Seismic Anisotropy and the Lithosphere Asthenosphere Boundary in Oceanic Basins

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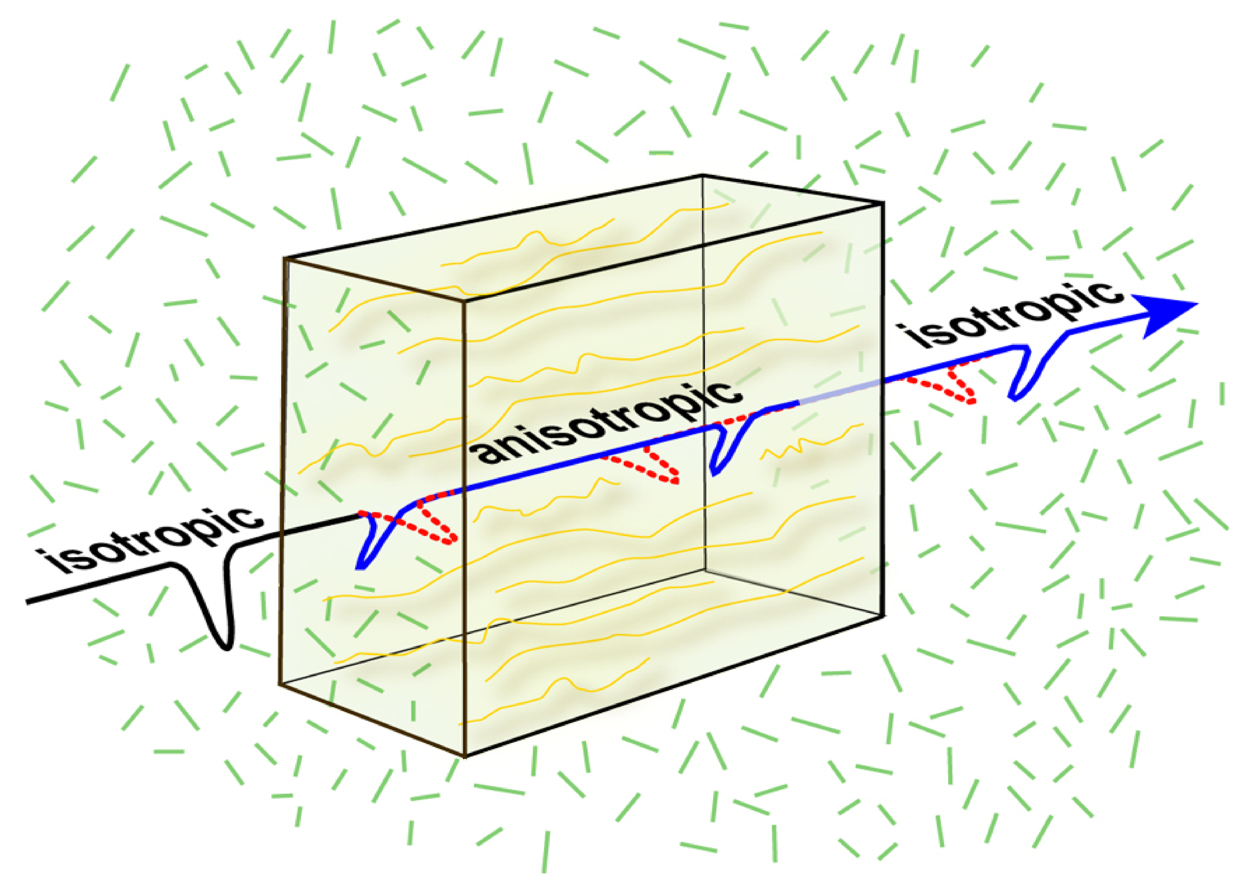
What is Seismic Anisotropy ?

Shear wave splitting in anisotropic media



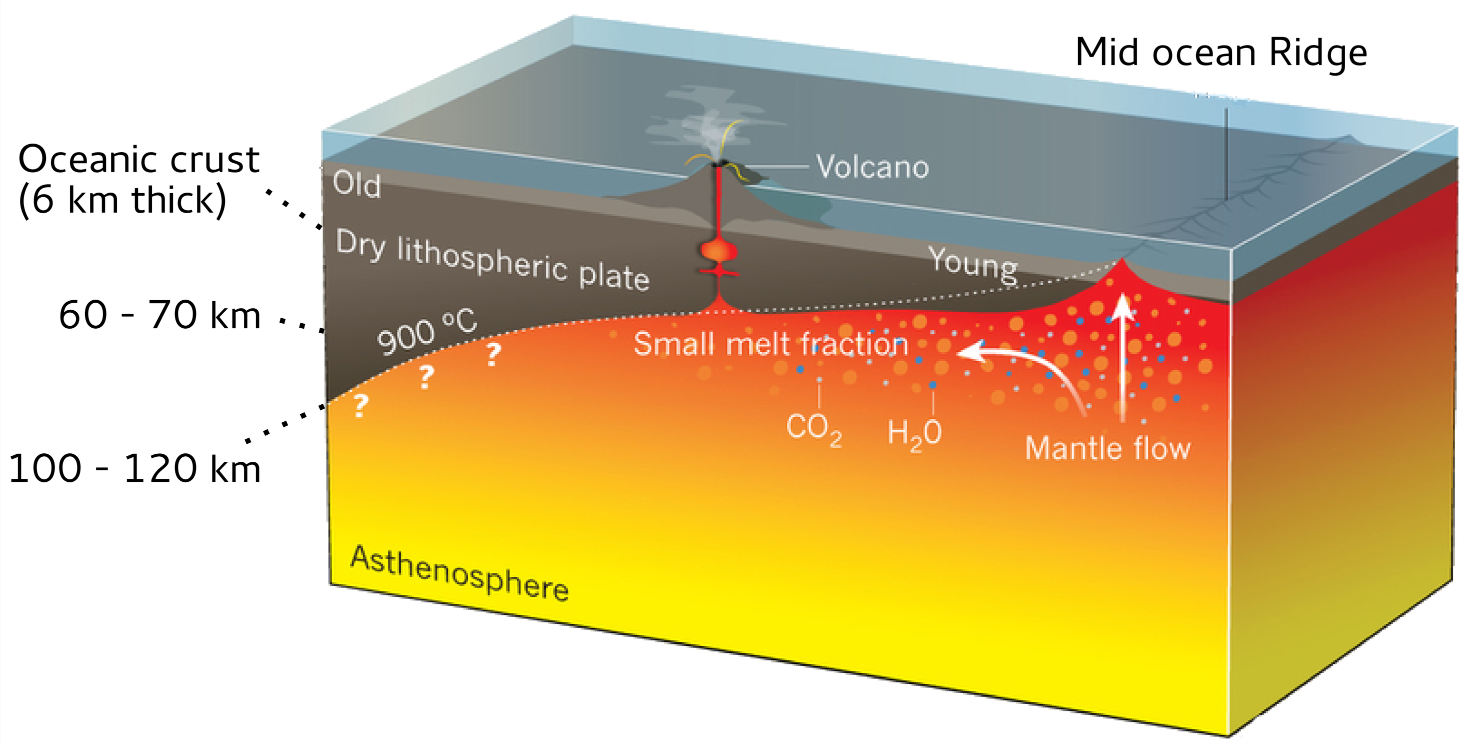
(After Crampin, 1981)

Radial anisotropy: $\xi = \frac{V_{SH}^2}{V_{SV}^2}$ (V_{SH}: Horizontally polarized shear wave; V_{SV}: Vertically polarized shear wave)

Fast axis direction of azimuthal anisotropy: ψ

The Lithosphere Asthenosphere Boundary

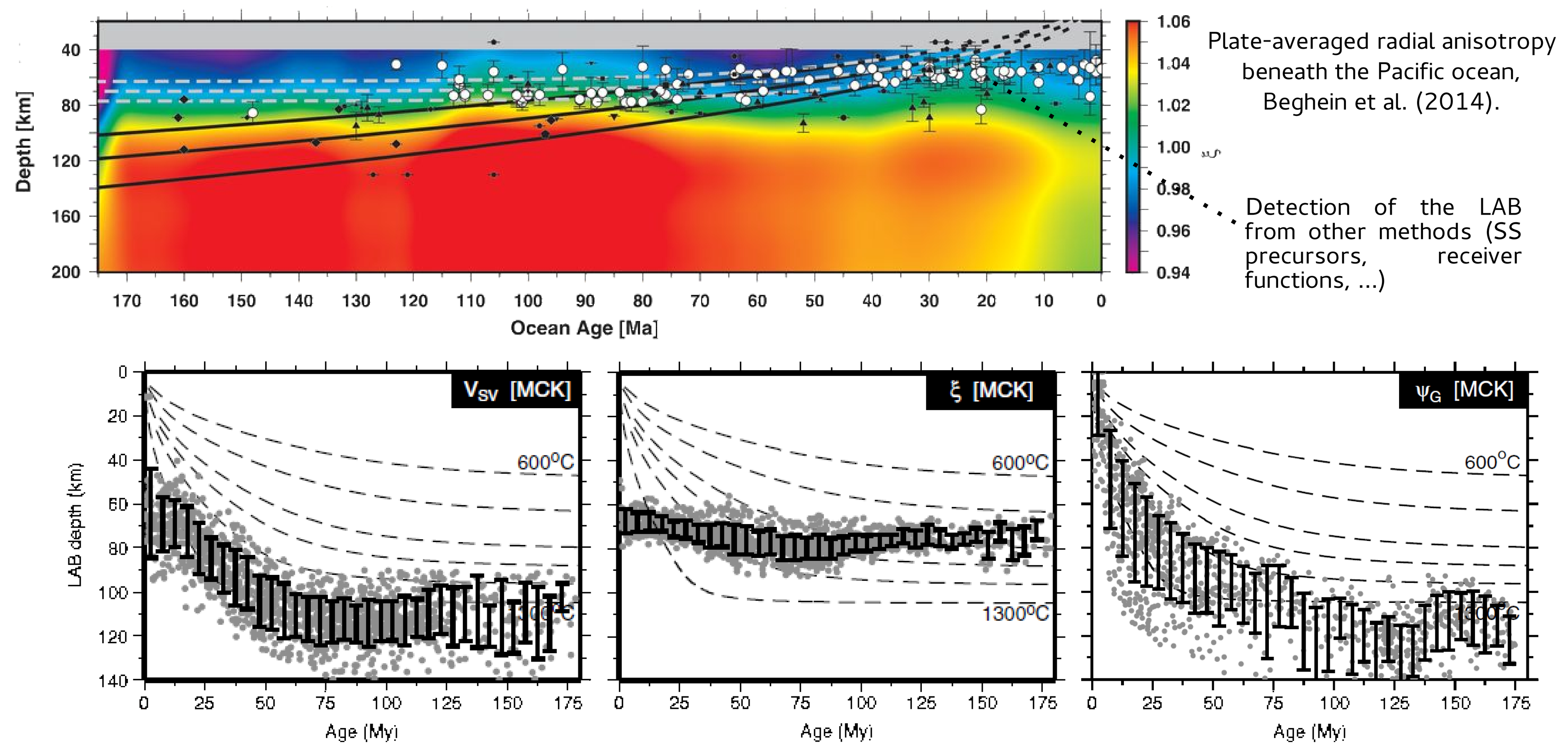
Near the boundary between the lithosphere and asthenosphere (LAB) of oceanic basins, seismic tomography models observe a drop of shear waves velocity, called the G discontinuity, at a depth increasing with plate age. Two main explanations have been proposed : presence of partial melt and influence of water.



A conceptual model of Earth's oceanic lithosphere and asthenosphere from Evans R.L. (2014)

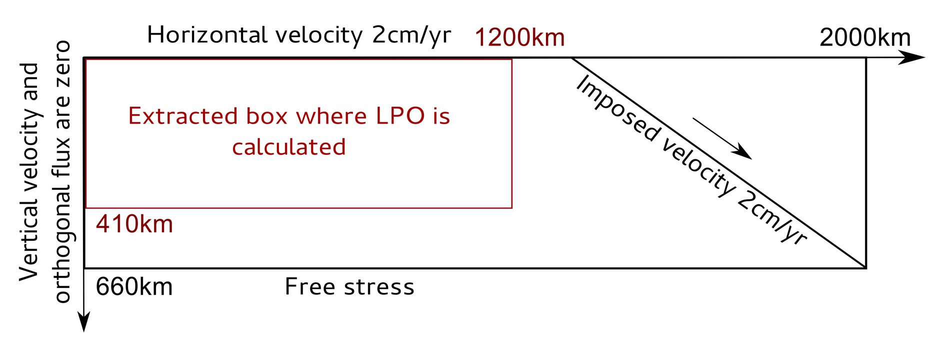
Motivations

The LAB induces also changes in seismic anisotropy. In the asthenosphere, the azimuthal fast axis is oriented subparallel to the plate motion, while in the lithosphere, the fast axis is 'frozen' subparallel to paleo-spreading directions, with a boundary roughly following the G. Conversely, radial anisotropy show an enigmatic strong vertical gradient, at a depth of about 70 km, regardless of plate age. The physical relating between this horizontal discontinuity and the LAB, if any, is largely unknown.



LAB depth from different proxies , using surface wave data, Burgos et al. (2014). Left: largest drop in SV waves velocity, Atlantic ocean; middle: largest increase in radial anisotropy, Atlantic ocean; right: change in correlation between the seismic fast direction and plate motion, Pacific Ocean. Dashed lines represent the McKenzie plate model.

The Geodynamic Model

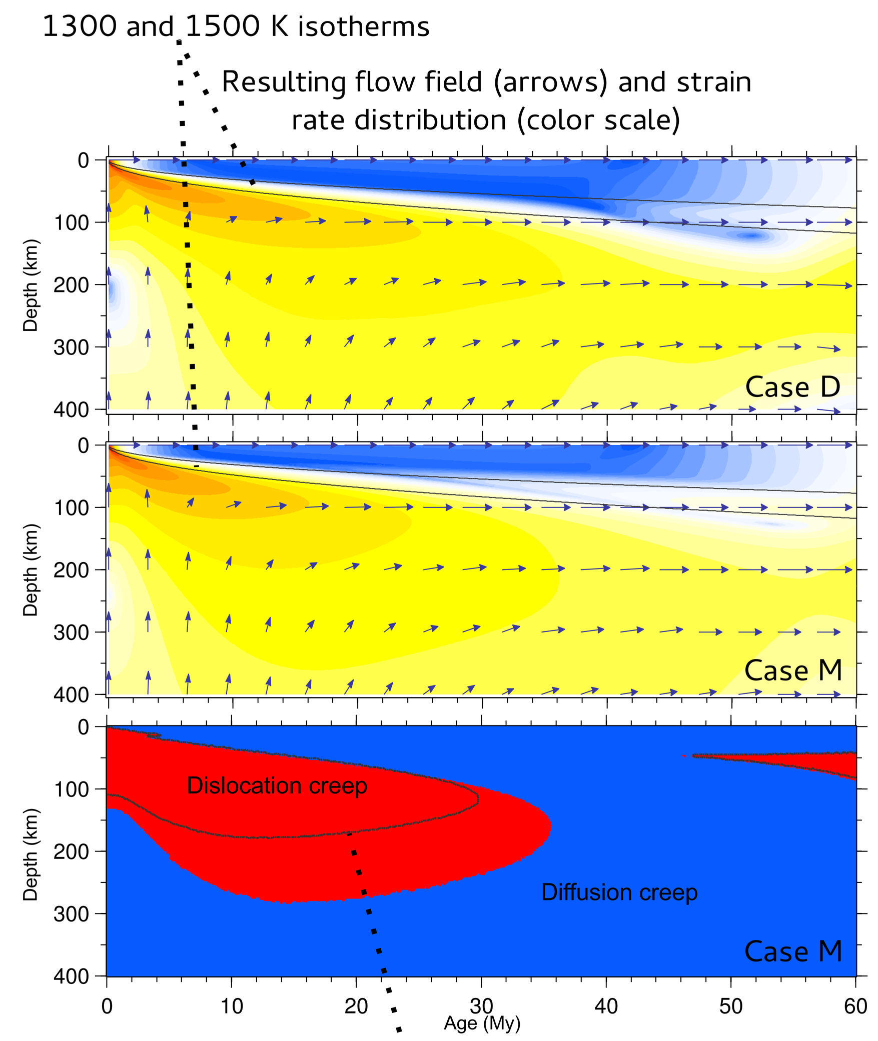


The conservation of mass, momentum, and energy equations are solved for an 2D incompressible Stokes fluid, under the Boussinesq approximation, using the finite-element code Fluidity (Davies et al., 2011).

Case D: deformation under dislocation creep only
Case M: mixed rheology , with dislocation and diffusion creep. The effective viscosity is :

$$\mu_{eff} = min(\mu_{disl}, \mu_{diff})$$

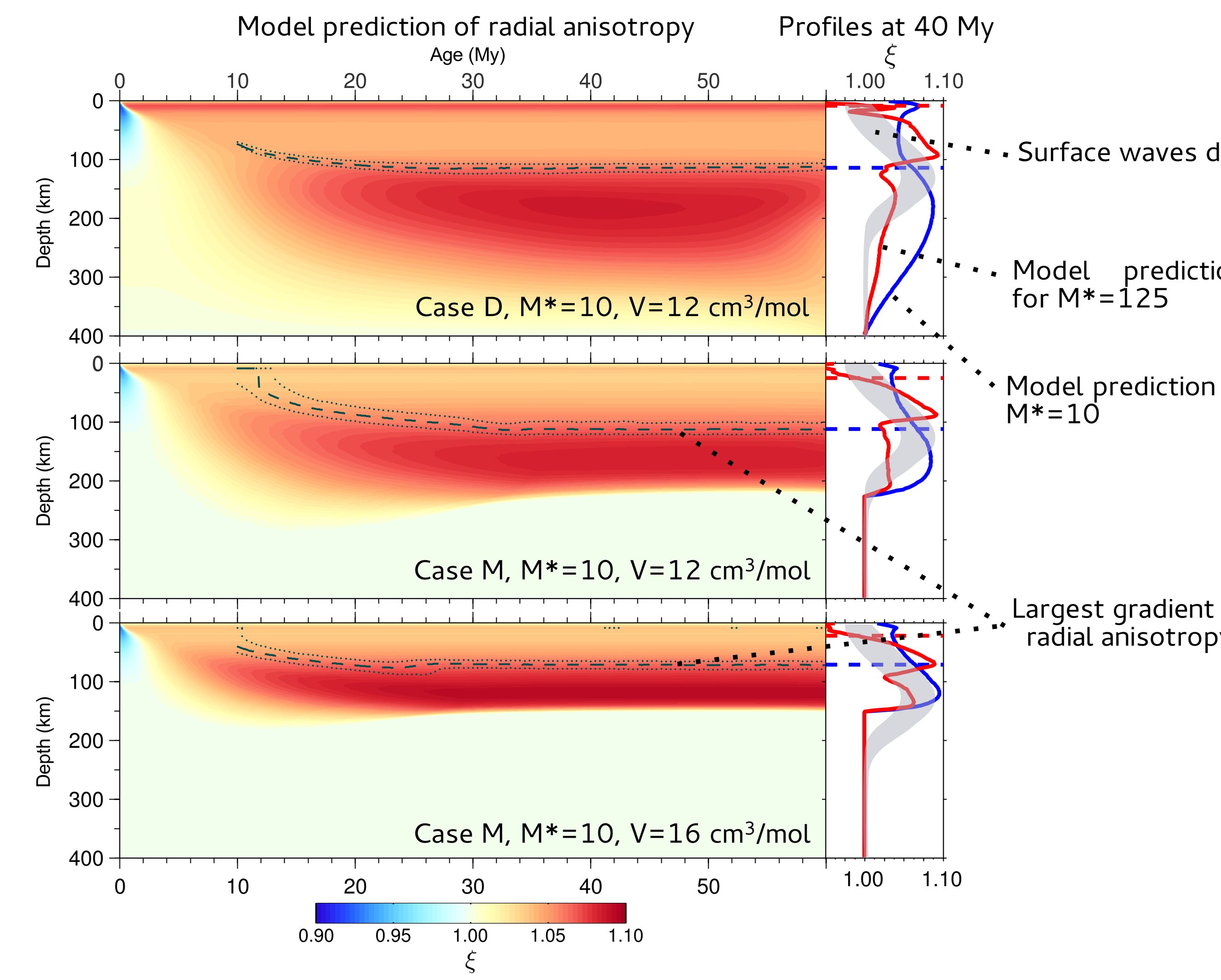
To calculate the crystal lattice preferred orientation (LPO) in the dislocation creep regime, we use the model D-Rex (Kaminski et al., 2004). In the diffusion creep regime, all crystals follow the fluid deformation.

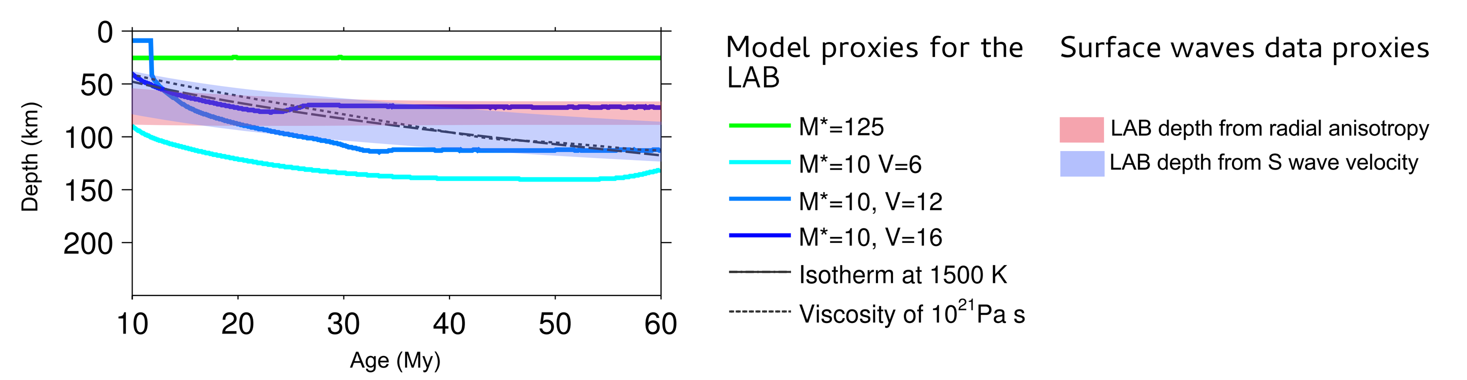


Transition of deformation mechanism for V = 16 cm^3/mol

Results

In D-Rex, the rate of evolution of the LPO is controlled by the (dimensionless) grain boundary mobility M*. Recent laboratory experiments show that a value of M* = 10 produces a much better fit for complex deformation histories (Boneh et al., 2015). The depth dependence of ξ is controlled by the activation volume of dislocation creep V.





Conclusion

Geodynamic modeling of LPO seismic observations shows how seismic anisotropy is developed, and subsequently preserved, through an upper mantle with a mixed rheology, and naturally producing the horizontal layering of radial anisotropy. The rate of LPO evolution can be constrained by the strength of anisotropy. Differences in rate of LPO evolution and deformation mechanism in the strain history controls the depth of the largest radial anisotropy gradient, which is horizontal and frozen in the lithosphere. Explaining the constant 70 km depth discontinuity of radial anisotropy does not need an additional mechanism, such as water or partial melt.

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