Lithospheric Rheology and Stress, Dynamics of Plate Tectonics, and Long-wavelength Mantle Convection

Shijie Zhong and Xi Liu

Dept. of Physics, University of Colorado
Boulder, Colorado, USA

A. B. Watts,
Dept. of Earth Sciences, Oxford University, Oxford, UK

2014 Mantle and Lithospheric Dynamics Workshop, Banff, Canada

Outline

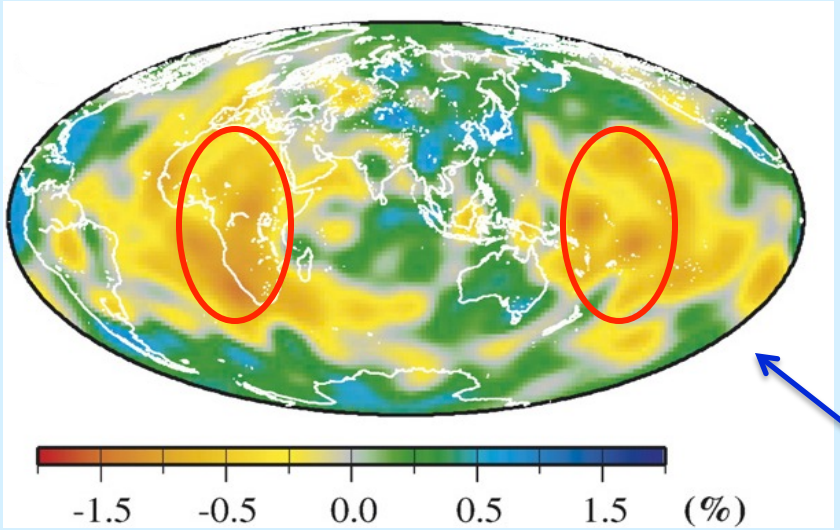
1) ***Dynamic generation of (very) long-wavelength mantle structure – controls of lithospheric viscosity.***
2) ***African and Pacific chemical piles? What would the geoid have to say about them?***
3) ***Dynamic generation of plate tectonics – large or small yield stress? or what's lithospheric stress anyway?***

The Degree-2 Mantle Structure

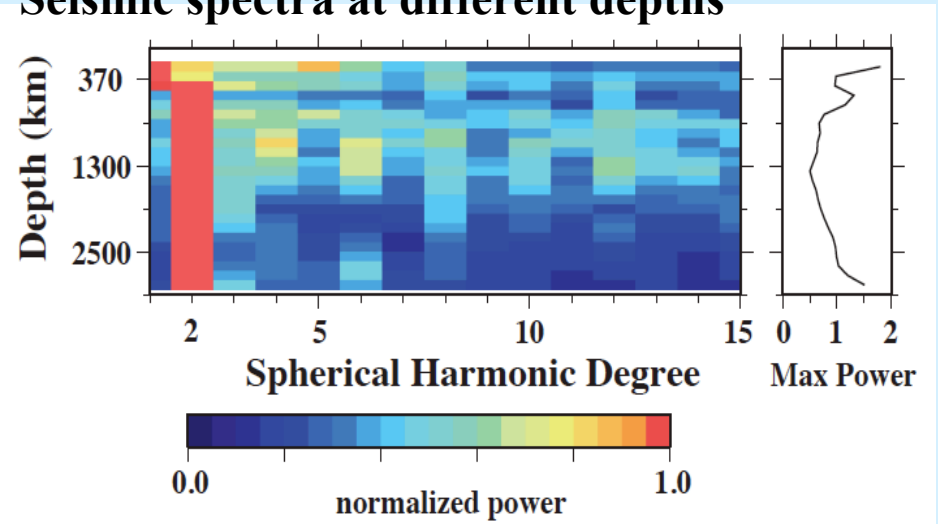
-- African and Pacific Superplumes and circum-Pacific subduction

[Dziewonski et al., 1984]

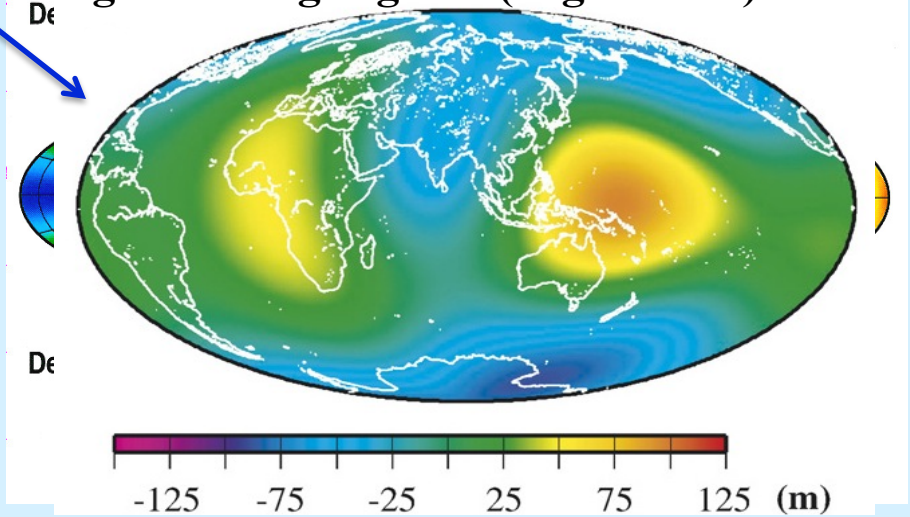
Shear-wave anomalies at 2300 km depth from S20RTS [Ritsema et al., 1999]



Seismic spectra at different depths



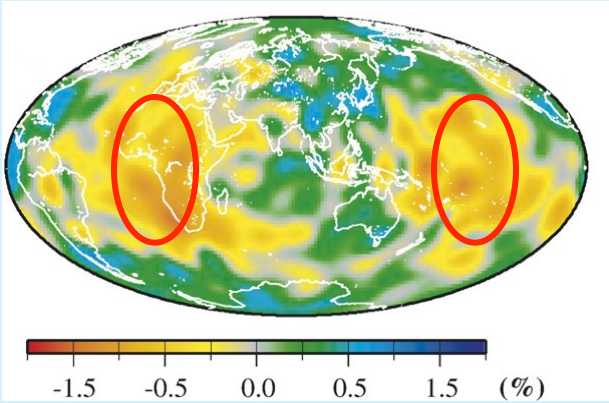
Long-wavelength geoid (degrees 2-3)



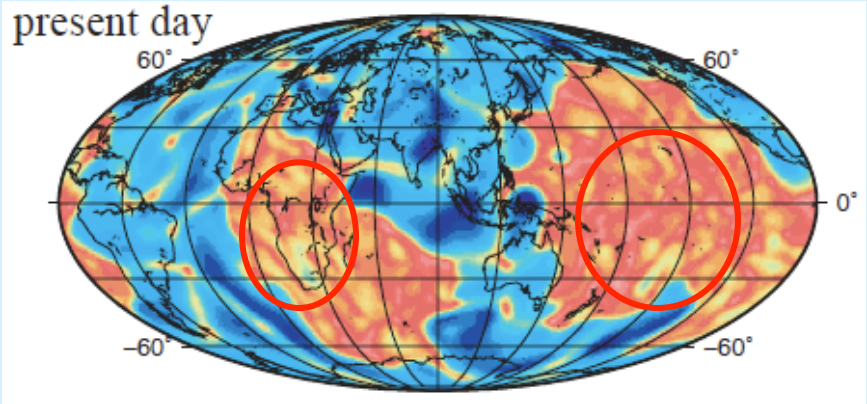
<u>Degree-2 structure:</u>
Dziewonski et al. [1984], van der Hilst et al. [1997], Masters et al. [1996, 2000], Romanowicz and Gung [2002], and Grand [2002].

Hager et al. [1985]

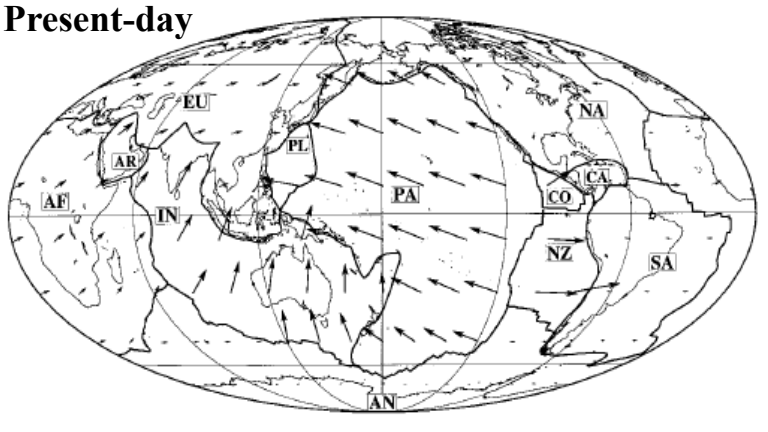
A Convective Origin for the Degree-2 Structure

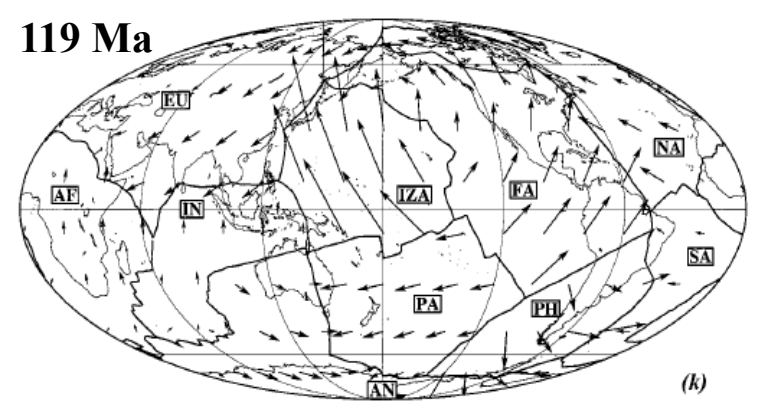


<u>Origin:</u> Controlled by plate motion and its history [Hager & O'Connell, 1981; Bunge et al., 1998].



[McNamara & Zhong, 2005; Zhang et al., 2010]

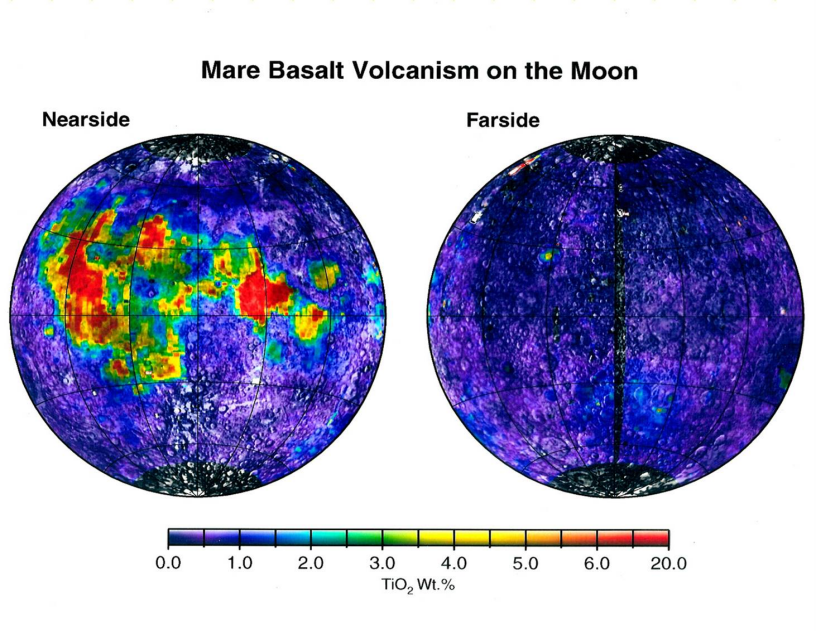




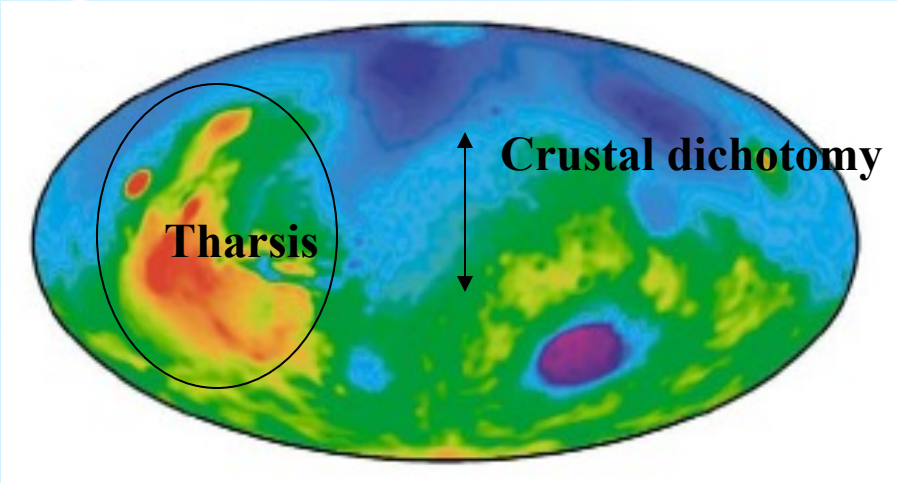
Engebretson et al. [1992]; Lithgow-Bertelloni & Richards [1998].

However, one may ask what causes the plate motion? The answer has to be mantle convection -- one of the chicken-or-egg questions. It has always been an interesting question as how to generate long-wavelength convection dynamically self-consistently.

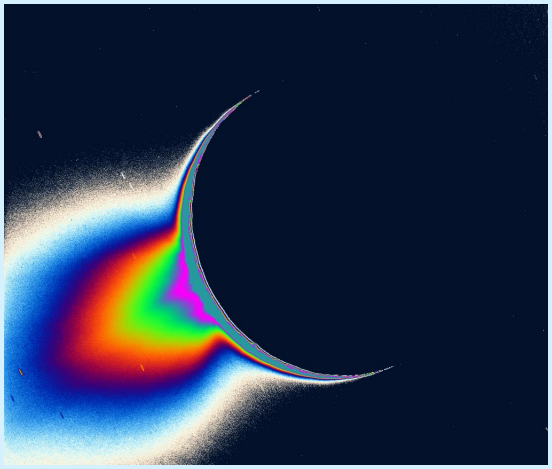
Longer wavelength than degree-2? Degree-1 or hemispherically asymmetric structures for the other planetary bodies?



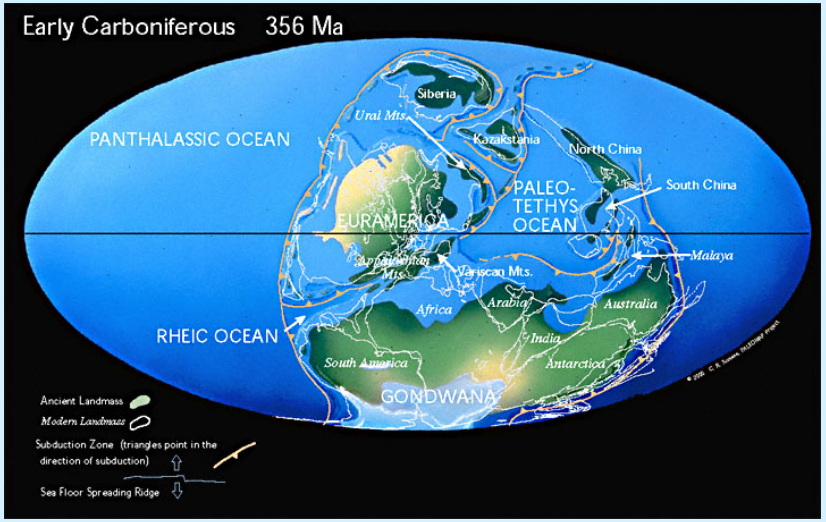
Surface topography on Mars

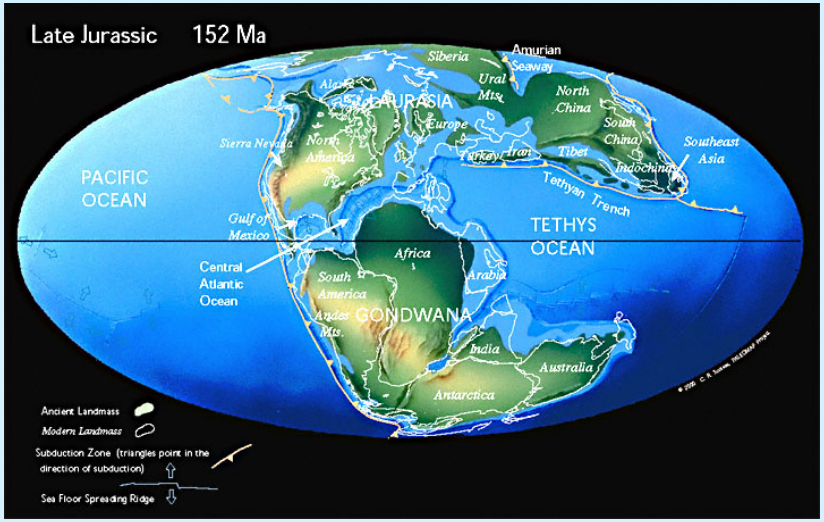


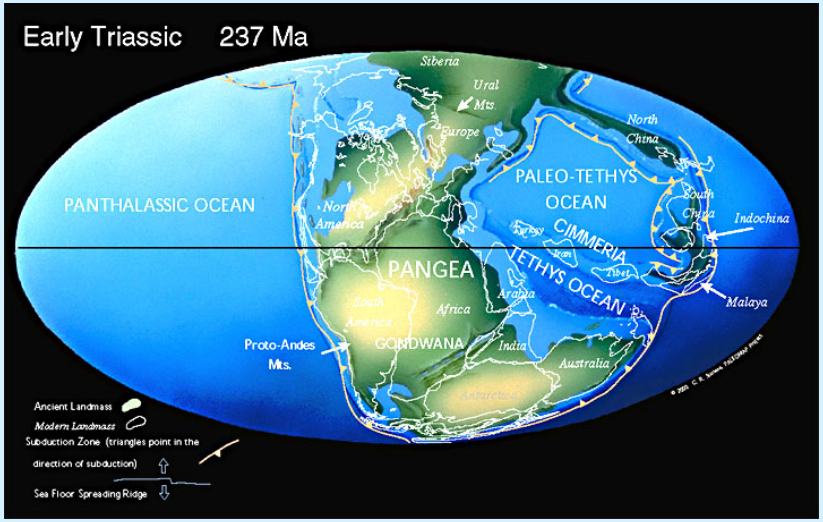
Icy satellite Enceladus



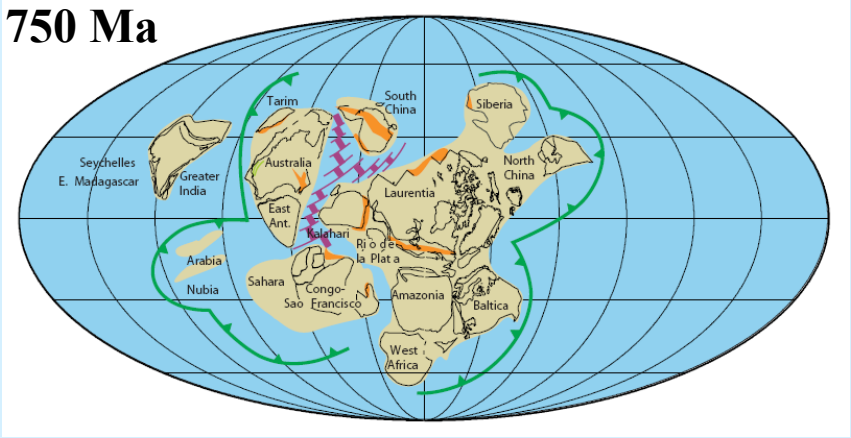
Supercontinent Pangea (330 -- 180 Ma) and Supercontinent Rodinia (900 -- 750 Ma)







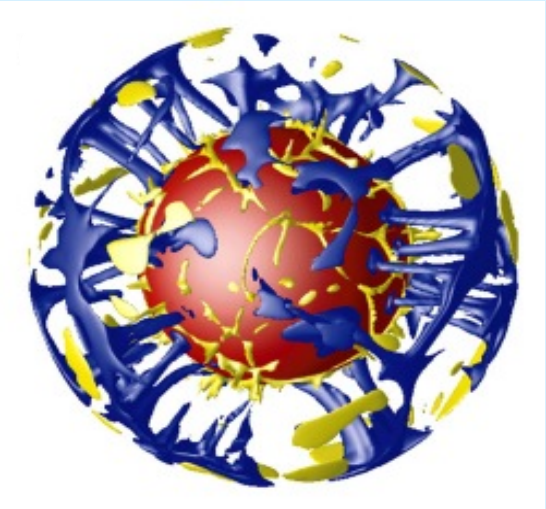
[Smith et al., 1982, and Scotese, 1997]



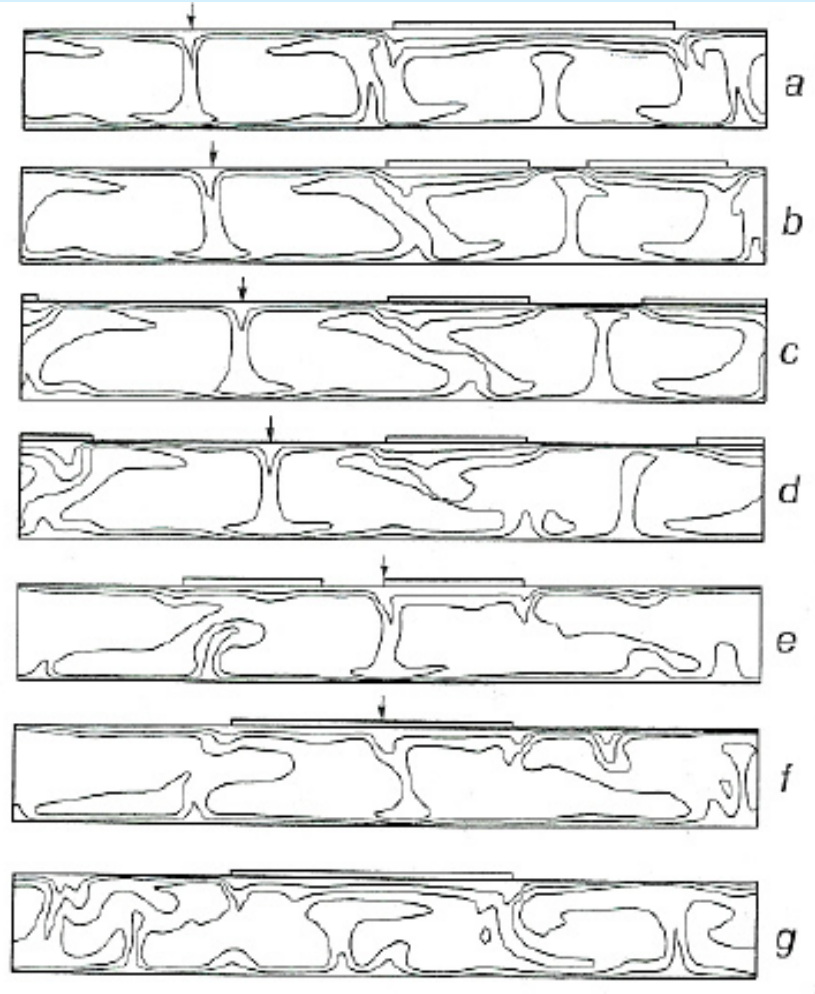
[Li et al., 2008; Hoffman, 1991; Dalziel, 1991; Torsvik, 2003].

Or degree-1 convection for the Earth – supercontinent formation?

But …

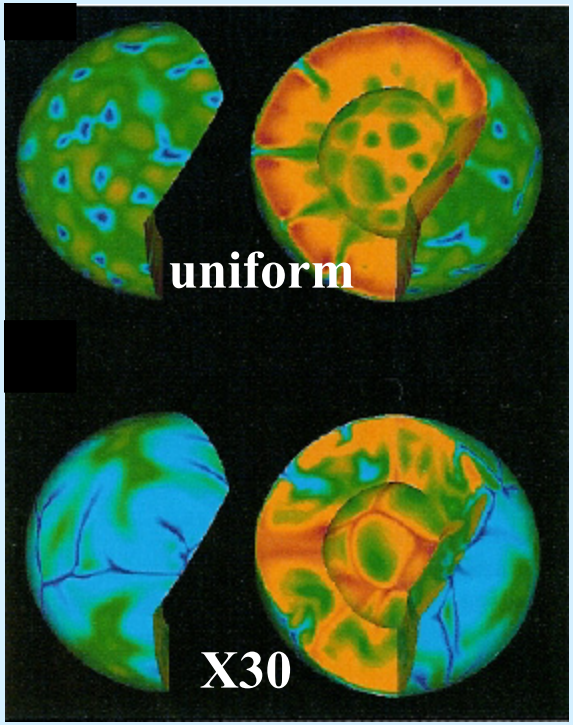


Degree-1 flow?



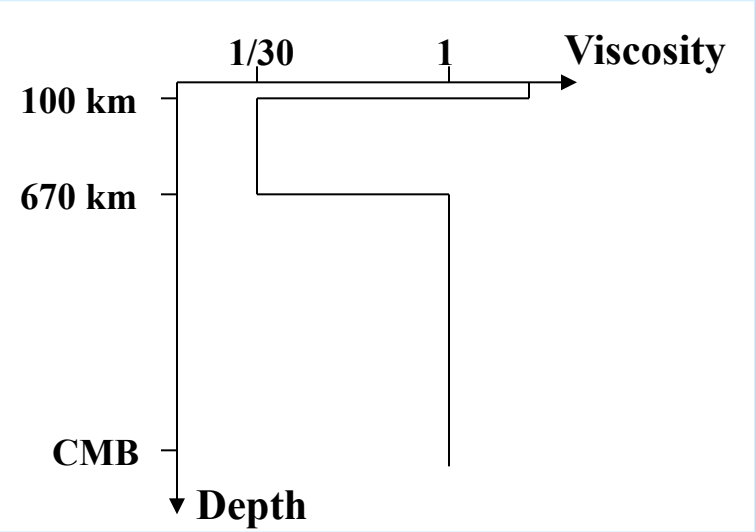
Gurnis [1988]; Gait & Lowman [2007]

Generation of long-wavelength mantle convection due to viscosity increase in the lower mantle



Bunge et al. [1996].

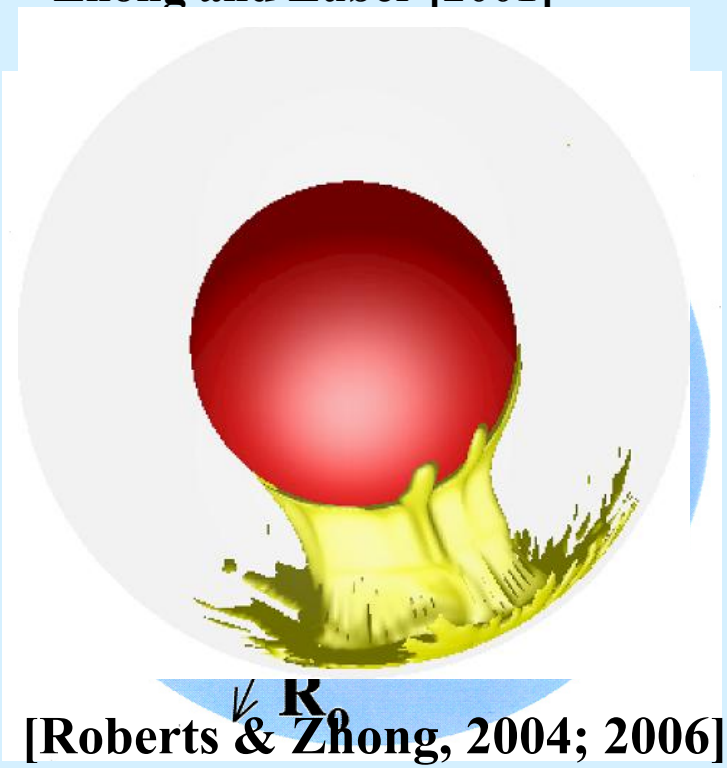
Largely at degree 6



Constrained by geoid modeling [Hager, 1991] and to some extent by postglacial rebound [Mitrovica et al., 2007].

The effect of a weak upper mantle
-- A Rayleigh-Taylor instability analysis in a sphere

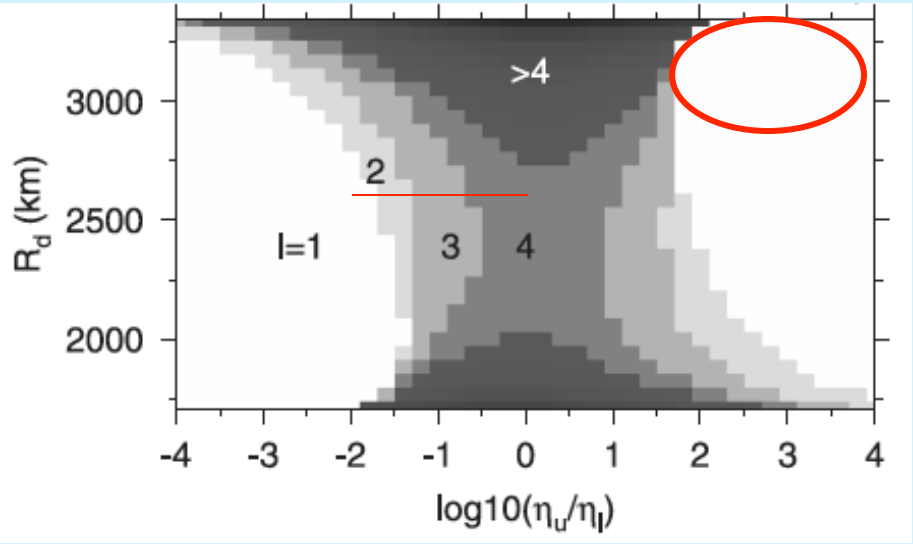
Zhong and Zuber [2001]

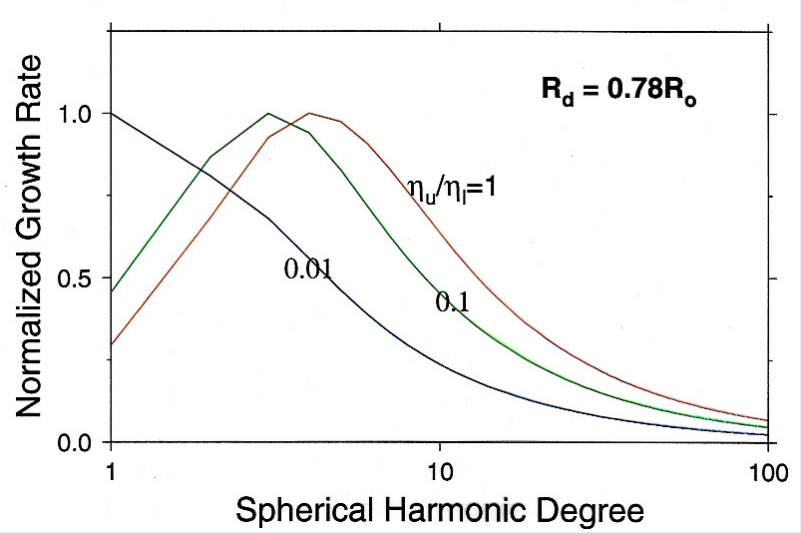


[Roberts & Zhong, 2004; 2006] for high viscosity stagnant-lid (Mars).

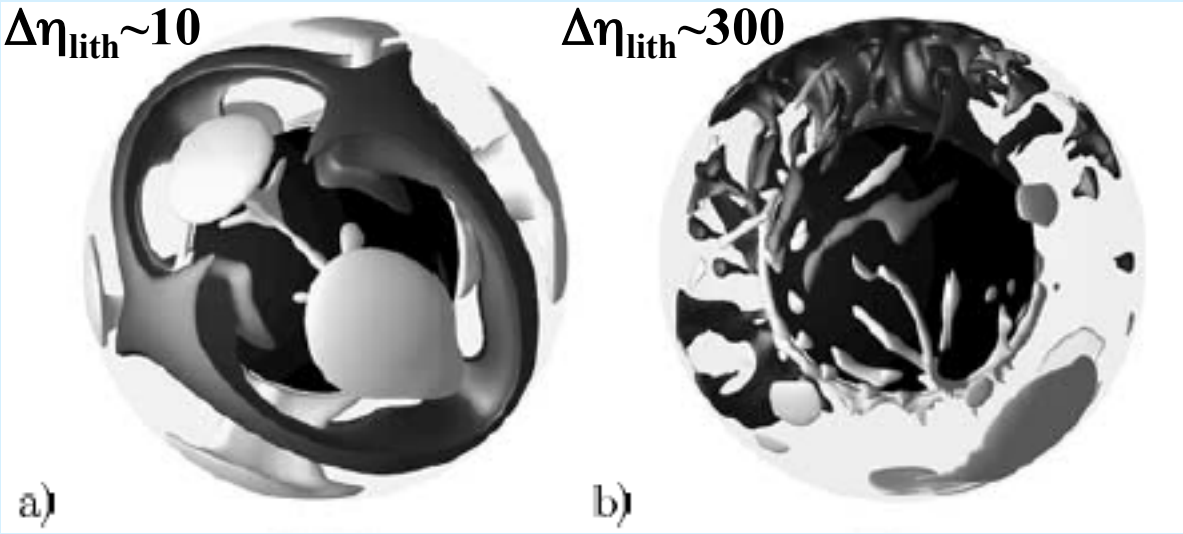
$\rho_l < \rho_u$ → gravitationally unstable.

Why only degree-6 from Bunge et al [1996] for mobile lid convection?

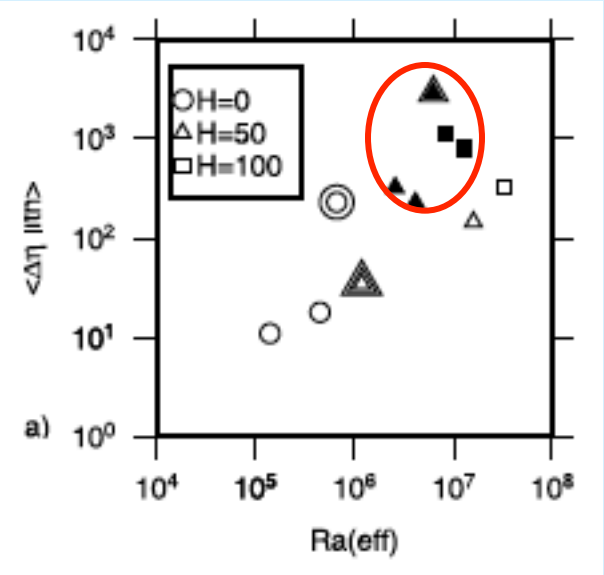




Controls of lithospheric viscosity on (very) long-wavelength convection



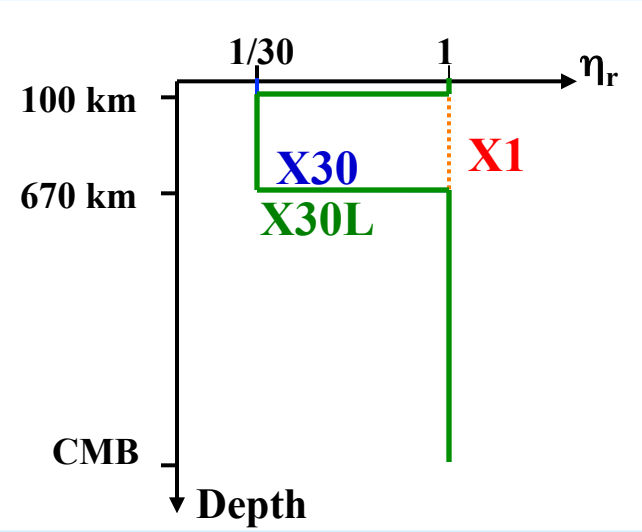
McNamara and Zhong [2005]

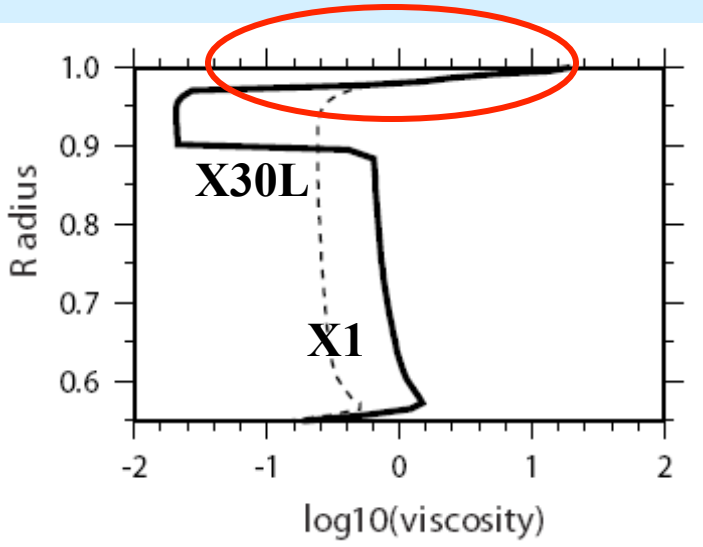


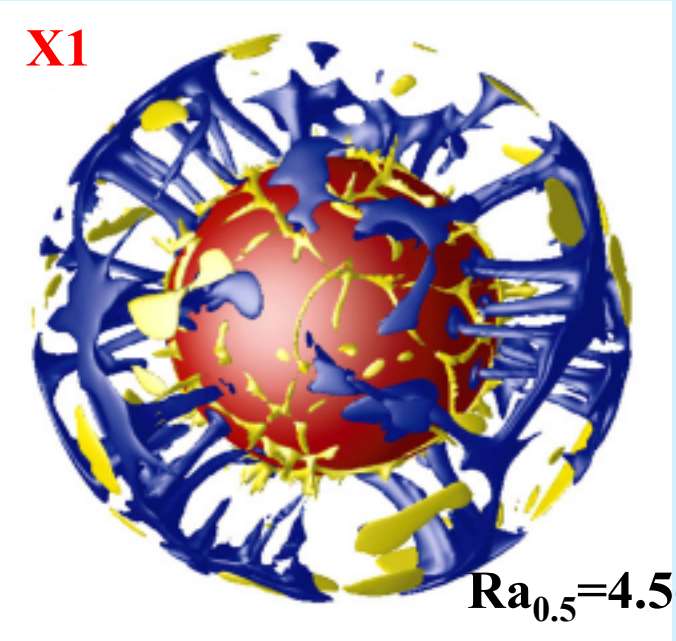
Degree-1 convection can be generated for higher Ra models, by combining moderately strong lithosphere with X30 increase in viscosity from the upper to lower mantle [Zhong et al., 2007].

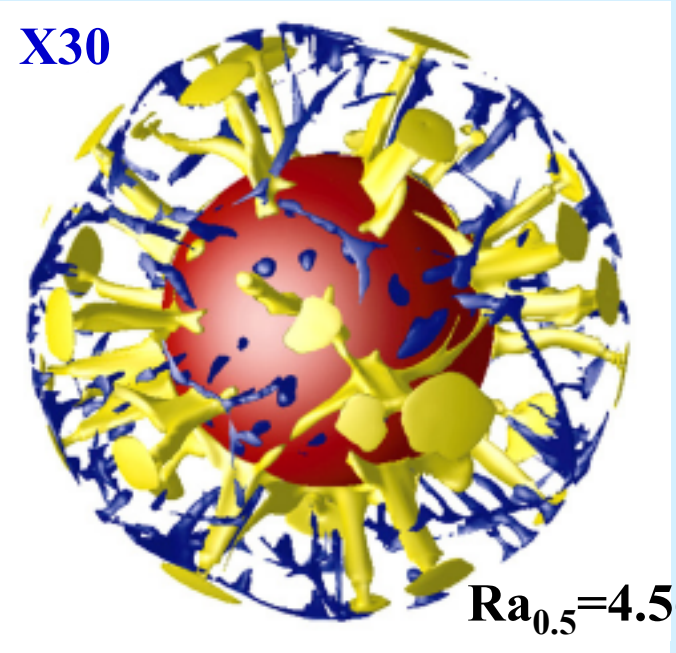
Degree-1 mobile-lid convection with realistic mantle viscosity

$\eta=\eta_r \exp[E(0.5-T)]$





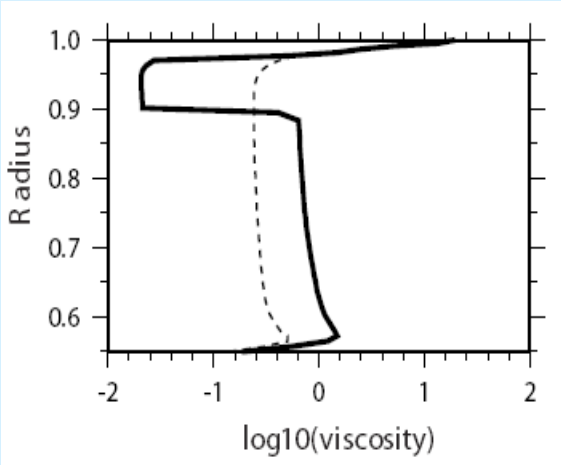


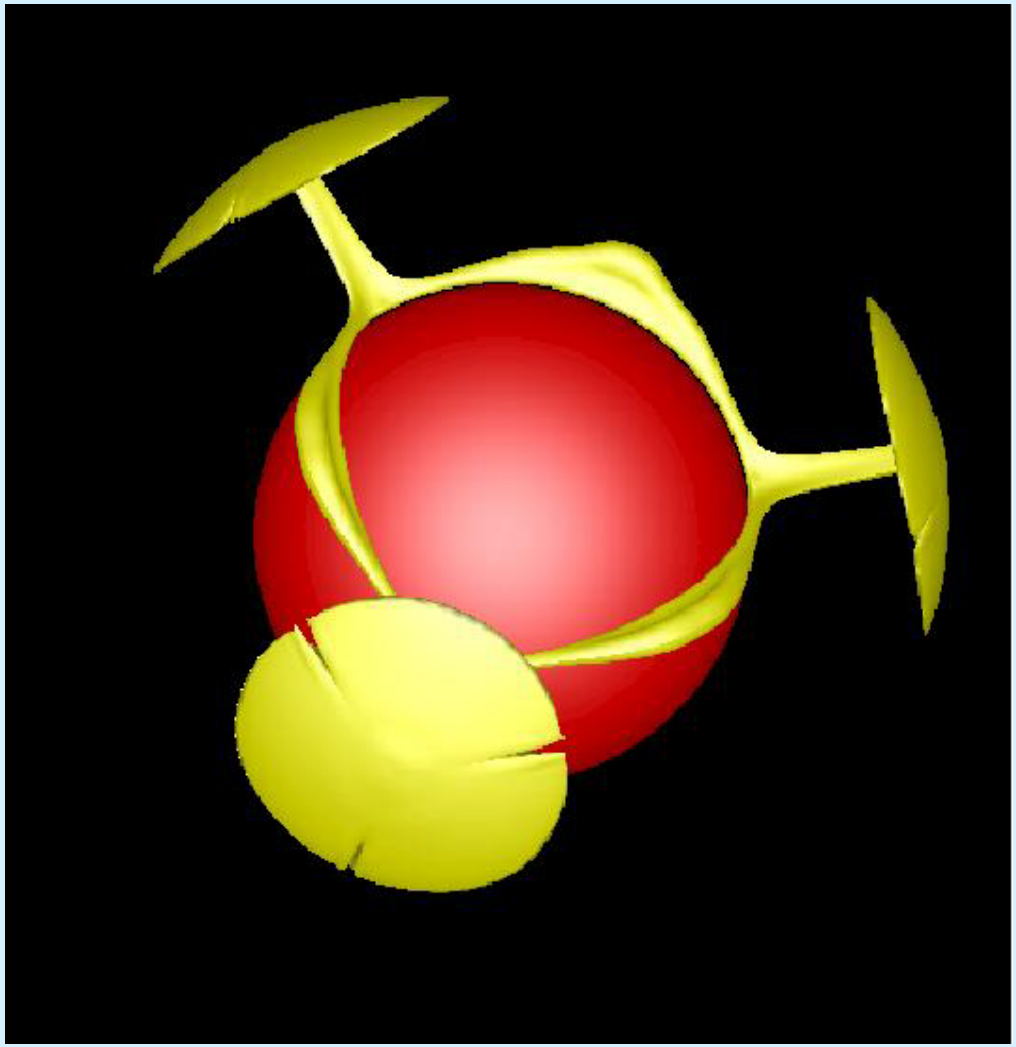


Movie 1: Evolving to degree-1 convective structure

Viscosity: η(T, depth).

$\eta_{lith} \sim 100\eta_{um}$
& $\eta_{lm} \sim 30\eta_{um}$

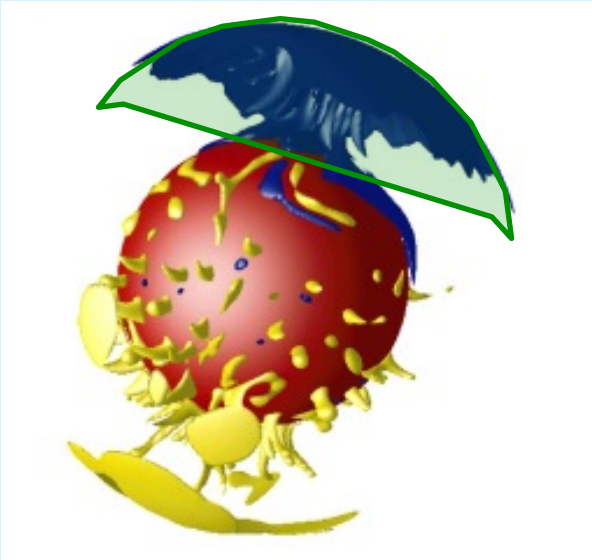


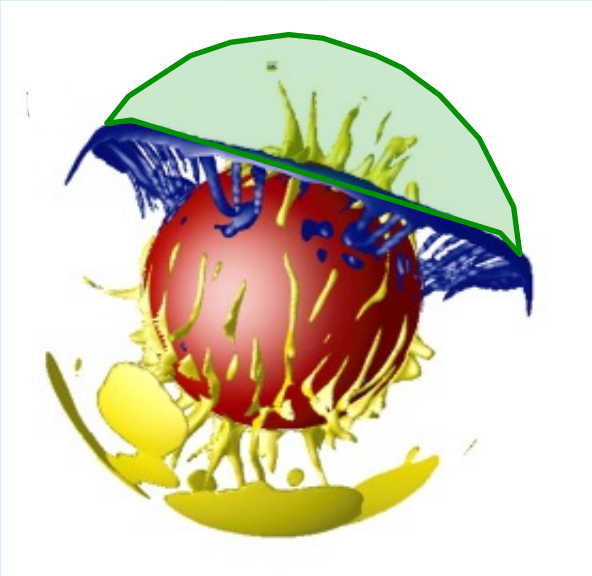


Independent of convective vigor, heating mode, & initial conditions.

Movie 2: A supercontinent turns initially degree-1 to degree-2 structures [Zhong et al., 2007]

An 1-2-1 model for the evolution of mantle structure modulated by continents [Zhong et al., 2007]





Degree-1 convection with one major upwelling system.

↓ **forming a supercontinent**

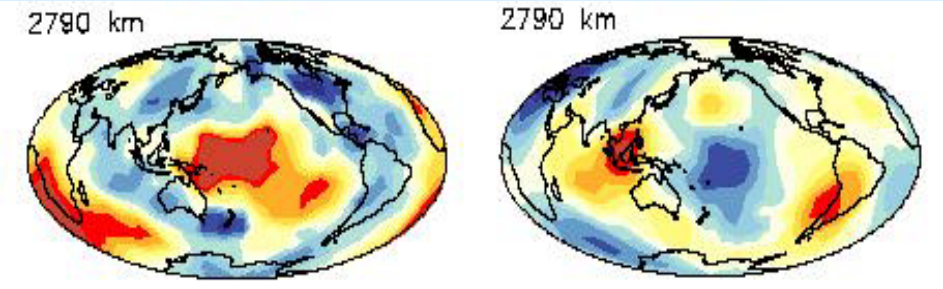
Degree-2 convection with two antipodal major upwelling systems, including one under the supercontinent.

↓ **breaking up the supercontinent**

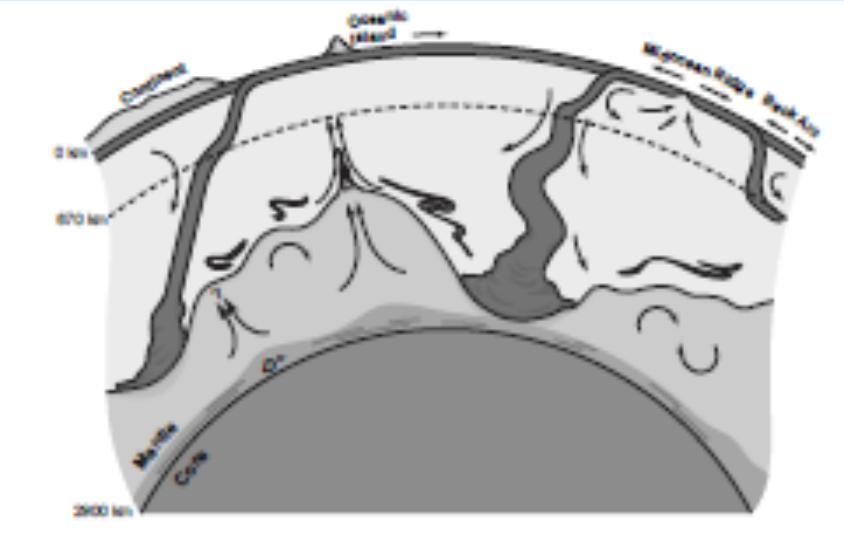
Mantle structure: 1→2→1 cycle.
At the surface: supercontinent cycle.

African and Pacific superplumes (LLSVPs) are thermochemical piles?

Vs Vc

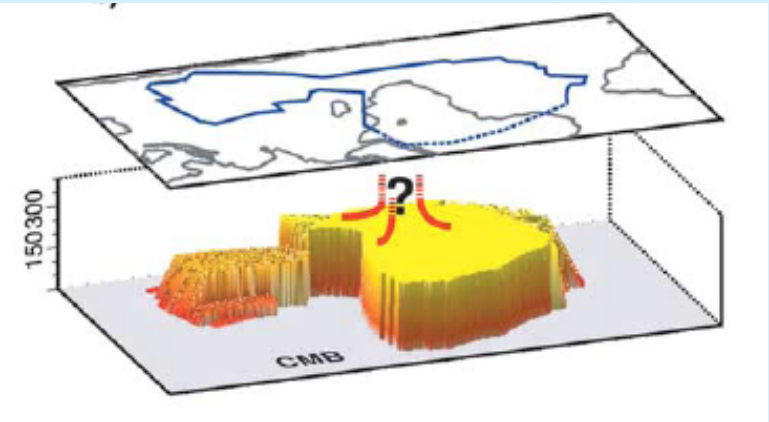


Masters et al. [2000];
also Su and Dziewonski [1997]

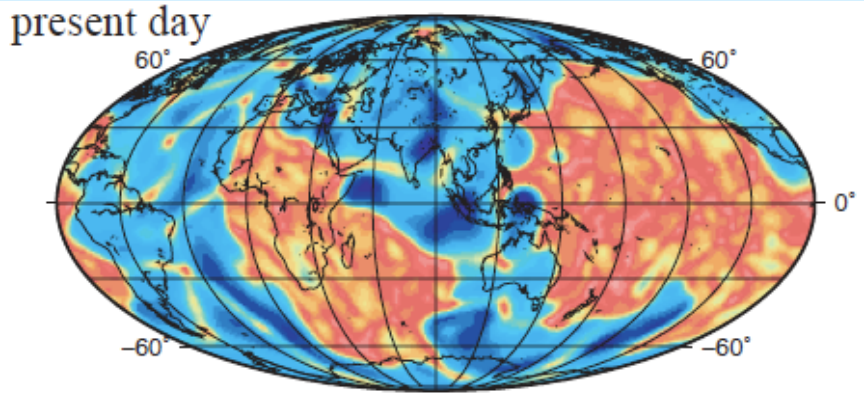


Kellogg et al. [1999]

African chemical pile extends up to 500 km above CMB



Wang & Wen [2004]



[McNamara & Zhong, 2005; Zhang et al., 2010; Deschamps et al., 2011, 2012; Nakagawa et al., 2010]

Outline

1) *Dynamic generation of (very) long-wavelength mantle structure – controls of lithospheric viscosity.*
2) *African and Pacific chemical piles? What would the geoid have to say about them?*
3) ***Dynamic generation of plate tectonics – large or small yield stress? or what's lithospheric stress anyway?***

Why do we care about lithospheric rheology and stress?

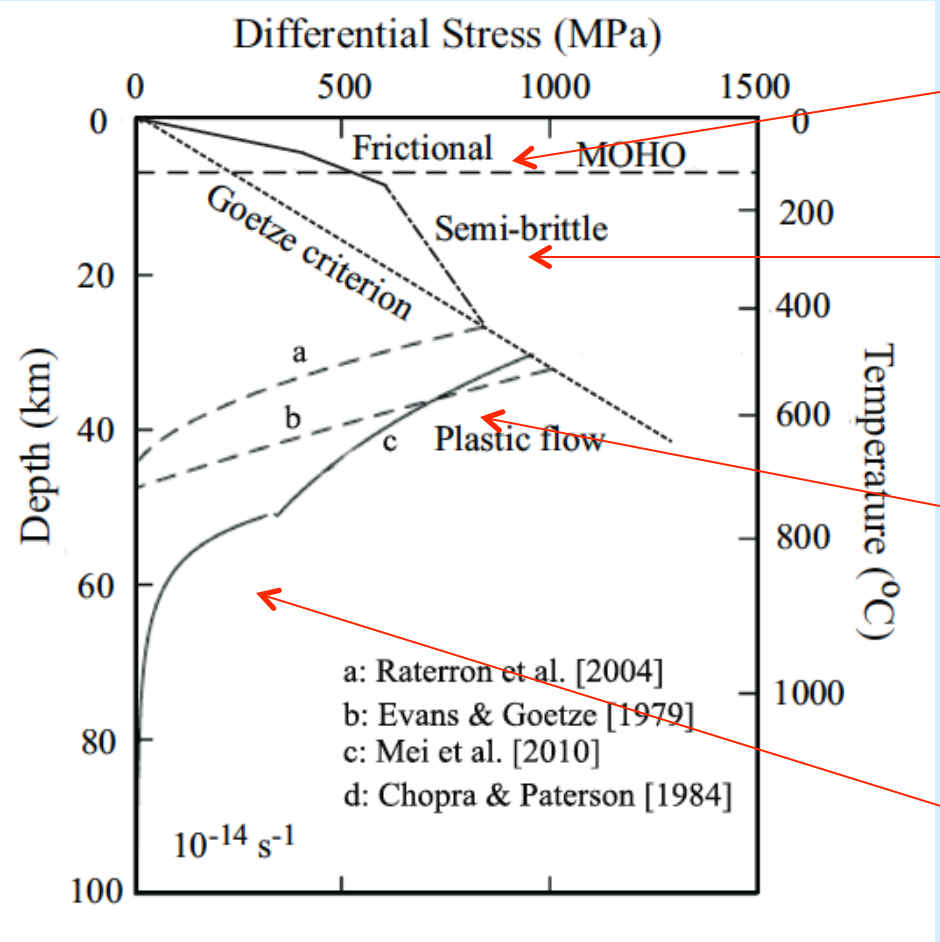
Many reasons (tectonics, earthquakes …), but here the issue is <u>the dynamic generation</u> of plate tectonics.

It has been suggested that small <u>coefficient of friction (μ<0.1)</u> or <u>yield stress (<100 MPa)</u> would lead to plate tectonics in models of mantle convection.

[Moresi & Solomatov, 1998; Trompert & Hansen, 1998; Tackley, 2000; Richards et al., 2001; O'Neill et al., 2007; Foyle & Becker, 2009; Gerya, 2010; Coltice et al., 2012]

Place observational (in-situ) constraints on <u>coefficient of friction μ, lithospheric rheology and stress</u> near Hawaii islands -- plate interiors near the largest loads [Zhong & Watts, 2013].

Three deformation regimes in lithosphere: A laboratory view [e.g., Mei et al., 2010]



1) Frictional sliding (shallow depths) or Byerlee's law:

$$\tau_{yield} = c_0 + \mu \rho g z,$$

μ~ 0.6

2) Semi-brittle (transitional regime, poorly understood)

3) Plastic flow:

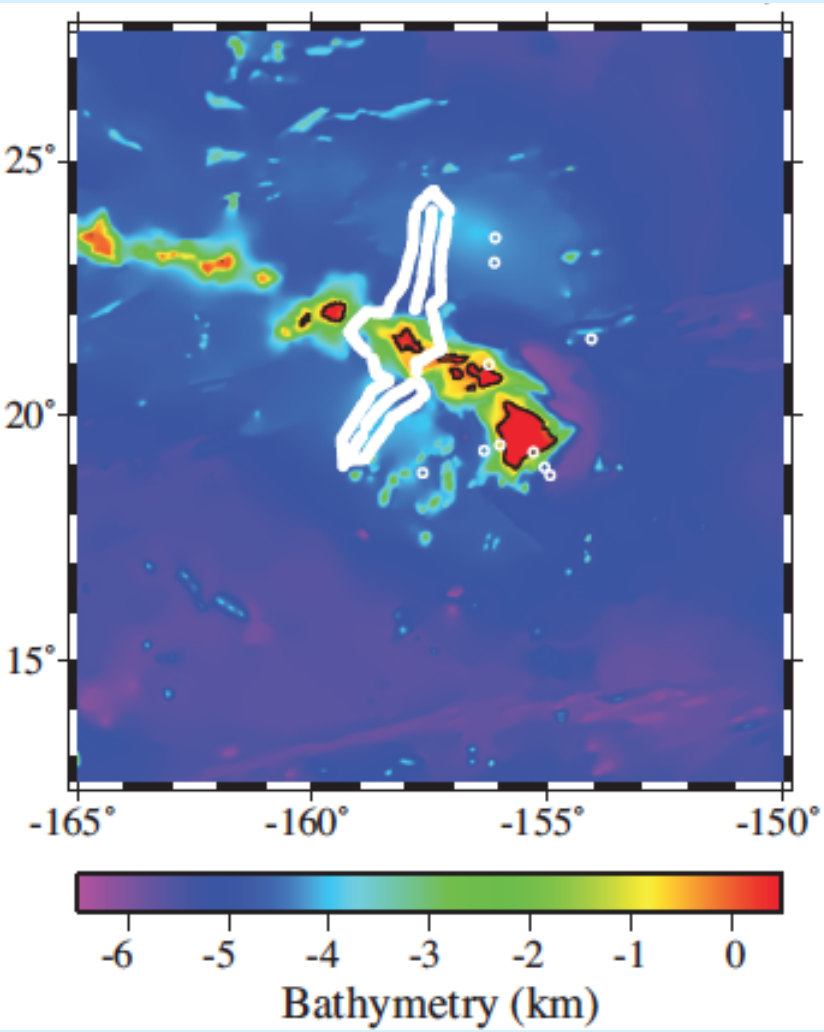
a) Low-temperature plasticity (<~800 °C):

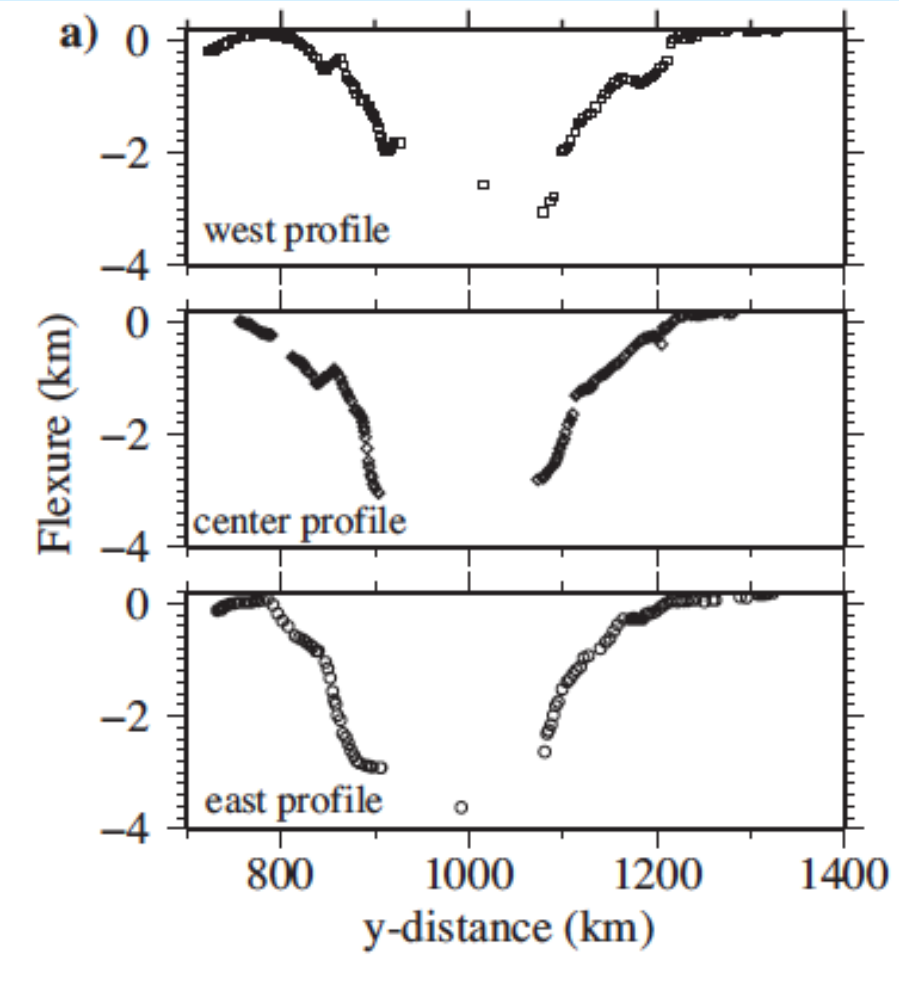
$$\dot{\varepsilon} = A\sigma^n \exp\{-\frac{E_p}{RT}[1-(\frac{\sigma}{\sigma_p})^p]^q\}$$

b) High-temperature creep (>~800 °C):

$$\dot{\varepsilon} = A\sigma^n \exp(-\frac{E_c}{RT})$$

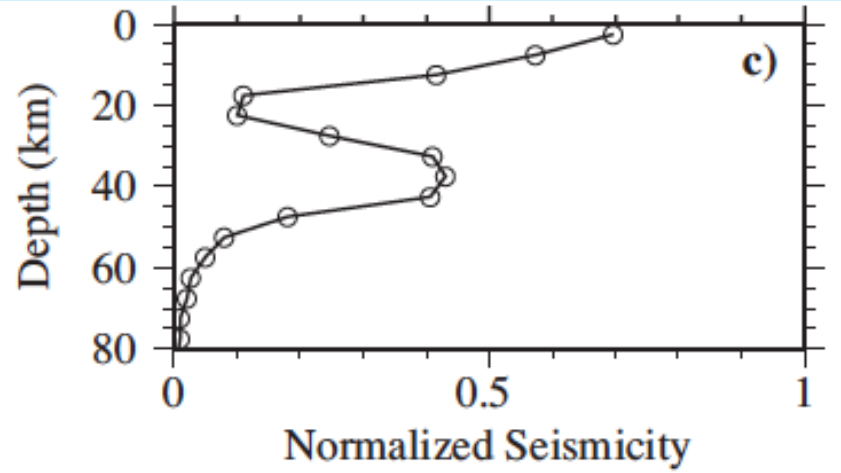
Hawaiian volcanic loads and lithospheric response – a natural laboratory

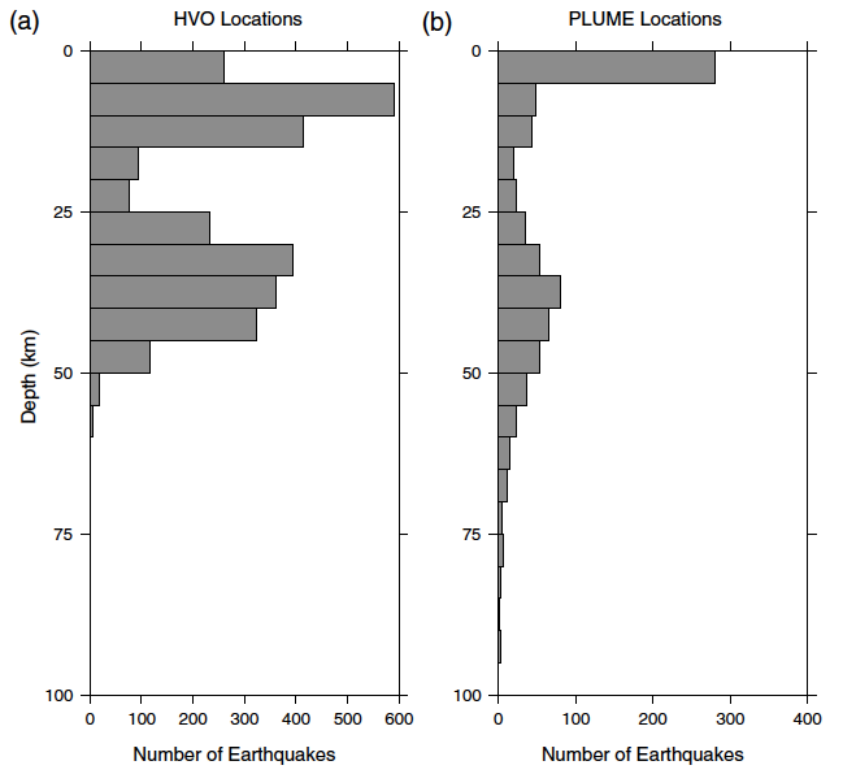


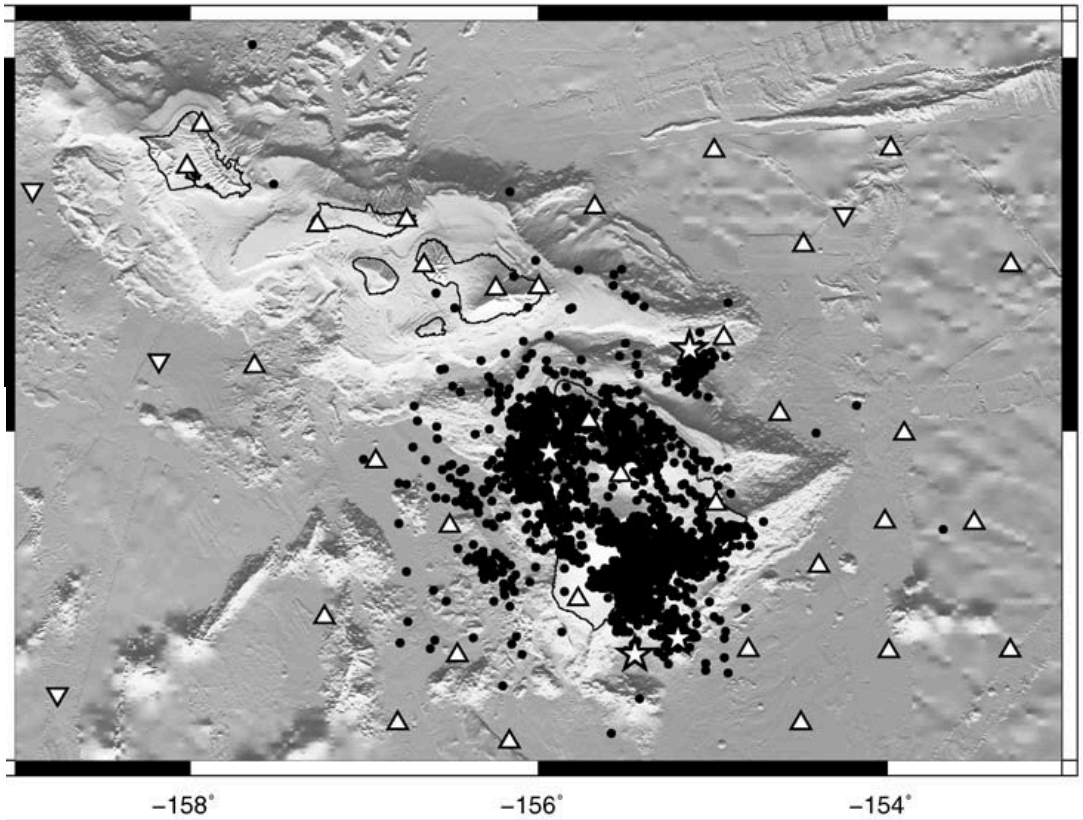


Watts et al., 1985; Zucca et al., 1982; Zucaa & Hill, 1980; Shor & Pollard, 1964.

Load-induced Seismicity in Hawaiian region

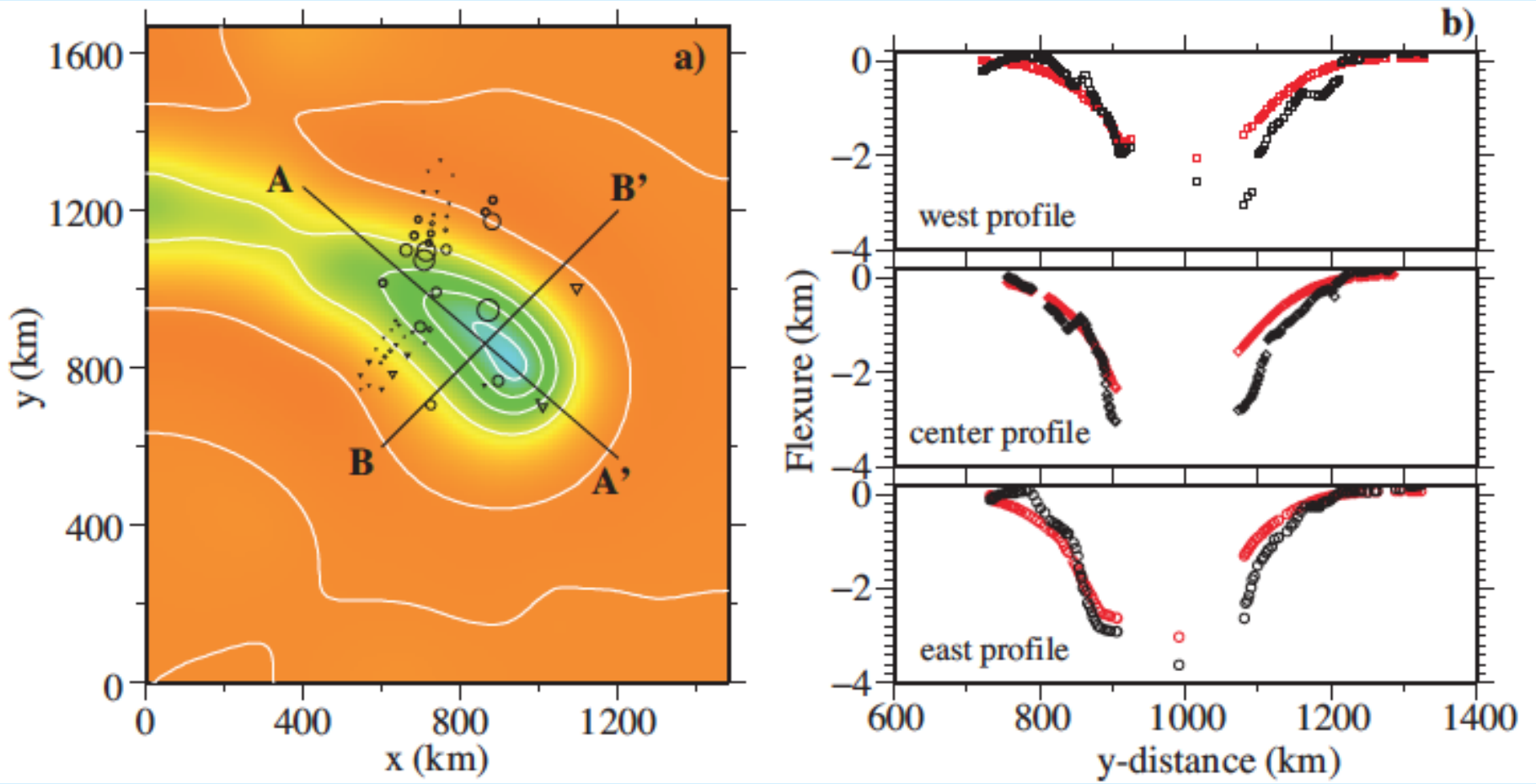






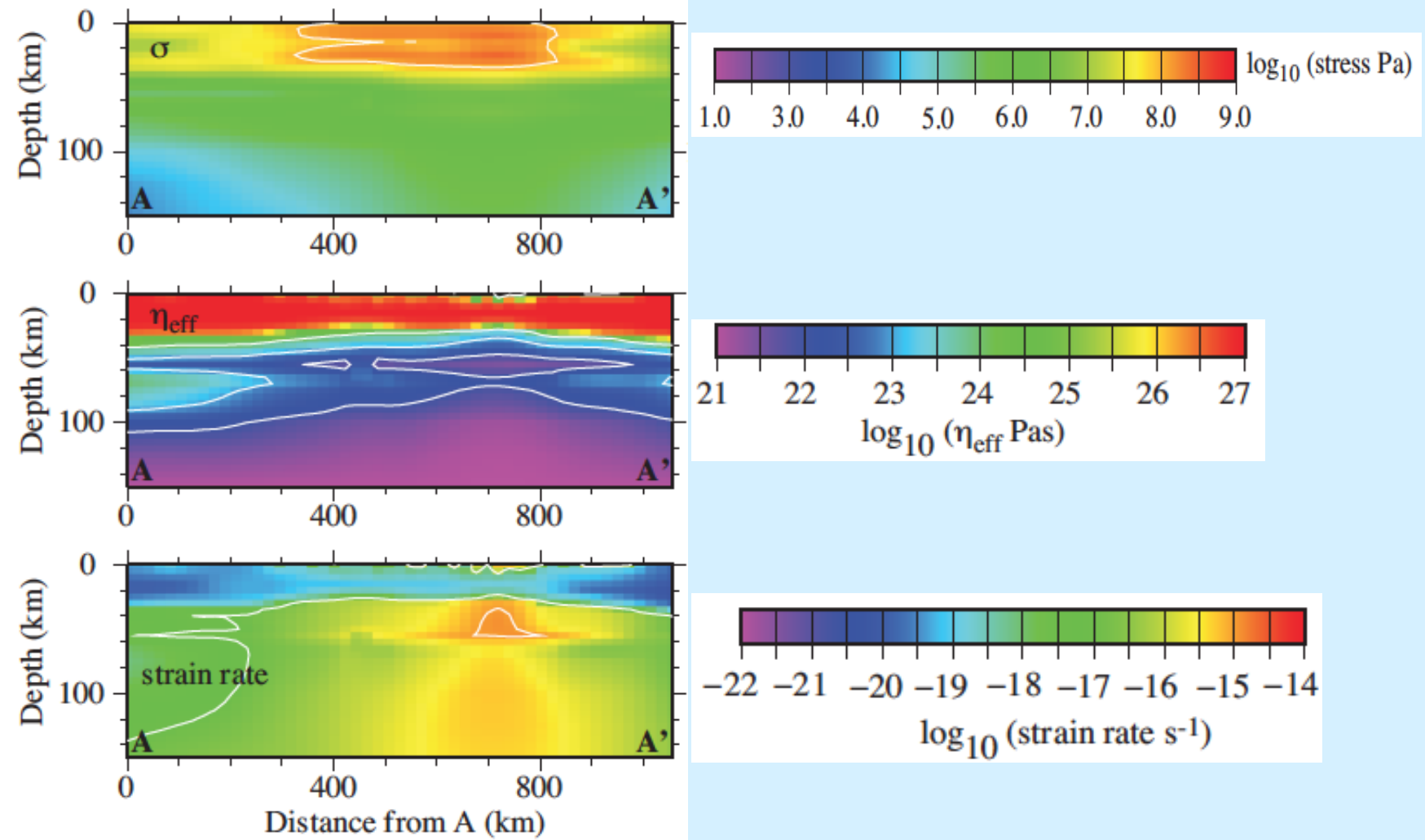
Archieta et al., 2011.
Wolfe et al., 2004.

A 3-D viscoelastic loading model with μ_f=0.7, low-T plasticity by Mei et al. [2010] (but a <u>reduced</u> pre-factor), and standard high-T creep

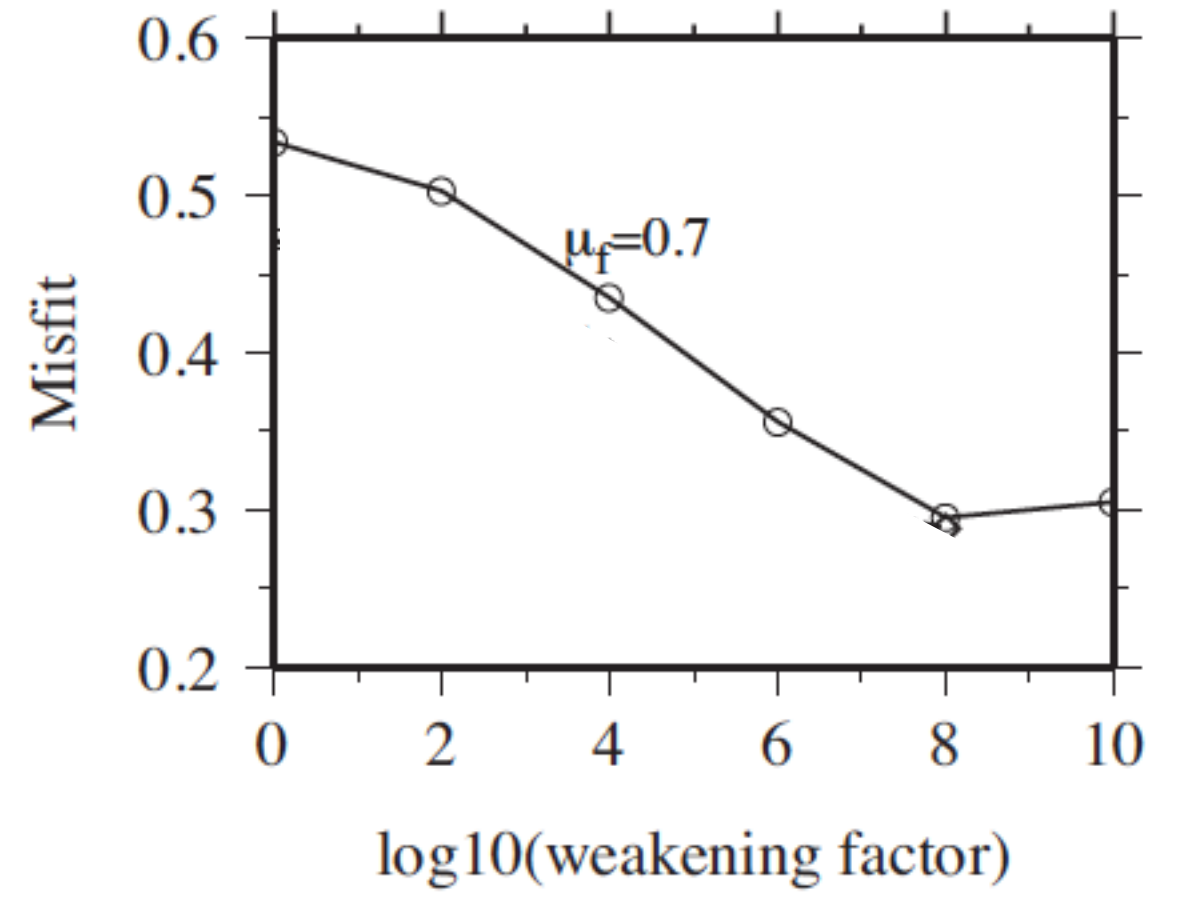


Zhong & Watts [2013]

Stress, effective viscosity and strain rate along AA' cross section



Controls of <u>low-T plasticity</u> and μ_f on flexure and a trade-off between them as seen by misfit

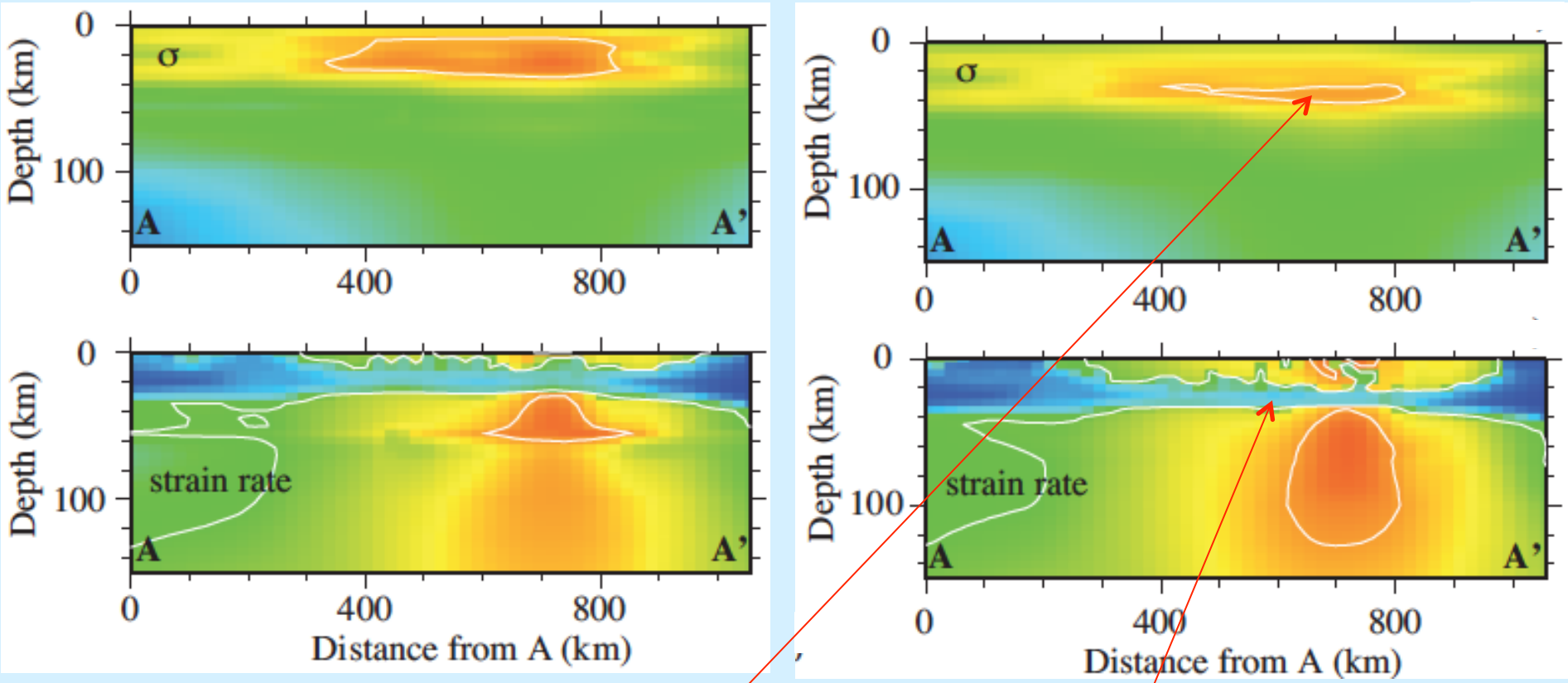


Zhong & Watts [2013]

Stress and strain rate for <u>two cases</u> with identically small misfit for flexure

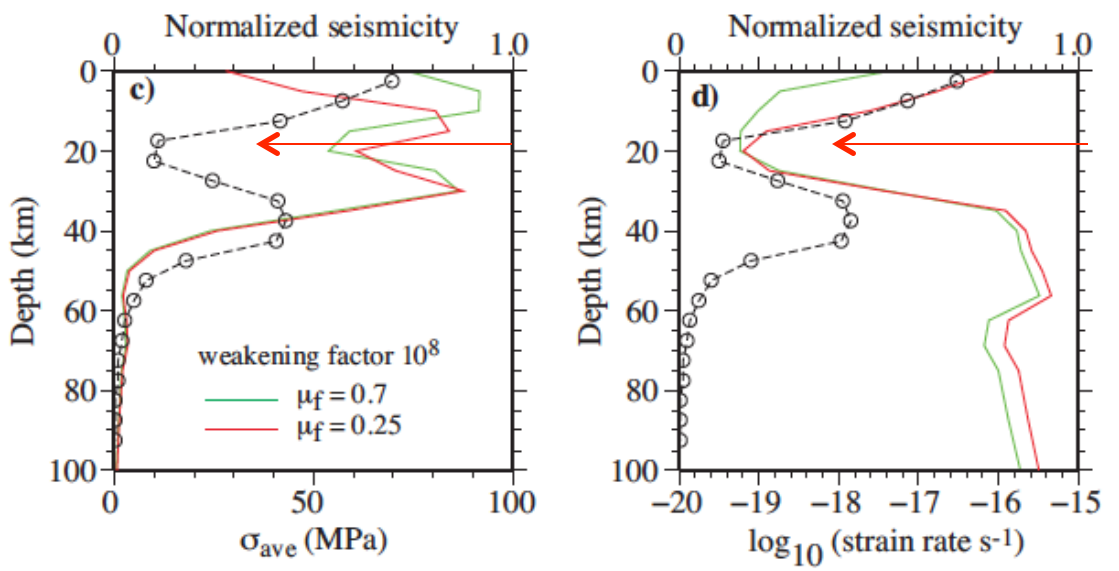
μ_f=0.25; weakening of 10^8

μ_f=0.1; weakening of 10^6



A smaller μ_f pushes high stress and low strain rate to larger depth

Seismicity removes the ambiguity and poses constraints on μ_f (>0.25)



μ_f at 0.25-0.7 appears to explain the seismicity pattern,

but not for smaller μ_f at 0.1.

Zhong & Watts [2013]

The State of Stress in the Earth's Lithosphere.

H. KANAMORI

Seismological Laboratory, California Institute of Technology - Pasadena, Cal. 91125

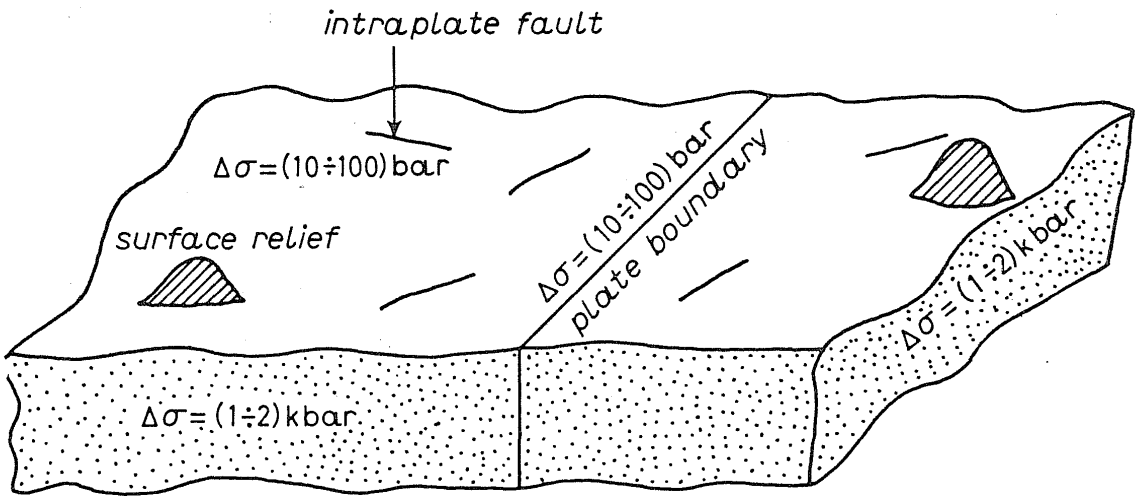


Fig. 11. – The low-stress model. The stress differences of at least 1 to 2 kbar due to surface reliefs are supported by the lithosphere, while the stress on intraplate weak zones and plate boundaries is maintained at a low level (10 to 100 bar).

Reprinted From
Physics of the Earth's Interior

Soc. Italiana di Fisica - Bologna - Italy

A provocative statement on generation of plate tectonics from mantle convection

- ***Assuming a weak lithosphere (a small μ or yield stress) as done in most convection models does not address how or why plate tectonics is generated.***
- ***The key is to understand why <u>lithospheric strength</u> evolves from strong plate interiors to weak plate margins.***

LETTER

doi:10.1038/nature13072

Plate tectonics, damage and inheritance

David Bercovici[1] & Yanick Ricard[2]

Conclusion

- ***Very long-wavelength convection is readily generated by moderately high lithospheric viscosity and weak upper mantle.***
- ***1-2-1 model for mantle structure evolution as modulated by supercontinent process.***
- ***Maximum lithospheric stress under Hawaii is ~ 100-200 MPa – probably the largest on the Earth. Coefficient of friction is in the range of 0.25-0.7, as constrained by seismicity and observed flexure.***