strongly on the rheological properties of deep Earth's minerals. Rheology is probably one of the less well known physical property under extreme P, T conditions. The reason is that we don't have yet any machine capable of achieving significant and controlled deformation at high-P, high-T. However, significant advances have recently been made to investigate plastic properties of high-pressure minerals. Most of them have been based on developments of the multianvil technology. On the one hand, the high-pressure assemblies have been modified in such a way that some plasticity is enhanced during a high-P, high-T experiment. Stresses or strains can't be measured and the information withdrawn are essentially microstructural. On the other hand, coupling of the multianvil technology with synchrotron-based diffraction experiments allows real-time measurement of elastic strains. Recent achievements of rheological studies at extreme pressures are presented. Success, pitfalls and perspectives are discussed.

Anomalies of temperature and iron in the uppermost mantle

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Anomalies of temperature and composition in the Earth's mantle induce anomalies of density and seismic velocities. Seismic tomography models do not give access to temperature anomalies and compositional changes simultaneously. To infer these anomalies, one needs an additional data set, such as gravity data.

First, we invert gravity anomalies (EGM96) and a global S-wave velocity model (S16RLBM) for the ratio $\zeta = dln\rho/dlnV_s$, which relates relative density anomalies to relative S-wave velocity anomalies. Data are filtered for the spherical harmonic degrees 11 to 16, and calculations are made separately for oceanic and continental regions. The resulting model of ζ is significantly different for the sub-continental and sub-oceanic mantle. Below continents (oceans), ζ has positive values down to z=220 km (z=140 km). The absolute values of ζ are small, less than 0.05. If one accounts for anelasticity, these values are consistent with mineral physics experiments. A variety of tests suggest that this model is robust.

We then invert relative V_s-anomalies (δV_s) and ζ for anomalies of temperature (δT) and global iron ratio (δFe). Positive V_s-anomalies are associated with negative temperature variations and iron depletion. For instance, the temperature and iron anomalies associated with δV_s=3% and ζ=0.03 are about δT=-150K and δFe=-1.25%, respectively. We then compute values of δT and δFe in the uppermost mantle down to 300 km. Below oceans and tectonic continents, the mantle is nearly homogeneous i.e., the mean values of δT and δFe are close to zero. On the other hand, down to z=250 km old cratons are significantly colder than average mantle and depleted in iron: the mean values of δT and δFe reach -300K and -2.7%, respectively. This result is important since depletion in iron induces positive buoyancy that may balance the negative buoyancy induced by low temperatures.

Lithospheric deformation and continental crust recycling: The roles of rheology and eclogitization

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A mechanical model with viscous (ductile) and plastic (brittle) rheologies is used to investigate the effect of eclogitization on the dynamics of convergence. Density increases by 300 to 600 km/m^3 during eclogitization of crustal rocks and continental lower crust and oceanic crust reach higher density than mantle. We explore cases of intracontinental deformation, subduction, and continental collision. We consider a wide range of parameters for friction, activation energy, and initial thermal state, and cases with or without eclogitization. The style of deformation appears to be primarily controlled by the presence or absence of weak zones. Simulations are run with a constant convergence velocity (1.5 or 4 cm/a) and the evolution of the compressive force through time is thus a critical test of the model viability. For intra-continental deformation, when the brittle crust is decoupled from the mantle, the mantle deforms by symmetrical bending or subduction, and a variable amount of lowermost crust is entrained with the sinking mantle. In these cases the compressive force remains in the (10^{12}-3×10^{13} N/m) acceptable range. Oceanic subduction only occurs if a low friction shear zone is specified in the brittle realm. During the transition from oceanic subduction to collision the oceanic crust in this shear zone remains trapped between colliding crusts. Eclogitization has no influence on the initial mode of deformation. In all simulations the influence of eclogitization increases progressively with time, in proportion of the amount of lower continental crust and oceanic crust entrained into the mantle. The structural evolution of the orogen model depends on eclogitization if initial decoupling occurs at mid crustal level. The simulations thus indicate that eclogitization promotes convergence and also, in some cases, enables the recycling of the whole continental lower crust into the mantle. We also suggest that eclogitization could regulate the long term convergence rate in orogens.

Numerical simulations of the cooling of an oceanic lithosphere above a convective mantle

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Numerical simulations of two-dimensional Rayleigh-Bénard convection are designed to study lithospheric cooling above a convective mantle. A strongly temperature- and pressure-dependent viscosity fluid is heated from below or from within. In a first set of simulations, an imposed velocity at the surface of the box mimicks the plate motion between a ridge and a convergent plate boundary. Dripping instabilities at the base of the lithosphere are not observed close to the ridge. Nevertheless, the material flows along the slope defined by the lower part of the lithosphere and feeds the first descending drip. Afterwards, cold downgoing instabilities develop continuously and randomly at the base of the lithosphere. Surface heat flow, subsidence and lithospheric temperature structure obtained by the convective simulations are compared to the predictions of three 2-D conductive models: the Plate, Chablis, and Modified Chablis models. These models differ by their applied bottom boundary condition which represents the lithosphere/asthenosphere convective coupling, i.e. by the presence or absence of instabilities developing at the base of the lithosphere. The conductive model which best explains the lithospheric cooling obtained by convective simulations is the Modified Chablis model. In this model a variable heat flow (depending upon the viscosity at the base of the lithosphere) is applied along an isotherm located in the lower unstable part of the lithosphere. In the second set of simulations, we model transient lithospheric cooling by imposing a zero temperature at the surface of the mantle with an initially homogeneous temperature. For a while the lithosphere cools approximatively as a conductive half-space and lithospheric isotherms remain flat. As instabilities progressively develop at the base of the lithosphere, lithospheric cooling departs from the half-space model. The Plate model fits better the transient lithospheric cooling in these simulations. We quantify the characteristic timescale of the exponential growth of instabilities as a function of the Rayleigh number and of the viscous temperature scale. This study emphasizes the role of the lithospheric isotherms topography on the development of instabilities.

Mixing of Heterogeneities in Mantle Plumes

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The long standing idea that the source of oceanic island basalts includes ancient subducted oceanic crust is strengthened by recent geochemical data on Hawaiian shield lavas. The isotopic variations in the Koolau and Mauna Kea lavas document the presence of two distinct recycled components: 1) ancient oceanic crust and pelagic sediments, 2) altered ultramafic lower crust or lithospheric mantle. Lassiter and Hauri [1998] suggest that both components are from the same packet of recycled oceanic lithosphere, thus implying that chemical heterogeneities a few kilometers thick can be preserved in the convecting mantle.

Here I investigate the role of mantle plumes in mixing mantle heterogeneities, in particular I address the following questions: How are mantle heterogeneities entrained by mantle plumes? Which regions of the mantle are more efficiently sampled by plumes generated at the core-mantle boundary? Is mixing more efficient in the plume head or in the tail? Are the implications from geochemistry consistent with fluid dynamical models?

I use the three dimensional convection code (Stag3d) by Paul Tackley to model the formation and rise of plumes, while the flow trajectory of the heterogeneities is mapped using tracers. For a simple axisymmetric case the passive tracers can be advected forward or backwards in time. The results indicate that the bottom 200-300 km of the lower mantle are most efficiently sampled by a thermal plume, and that the heterogeneities are considerably stretched and mixed in the plume head. Further more realistic models include 1) a chemically denser recyled crust, 2) a more complex thermo-chemical boundary layer and 3) a background velocity field superimposed on the flow field associated with the mantle plume.

Factors controlling the asymmetry of lithosphere extension inferred from numerical experiments

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It has been argued that extension of the lithosphere may occur either in a pure shear mode or in a simple shear mode of deformation. The case for an asymmetric mode of extension has been made based on two types of observations. Low angle detachments, which have been active over a longer time interval, have been interpreted as whole lithosphere scale shear zones offsetting the areas of mantle lithosphere and crustal thinning. Strong asymmetry in crustal structure between the conjugates of passive continental margin systems has lead to interpretations of the conjugates in terms of an upper / lower plate geometry. We use 2D thermo-mechanical Arbitrary-Eulerian-Lagrangian finite element techniques to study factors controlling the asymmetry of lithosphere extension. In order to be able to model the interaction between lithosphere processes and the underlying asthenosphere, the models extend from the surface to 660 km depth. The upper surface of the model is free to move. Crust and mantle-lithosphere rheology are represented by a plastic failure criterion following Byerlee's law and temperature dependent non-linear creep, the asthenosphere follows a temperature dependent non-linear creep flow law. We focus on the role of plastic strain softening and on the effect of (de) coupling between crust and the mantle lithosphere, starting from a lithosphere with a "normal" crustal thickness. Our numerical model results indicate (at least) three modes of deformation, 1) symmetric extension, 2) (transient) whole lithosphere scale asymmetric extension, and 3) crustal asymmetry concomitant with a whole lithosphere scale symmetric style of deformation. Asymmetric extension with a fault zone continuing from the surface into the upper mantle lithosphere is promoted in models with a large ratio between the initial strength and the final strength of the strain softened brittle / plastic shear zone and by strong coupling between crust and mantle lithosphere. In these models asymmetry is a transient feature where the single fault zone is abandoned when viscous processes start to dominate and when the brittle shear zone is warped up into a less favourable position for further deformation. When the strength reduction upon "faulting" is lower, the coupled models show a fully symmetric style of extension, with a conjugated brittle fault system cutting the crust and upper mantle. When low viscosities in the lower crust facilitate decoupling between the brittle upper crust and the brittle upper mantle, decoupling of the "fault" zones in the lower crust results in an asymmetric crustal necking profile with strongly differing width and structure of the conjugated margins. The overall style of extensional deformation, however, is still dominated by pure shear deformation.

The Influence of Dynamic Recrystallization of Olivine on the Development of Seismic Anisotropy in the Mantle

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The interpretation of seismic anisotropy in the mantle requires a knowledge of the relationship between the lattice preferred orientation (LPO) of crystals and the convective flow field. In order to better understand this link, we present a model for the evolution of LPO in olivine aggregates that deform by both intracrystalline slip and dynamic recrystallization. Dynamic recrystallization depends on the dislocation density of the grains, which is a function of the applied local stress. Grains with a large density of dislocations lower their bulk strain energyby nucleating strain-free subgrains at a rate proportional to a dimensionless nucleation parameter λ^*. Grains with high energy are then invaded by grains with low energy by grain boundary migration, at a rate proportional to a dimensionless grain boundary mobility M^*. The value of λ^* is constrained by observed LPO patterns in experimentally deformed olivine aggregates,and M^* is constrained by the temporal evolution of the strength of the LPO. For $M^* = 75 \pm 25$ and $\lambda^* > 3$, the model predictions agree well with the experimental results. When an initially isotropic aggregate is deformed uniformly, the mean orientation of the a-axis first follows the long axis of the finite strain ellipsoid and then evolves by GBM toward the softest orientation. At larger strain, it rotates towards the long axis of the strain ellipsoid corresponding to infinite time (the "infinite strain ellipsoid" or ISE). The mean a-axis orientation will align with the flow direction only if changes of the ISE orientation along flowlines are sufficiently slow, i.e. if $\Pi \equiv |\mathrm{D}\Theta/\mathrm{D}t|/\dot{\epsilon}_o \ll 1$, where Θ is the angle between the flow direction and the long axis of the ISE. We predict that the best approach to interpret seismic anisotropy is the direct calculation of synthetic seismograms from 3D mantle flow models for which the LPO can be calculated using the present model.

Modeling Flow and Melting within Subduction Zones

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A number of models for flow and transport within the mantle wedge above subducting slabs have been developed based on geochemical data. Specifically, models have been advanced which appeal to a pre- existing condition for mantle material entering the wedge. Here melting and chemical depletion at a back-arc spreading center (BASC) is used to explain characteristically low concentrations of high field strength elements (e.g., Nb, Ti, Zr, Hf) in arc volcanics. Results are presented from two- dimensional numerical experiments which include both flow and melting of the upper mantle at subduction zones in order to test geochemically based models. Variable viscosity experiments incorporate effects of both a dynamic slab and back- arc spreading. Experiments characterize slab and wedge temperature distributions for a range in subduction and back-arc parameters including subduction rate, retrograde versus down-dip slab motion, viscosity structure and trench-back-arc separation distance. Experiments also include the generation and tracking of melt within the mantle matrix using distributed tracer particles. Both batch and fractional melting models are employed for extracting melts from beneath the BASC and within the arc. The average degree of depletion for material in the wedge is recorded over a range in subduction and extraction parameters. Interestingly, the largest depletion effect is seen for intermediate back-arc spreading rates. Models also include the role of melting and melt extraction on mantle viscosity within the wedge and related subduction evolution.

Subduction initiation: Weak fault by water?

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Subduction initiation remains one of the unresolved challenges of plate tectonics. Up to now fluid-dynamical approaches have been used in which the yield strength of the lithosphere has been assigned arbitrarily. Our analysis focuses on the solid mechanical processes prior to and during yielding; i.e. we investigate the formation of a weak zone, the proto-subduction zone. It is long known that the negatively buoyant oceanic lithosphere, once pushed down deep enough on such a weak zone can in a second finite amplitude instability overcome flexural bending and shear-resistance.

While realistic analytical, analogue and numerical models of subduction postulate the presence of a weakness, the problem of weak zone formation has not been looked into until very recently. In a first nonlinear elastic fracture mechanical approach we investigated whether thick sediment piles at passive margins on the old oceanic lithosphere can cause failure of the lithosphere. An earlier model study came to the conclusion that the old oceanic lithosphere should be strong enough to support such a weight. However, the oceanic lithosphere strength envelope since have been revised to much lower strength values.

Following our studies based on fracture mechanics we now present a coupled solid-fluid mechanical approach in which the laboratory creep laws are implemented directly. Our previous study focused on the effect of strain hardening, here we are able to investigate systematically the roles of creep rate sensitive, temperature, and water effects on weak zone formation. We use a finite element approach on a generic oceanic lithosphere with and without adjacent continental lithosphere. Thermal initial conditions are implemented analytically, using a cooling half space model with thermal parameters that give a statistical best fit of worldwide ocean floor topography as a function of thermal age. A skewed triangular sedimentary loading function is introduced as a nodal load on the continental margin. The peak sedimentation rate is set to 15cm/kyr over a time span of 100 Myrs. Sedimentary and water buoyancy responses to bending deformation are introduced by buoyancy forces applied to the bottom of the plane strain model where passive mantle flow mimicked by two families (z and x) of nodal nodal dashpots (10^{21} Pas) at the base.

1) Strain Rate weakening Our formulation provides a self-consistent, laboratory data based constitutive equation of a thermo-elasto-visco-plastic mechanical lithosphere (the solid layer) resting on a fluid dynamic rheological sublayer. In this

approach the boundary between the fluid and solid domains is clearly defined by the validity domain of the Peierls mechanism. In laboratory experiments the boundary was found to be about Tbase =1200K corresponding to fluid like deformation above and solid like deformation below this temperature. Geological extrapolations have been based on estimates of the thickness of the mechanical lithosphere from nonlinear flexure giving Tbase =1073 or 1173 K, respectively. 1200 K are given as an upper limit of shear localization in upper mantle peridotites. Our formulation embeds localized weakening in shear zones within the solid layer through the yield phenomenon of the Peierls mechanism. Above the yield stress the solid becomes weak and flow starts with a characteristic strain rate. Further strain-rate weakening due to enhanced shear localization results from both Peierls and power law nonlinear flow laws but not from the diffusion creep law. Weakening in the diffusion creep regime can, however, become important through grain size reduction effect after yield. Grain size reduction has not been considered here because it is an evolved weakening mechanism requiring large strain after subduction initiation.

2) Strain weakening Another weakening mechanism, which does not require an intermediate phase of power law creep for grain size reduction may however play a role. In a wet mantle free fluids can self organize into void sheets under the action of an applied shear stress. We have tested this approach and find that in a compressive environment the effect is negligible. Under global tension the weakening due to free fluids can become important. A tensile subduction initiation mechanism envisioned by refs is mechanically plausible with our rheology. An argument against a tensile mechanism is that passive continental margins are on a global scale under compression.

3) Wet versus dry lithosphere We come back to the wet versus dry rheology and investigate whether subduction initiation is possible under the combined action of sediment load and ridge push. Ridge push results from thermal expansivity and acts as a lateral push. The sediment load minus water depresses the oceanic lithosphere vertically on passive continental margins. The sediment load acts locally on the future subduction zone. Erosion on the continents furnishes a steady supply of sediments. Thus the sediment load can locally reach values, which are one order larger than ridge push. Sedimentary basins on continental margins are continuous but have varying thickness. Our model treats sediment basins as a line loads (plane strain model). We thus assume the continental margins have been uniformly filled. We first discuss the case of a lithosphere under an increasing sediment load only.

Both wet and dry lithosphere fail under the sediment load. However the style of failure is very different. In the dry lithosphere case, only the top 10 km (plus 10km which are not considered) fail in a solid mechanical manner while the lower part deforms in a diffuse way. Ultimately, the sediment load triggers a Rayleigh Taylor instability in the weak rheological sublayer which then delaminates from the mechanical lithosphere. Our cooling half space model thus develops dynamically into a cooling plate model between 10 and 100 Myrs.

The wet lithosphere case shows an entirely different behaviour. It fails on its whole mechanical thickness and subduction can initiate. The presence of water on the passive continental margin thus appears to be the first and foremost factor to control subduction initiation on the Earth.

Are Mantle Plumes Adiabatic?

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We have examined in a cartesian box with an aspect-ratio six The issue concerning the state of adiabaticity of mantle plumes. We have investigated in the quasi steady-state regime high Rayleigh number convection (Ra 10^7) . with both depth-dependent viscosity and thermal expansivity for both the Boussinesq and the extended Boussinesq approximations. We have also assessed the influence of various forms of thermal conductivity and internal heating. For a rapid local scanning of the conditions of adiabaticity, we have generalized the classical Bullen's parameter equation from one-dimension to multi-dimensions. This is analogous to what the oceanographers do with their construction of the local Brunt-Vaisaila frequency maps. For assessing the local state of adiabaticity inside plumes and in their surroundings, we have extracted from the local geotherms and the local thermodynamic properties the corresponding Bullen's parameter profiles and the 2-D maps portraying the state of adiabaticity in the mantle. Histograms characterizing the frequency of adiabaticity are also employed for quantification purposes. In general, superadiabatic thermal gradients are found inside the thick plume limbs and sometimes along the central part of the plume. The centers of plume heads are subadiabatic or nearly adiabatic, but the edges of the plume heads are strongly subadiabatic. Alternating strips of subadiabaticity and adiabaticity are found in the downwellings. The ambient mantle outside the plumes is generally adiabatic and is sometimes perforated with islands of marked deviations from adiabaticity.

Convective destabilization of a thickened continental lithosphere

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One consequence of orogenic building (by thrusting or homogeneous thickening) is to lead to a thickened lithosphere. Many studies have focused on consequences of convective removal of thickened lithospheric roots in terms of stress, strain, and metamorphism (Houseman et al., 1981; England and Houseman, 1989; Molnar et al., 1998). But first of all it seems important to understand how a thickened lithosphere reaches back equilibrium.

Two mains mechanisms may take place: 1) a large part, or the whole mantle lithosphere peels suddenly from the upper lithosphere by the propagation of a strongly localised shear zone, and sinks into the mantle (Bird (1979), Houseman (1996), Schott and Schmeling (1998)), or 2) a small scale convection progressively thins the lithosphere thanks to the development of instabilities due to lateral density contrast at the edge of the root and/or because of a density gradient at the base of the root (Doin et al., 1997).

We have used a 2D convective code to study the convective destabilization of a thickened continental lithospheric layer. The numerical simulations of convection include a Newtonian and a non Newtonian rheology with a viscosity depending exponentially on temperature and pressure, or following the Arrhenius law. The way the lithosphere reaches back equilibrium in 2D convective simulations can be fitted by using a 1D conductive CHABLIS modified model. This comparison shows that the thinning is triggered by a heat transfer due to small scale convection at the base of the lithosphere. The parameters controlling this heat transfer have been established by Dumoulin et al (1999). We show that after a thickening of the lithosphere, it slowly tends towards its equilibrium state (in about 100 Ma) compared to quickly destabilization (in about 10 Ma) obtained by Houseman et al. (1981) (in simulations involving an isoviscous fluid, 10^{21} Pa.s) and Marotta et al. (1999) (in simulations with a weak temperature dependence of the rheology). We have also studied the influence of the thickening factor and the thickened zone width, using the 2D convective code; we evaluate a limit width for which the main mechanism of erosion is no longer sublithospheric but is controlled by small scale convection on the edges of the root.

Dynamics of Slabs: Insight from Numerical Eexperiments

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We analyze the long-term dynamics of visco-elasto-plastic slab falling into a passive uniform or viscous stratified mantle in a self-consistent way by means of the results of numerical experiments with the aim to clarify the dynamics of trench retreat and to identify the factors that influence the process. We perform a comprehensive analysis of slab rheology and influence on slab dip and trench retreat. Particular emphasis is placed on the interaction with the 670 km discontinuity.

Modeling Archean sub-continental mantle dynamics and heat flow using diamond stability constraints

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Models for the thermal history of the Earth have, historically, frequently ignored a conflict in geological observations concerning the Archean thermal regime. The Archean continental thermal regime was, in most places, relatively mild in comparison to the higher upper mantle temperatures inferred for the Archean. Recent numerical experiments (eg. Lenardic 1998) involving simple convecting systems have provided a physical basis for the postulation that this inconsistency can be resolved if the oceans carried a greater proportion of the global heat flux in the Archean.

A particle-in-cell finite element code (Moresi et al, 2001) was employed to extend realistic convection models of the modern Earth (Lenardic et al., 2000) to Archean conditions. The models incorporate a strongly temperature viscosity, yield criterion for the lithosphere which allows a plate- like mode of convection to develop, and chemically distinct continental crust. A mode of convection with mobile oceanic regions and continents with relatively thick thermal roots is stable up to the high Rayleigh numbers (a measure of convective vigour) applicable to the Archean. Given this observation, the increase in the partitioning of heat flux implied for simple Archean convection models, similarly occurs for systems capable of generating plate-like motion. This result is robust under a wide range of parameter variations. The continental crust modulates the thermal conditions of the mantle, extending the depth of the thermal boundary layer beneath the continent, thus providing a mechanism for stabilising the sub-continental thermal field. This effect can explain the primary features of the Archaean thermal record.

Constraints provided by Archaean diamonds not only requires that the conditions necessary for diamond stability existed in the Archaean continental lithosphere, but also that those conditions have remained relatively unperturbed for 3Gyr (eg. Boyd et al., 1985). We have examined the length of time the sub-continental mantle lithosphere spends in the diamond stability field for a number of convection models which satisfy the Archean heat flow partitioning contraints in the mean. While such systems can account for the provisional existence of Archaean diamonds, they do not directly explain their longevity even in the centre of a craton. We therefore postulate that the thermal stability required for the preservation of Archaean diamonds can only be provided if the cratonic roots are also chemically stabilised against convective recycling.

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Towards a consistent thermo-mechanical model of plate tectonics for oceanic-lithosphere steady flows

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Thermal mantle convection is often put forward as the driving mechanism of plate tectonics. However, the nature of the coupling between the lithosphere and the underlying mantle is still not clearly understood. To this end, we are aiming to develop an analytical theory plate tectonics, allowing coupled treatment of the thermomechanics of both regions and incoporating rock constitutive models. The goal of our poster will be to illustrate such an analysis in the particular case of the seafloor-spreading dynamics.

In our analysis, as intuition and previous works suggest, seafloor spreading is accompanied by cooling of the upwelling flow of hot mantle as it moves towards the surface and separates on either side of the stagnation point corresponding to the ocean ridge. No a priori distinction is made between the lithosphere and underlying mantle, both being modelled as having the same homogeneous composition. The distinct properties of the two regions appears later as natural consequence of the model. A plastic/viscoplastic incompressible constitutive relation is used, based on the rheological profile and whose behaviour is either brittle and pressure dependent, or ductile and strongly temperature sensitive. Thus, expressing mechanical equilibrium and advection-diffusion of temperature, together with boundary conditions at the free interface with the water, a complex problem which couples flow, stress and temperature fields is obtained.

Approximate solutions of this thermo-mechanical problem can be constructed using a thermal boundary layer hypothesis suggested by the order of magnitude of the Péclet number. This non-dimensional number is defined as $Pe = U\,L/\kappa$, where κ, U and L are respectively the thermal diffusivity, velocity and length scales. Choosing geodynamical orders of magnitude, i.e. for U the surface velocity and for L the distance x to the ridge, it is found that $Pe \gg 1$.

As a result, the flow can be divided into at least three regions:

1. a thin thermal boundary layer (the lithosphere) whose thickness is of order $Pe^{-1/2}\,x$ and in which variations take place over horizontal distances much greater than vertical ones;

2. an external region beneath the thermal layer, characterised by horizontal and vertical length scales of the same order (x) and in which thermal diffusion is negligible;

3. and finally, a comparatively small region near the ridge in which the flow description is more complicated and a numerical approach would be needed.

As will be explained in the poster, the existence of two different length scales in the boundary layer implies that its dynamics is controlled by the horizontal strain rate $D_{xx}(x)$. Knowledge of D_{xx} determines indeed the temperature profile, the overall horizontal force acting on the layer and the position of brittle-ductile transition. By matching stresses and velocity with the external region, D_{xx} is determined at any distance x. As a result, one obtains non-local, nonlinear relations between D_{xx} and the shear stress just outside the layer. In principle, these provide boundary conditions for the flow external to the layer expressing the coupling between the external and surface-layer problems.

In attempt to obtain a consistent solution of the coupled problem, we consider ductile flow at constant temperature T_0 in the external region. Assuming different constant surface velocities on the two sides of the ridge, a hypothesis which is justified a posteriori, a self-similar external solution is obtained. The shear stress calculated from this solution is then used to determine the D_{xx} from the boundary layer problem. It is found that, at high enough external temperatures, the variations in surface velocity are indeed negligible. Significantly, it is found that the upwelling of the mantle in the external region is driven by the motion of the surface (plate), rather than the converse, and that the flow should be symmetric about the ridge. Apart from material properties of the rocks, the solution is uniquely determined by the surface velocity at the ridge and the upwelling temperature T_0, parameters which may depend on the particular ridge considered. Thermal definition of the lithosphere is then upheld and the plate imposes a constant-velocity boundary condition on the external convection.

Validity of the above description supposes sufficiently high T_0, otherwise we find rapid and significant variations of the surface velocity with x, rendering invalid both the assumption of constant surface velocity in the external problem and that of slow horizontal variations in the boundary layer. Furthermore, order of magnitude estimates suggest that the supposed incompressibility of the flow is questionable. Both points will be addressed in the poster and suggest directions for future work.

Comparing global mantle models: from mapping to hypotheses testing

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Seismic tomography shows that subducted oceanic lithosphere deforms in a complex variety of ways as it descends through the mantle. To understand better the mechanisms of this deformation, I study the dynamics of thin viscous sheets of arbitrary shape subject to arbitrary loading. An analytical "shallow sheet" scaling analysis reveals that two distinct types of deformation can occur, depending on the sheet's principal curvatures and the wavenumbers of the applied load: an "inextensional" (bending) mode and a "membrane" (stretching) mode. In general, high curvature favors membrane deformation, whereas low curvature favors inextensional deformation. By using the scales revealed by the shallow-sheet analysis together with asympotic expansions in powers of a small slenderness parameter, I reduce the three-dimensional viscous flow equations to a set of equivalent two-dimensional equations for the velocity of the sheet's midsurface. General nonorthogonal coordinates are employed, which allows the use of a Lagrangian grid that deforms with the sheet. The set of "thin-sheet" equations is completed by kinematic evolution equations for the sheet's shape (metric and curvature tensors) and thickness. Numerical solutions of the equations for gravity-driven deformation show that the inextensional mode has a boundary-layer character, and that compressive membrane states are inherently unstable to buckling instabilities. A model for a two-dimensional viscous jet falling onto a rigid plate exhibits steady periodic folding, the frequency of which varies with the jet's height and extrusion rate in a way similar to that observed experimentally. A simple model for subduction, in which a thin sheet extruded horizontally deforms under its own weight, exhibits "trench rollback" at a rate comparable to those observed on earth.

Role of the mantle dynamics on water distribution in the Earth's mantle

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The effects of the dynamics on water distribution in the Earth's mantle have been investigated in a 2D cartesian box. The model takes in account water partitioning between mantle transition zone and upper and lower mantle (i.e between resp. Olivine-Spinel and Spinel-PostSpinel). The momentum equation is expended in harmonic series along horizontal direction and solved by finite difference along the vertical direction. Water and Temperature fields are solved using finite differences and ADI sheme. We have also modelled water extraction at the ridge and reinjection at the trench.

First results suggest that the importance of storage in the transition zone is driven by the diffusivity of the water. For characteristic temperature of the transition zone (1400 ^{o}C), intra-cristalin diffusivity (D_{H_2O}) is about 10^{-8} m^2/s. However, the diffusion along the grain boundaries, could be two orders of magnitude higher (i.e 10^{-6} m^2/s). This extremal values correspond to opposite situations. When diffusivity is high, a reservoir can set in the transition zone. Otherwise, the water concentration of the mantle remains homogen, mixed by the dynamics.

Thermochemical Convection and He Concentrations in Mantle Plumes

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Mantle convection is still a subject of controversy since geochemistry indicates the existence of at least two chemically different reservoirs isolated over billions of years, while seismic tomography shows that slabs penetrate in the lower mantle. Recent tomographic studies suggest the existence of chemically denser material at the bottom of the lower mantle. The presence of such a layer, if primitive, may reconcile geochemical and geophysical observations.

We use a 2D numerical convection model with a chemically denser layer at the bottom of the mantle and we vary the ratio of chemical to thermal density contrast ($\Delta\rho_c/\Delta\rho_{th}$). We find that for $\Delta\rho_c/\Delta\rho_{th}$ greater than 0.5, convection takes place in two units separated by an intermediate thermal boudary layer. This thermal boundary layer forms a topography which could make its seismic detection difficult. Moreover, the chemically denser layer remains stable and poorly mixed for ages comparable to the age of the Earth.

To investigate if such configuration could explain both the undegassed nature of plumes and the presence of recycled oceanic lithosphere in the source of OIB, we use a numerical model which includes: 1) active tracers to study the dynamics of a primitive and chemically denser layer at the bottom of the lower mantle, and 2) passive/active tracers to follow the subduction of oceanic crust and of the depleted oceanic lithosphere. The model has temperature dependent viscosity, depth dependent thermal expansion coefficient and includes phase transitions. Internal heating depends on local concentrations of radioactive elements producing internal heating (^{238}U, ^{235}U, ^{232}Th, ^{40}K). The model calculates the time evolution of the ^{4}He and ^{3}He concentrations in the mantle, in the denser layer and in the subducted material. We then focus on the ^{4}He and ^{3}He concentrations in ascending mantle plumes.

On the importance of compressibility in D”-layer dynamics

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The entrainment dynamics in the D”-layer are influenced by multitudinous factors, such as thermal and compositional buoyancy, and temperature- and composition-dependent viscosity. Here we are focusing on the effect of compositionally dependent viscosity on the mixing dynamics of the D”-layer, arising from the less viscous but denser D”-material. The marker method, with up to ten million markers, is used for portraying the fine scale features of the compositional components, D”-layer and lower mantle. The D”-layer has a higher density but a lower viscosity than the ambient lower mantle, as suggested by melting-point systematics. Results from a 2D-FD numerical model including the extended Boussinesq approximation, and therefore accounting for the compressibility of the mantle material, show that a D”-layer that is less viscous than the ambient mantle by 1.5 to 2.4 orders of magnitude, can NOT efficiently mix with the lower mantle, even though the buoyancy parameter is as low as Rrho = 0.2 to 0.3. This is in contrast to laboratory experiments, that overestimate the Rrho needed to stabilize the D”-layer at the CMB, when e.g. water is used to model convection. However, very small-scale Schlieren of D”-layer material are entrained into the lower mantle. These small-scale lower-mantle heterogeneities have been imaged with 1-D wavelets in order to delineate quantitatively the multiscale features. They may offer an explanation for small-amplitude seismic heterogeneity inferred by seismic scattering in the lower mantle. This particular type of mixing dynamics transforms to a kind of “lava lamp” mode, when mechanical heating is neglected and the buoyancy parameter is kept the same. Preliminary results suggest, that this is due to the convection velocities being overestimated, when adiabatic heating/cooling is not taken into account in the non-dissipative (Di=0) case.

Movies can be found at http://www.geo.uu.nl/∼bert/DDP

Spectral modelling of mantle convection in a non orthogonal geometry: Application to subduction zones

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A two-dimensional numerical convection model is presented, in a parallelogram-shaped domain adapted to the geometry of subduction zones. The Navier-Stokes problem is solved by a pseudo-spectral solver (named projection-diffusion) coupled to the Richardson iterative scheme. In the convective domain, representing the upper mantle, plate tectonics are prescribed by imposing the velocities on the boundary. In particular, the subducting plate, located on a lateral side of the box, moves and dips at constant velocity. The model is validated for the zero obliquity configuration, first in the case of free boundaries, then in the case of a rigid upper boundary and, at last, in the case of both a rigid upper and lateral boundary. Then, we report preliminary results of experiments for a non zero obliquity at a Rayleigh number value of 104, a motionless overriding plate, subduction velocities smaller than 4 cm/yr and subduction dip angles higher than 60. Concerning the evolution of the flow and thermal structures with the subduction movement and geometry, the model predicts the occurrence of a recirculation process, characterised by the formation of an extra convection cell. This recirculation starts at a critical subduction intensity depending on the subduction geometry. The lower the subduction dip angle, the lower the critical subduction velocity. A critical subduction geometry is also suggested since the extra cell exists even at very low subduction velocities for dip angles lower than 65.

Secondary Instabilities developed in Upwellings of high Rayleigh number Convection : Implications for Hotspots

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1 Abstract.

Thermal plumes are ubiquitous at high Rayleigh number convection. The idea of secondary instabilities developed from plumes has come from recent high Rayleigh number simulations by Vincent and Yuen (2000). There is also evidence from seismic tomography (Bijwaard and Spakman, 1999, Goes et al., 1999) to suggest that mantle plumes are able to branch to some extent. Skilbeck and Whitehead (1978) and Whitehead (1982) have performed laboratory experiments showing the formation of secondary instabilities from a tilted plume in a set-up in which the conduits were made by injecting oil into a more viscous heavier liquid. The purpose of this work is to show how these secondary instabilities can develop as a consequence of the interaction of the shear flow developed by the large-scale circulation and the rising plume. This will be conducted within the framework of a constant viscosity fluid because of the need to understand the fundamental physics and the parameter space in Rayleigh number before pursuing other types of rheology. First, we will employ a two-dimensional axisymmetric spherical-shell model of Moser et al. (1997) within the framework of a Boussinesq fluid. When scaled to the Earth's mantle, the computational domain of this model will have an aspect-ratio of around six. A large enough aspect-ratio is needed to generate a potent enough large-scale circulation, which is needed to bend the plumes sufficiently for inducing the secondary instabilities found in the Whitehead experiments. Second, for comparing with 3-D situations, we have used a three-dimensional Boussinesq model taken from Dubuffet et al. (2000) with a large-enough aspect-ratio of $5 \times 5 \times 1$.

In the case of constant viscosity, the nature of these secondary instabilities branching off from primary plumes requires high-resolution, because what we are trying to detect represent secondary spatial bifurcations from an already developed primary hydrodynamic instability endowed with already a boundary layer character. Therefore, we would expect length-scales to be of the same magnitude or even smaller than the boundary-layer flows at the Rayleigh number being investigated. These instabilities develop at Rayleigh numbers for 2-D case at around 3×10^7 and in 3-D in excess of 10^8. Hence, very high spatial resolution are needed for 2-D at least 150 second-order, equally spaced finite-difference points along the vertical direction and for the more difficult 3-D situation at least 200 points along the vertical.

Fig. 1 shows a cartesian rendering of the 2-D simulations for six different Rayleigh numbers ranging from 3×10^6 all the way to $Ra = 10^{10}$. A fourth-order finite-difference scheme with equally spaced points have been used with grid points of 750 points in the radial and 4000 points along the tangential directions used for $Ra = 10^{10}$. We can observe that at $Ra = 3 \times 10^7$ there are signs for plume branching, whereas for the lowest Ra of 3×10^6, the upwellings more or less upright and do not suffer any bending from the large-scale shear flow. Thus this secondary plume bifurcation takes place at a Ra a little bit higher than 10^7 in 2-D. As Ra moves above 10^8, the tendency to plume branching increases and is accompanied by multiple foldings of the upwellings (see $Ra = 10^9$). These represent secondary instabilities emanating from primary upwellings, reminiscent of the instabilities found by Whitehead in his laboratory experiments.

We emphasize that in contrast to Whitehead's experiments which were conducted in a kinematically constrained situation and with two different types of fluids, the instabilities obtained here come from a thermal convection simulation with a homogeneous fluid. At higher Ra, above 5×10^9, the folding instabilities disappear as the upwellings become extremely thin and finally a quasi-layered regime emerges with small plumes hovering close to the thermal boundary layers for $Ra = 10^{10}$. This situation with separate convective systems at the top and bottom of the layer is similar to the results obtained for thermal convection at finite Prandtl number (Vincent and Yuen, 2000).

We have also investigated the three-dimensional situations in which plumes can be bent significantly and can develop secondary instabilities. For lower Rayleigh numbers we have gone up to 5×10^7 in the aspect-ratio $5 \times 5 \times 1$ box and did not find any noticeable signs of plume bending. Then at $Ra = 10^8$ there appears a bifurcation in the flow pattern and plumes are found to be bent severely by the large scale circulation produced at this high Ra and have gone to $Ra = 5 \times 10^8$ for further verification. We have used up to $1025 \times 1025 \times 257$ points in this simulation. Figure 2 shows a 3-D isosurface rendering of the $T = 0.55$ surface. We see clearly the distinct presence of a long snake-like structure which is caused by severe bending of a plume. It has clearly a complicated curved 3-D structure. A two-dimensional cross-section of the thermal fields (here the red denotes hot temperature between 0.6 and 1.0) is shown in Fig. 3. The plumes outlined in the 2-D cross section look strikingly similar to the 2-D upwellings observed in Fig. 1 for $Ra = 3 \times 10^7$. If this trend can be extrapolated upwards in Ra, then we may expect some sort of layered convection to take place in 3-D configuration for Ra between 3×10^{10} and 10^{11}. However, such high Ra calculations are slightly beyond the reach of massively parallel computers today, as they will require at least $5000 \times 5000 \times 1000$ grid points.

We have demonstrated here within the framework of a constant viscosity model that the secondary Whitehead instabilities can develop in a self-consistent manner in both 2-D and 3-D large aspect-ratio convection with bifurcation Rayleigh numbers of $O(10^7)$ and $O(10^8)$ respectively. These results would suggest the possibility of these secondary instabilities to arise under lower Rayleigh number conditions, since it has been shown by Malevsky and Yuen (1993) that non-linear (non-Newtonian) rheology can lower the threshold Ra for secondary instabilities to develop by at least an order in magnitude or a factor of two lower in the surface Nusselt number. Since plate dynamics involves a highly nonlinear plastic rheology, we surmise that such secondary plumelike instabilities can be developed in the upper mantle as a consequence of the interaction between the shear flow generated by plate tectonics and plumes generated from the 660 km boundary (Turcotte and Allegre, 1985, Cserepes and Yuen, 2000). These secondary instabilities can explain many hotspots without a definite age-progression trend as in the western Pacific (Mc Nutt et al., 1997).

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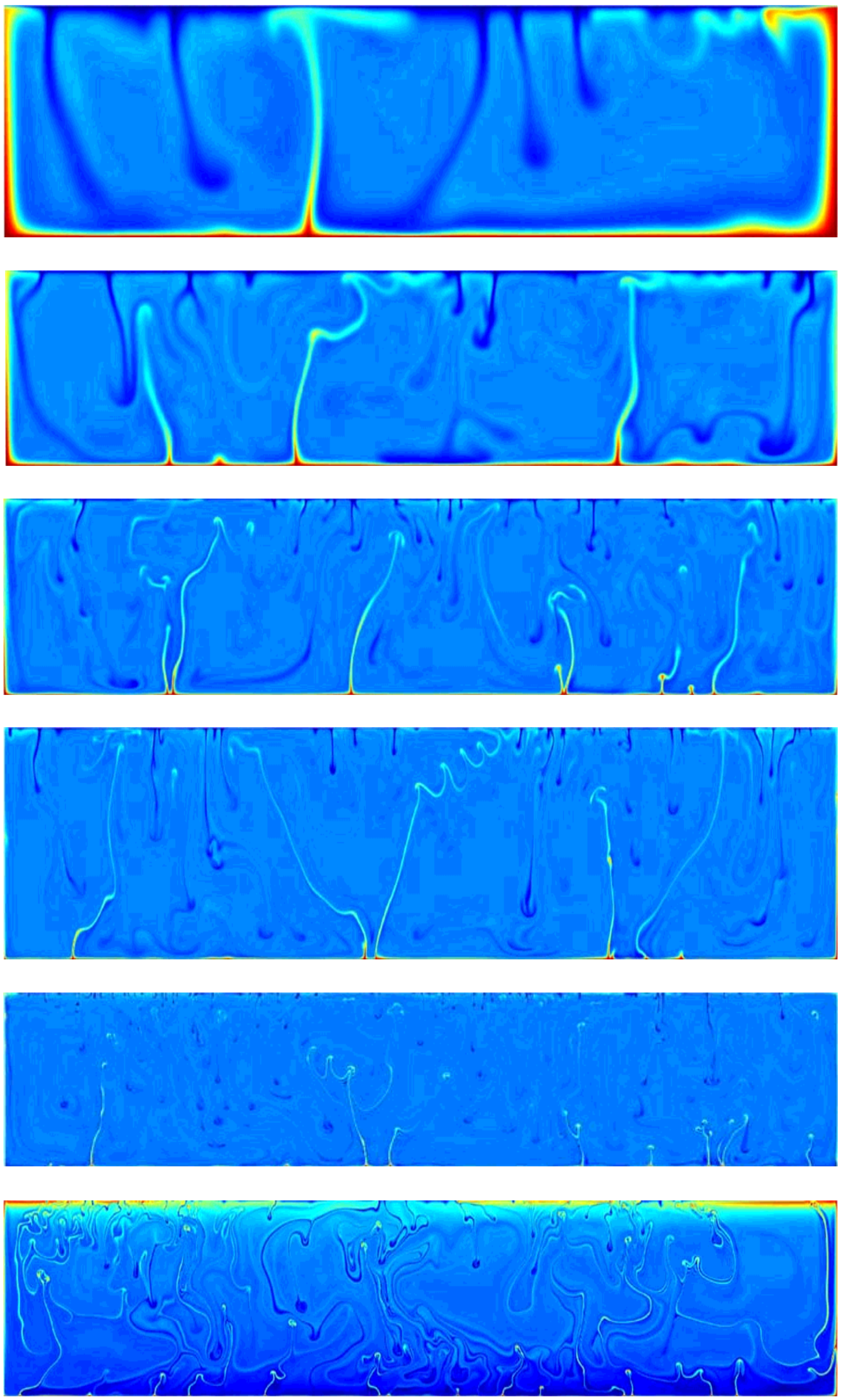

$Ra = 3 \times 10^6$

$Ra = 3 \times 10^7$

$Ra = 3 \times 10^8$

$Ra = 10^9$

$Ra = 5 \times 10^9$

$Ra = 10^{10}$

Fig. 1. Temperature fields of 2-D numerical simulations with purely basal heating.

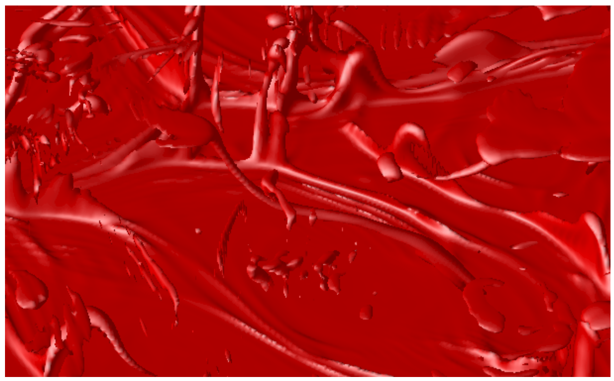

Fig. 2. Temperature isosurface of T =0.55 in 3-D convection with $Ra = 10^8$ in an aspect-ratio $5 \times 5 \times 1$ box, with unity being the depth.

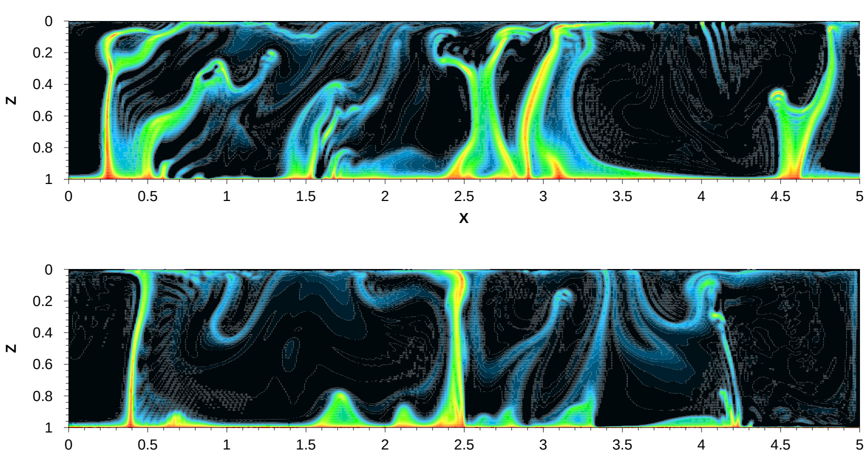

Fig. 3. Two-dimensional cross section of 3-D temperature field of $Ra = 10^8$ in large aspect-ratio box. Red color shows all T between 0.6 and 1.0.

Geophysical Application of Filtering and Visualizing Using Multidimensional Wavelets

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Wavelets are linear mathematical transformations (e.g. Resnikoff and Wells, 1998) which can analyze both temporal signals and spatial images at different scales. The wavelet transform is sometimes called a mathematical microscope. Large wavelets give an approximate image of the signal, while smaller and smaller wavelets zoom in on small details. Until recently most of the applications of wavelets in geophysics have been focused on the use of one-dimensional wavelets to analyse time-series of the Chandler wobble or one-dimensional spatial tracks, such as topography and gravity anomalies. Recently fast multi-dimensional wavelet transforms(Bergeron et al., 1999, 2000 a and b, Yuen et al., 2000) , based on higher-order-derivatives of the Gaussian function , have been developed and they have allowed us to construct rapidly two- and three- dimensional wavelet-transforms of geophysically relevant fields , such as geoid anomalies, temperature-fields in high Rayleigh number convection , and mixing of passive heterogeneities. We have also computed with wavelet transform the correlation of high-resolution seismic tomography between two depth levels and have constructed a two-dimensional map showing the correlation of seismic tomography for the transition zone under Europe. With wavelet transform one can view the character of a field at a particular length-scale. This continuous transform is different from orthogocal wavelet functions, which locks on to a particular hierarchy of length-scales. like a lattice. We have constructed scalograms of the two-dimensional wavelet-transformed quantities of the geoid anomalies, temperature fields in thermal convection and the scalar field depicting the hetereogeneities in mixing driven by thermal convection. With these scalograms one can discern the detail changes for the different length scales ranging from long wavelength to short wavelength , ranging from about one-tenth to one-hundredth of the entire domain . We have also constructed two proxy field variables, which are associated with the maximum of the wavelet transformed quantity, E-max, and the associated local wavenumber of the wavelet, k-max. Both E-max and k-max are functions of the spatial variables and their peak functional values indicate places with sharp variations in the field values, be it the geoid, seismic velocity anomalies or the thermal fields in convection. Both mixing dynamics and thermal convection in high Rayleigh numbers display different behavior at different length-scales, as revealed by the wavelet-transformation. Long-wavelength anomalies are mixed much faster than intermediate wavelength anomalies. The thickness of the thermal boundary layer depends also on the length-scale of the wavelet interrogation. In the geoid the ridges and subduction zones start to be discernible at around 400 km wavelength scale wavelet . Their appearance has been verified with smaller-scale wavelets down to a wavelength of around 120 km. Many spherical harmonics are required to mimick the feature capturing capability of a single wavelet at a particular scale in the geoid problem for order of the spherical harmonic going out to 256. We have compared geoid results from using a single wavelet and from a nonlinear band-pass filter in the spectral space. We have also developed analytical spherical wavelets , which involves a Bessel function and a Gaussian function, with the arc length being the argument. These spherical wavelets have been applied to decomposition of the geoid anomalies.

The two-dimensional correlation map of the seismic tomography shows that between a depth of 500 and 600 km under Europe there is an ellipse- shaped object with an area of 2000x 4000 km having a strong correlation for length scales of around 400 km. From the wavelet correlation spectra we can extract an horizontal length scale of around 100 km, which may be related to the interaction of the subdcuted material with the ambient mantle. The correlation results suggest that the thickness of the recumbent fast (cold) material is between 100 and 150 km. We hope to show here the possibilities of using multidimensional wavelets in looking at the multiscale features of several interesting geophysical problems.

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