Mantle Convection and Plate Tectonics: state of the art and open questions

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Plate tectonics / Earth unusual ?

- Mars: rigid lid
  - Had plate tectonics early?
- Venus: rigid lid
  - Plate tectonics->rigid lid?
  - Episodic overturn?

Early Earth had different type of plate tectonics?

- Reasons:
  - Oceanic crust too thick=> slab buoyant
  - Inherent scaling of plate-mantle dynamics
- Some possibilities:
  - Sub-crustal subduction
  - Distributed plate boundaries
  - No plate tectonics (rigid lid)

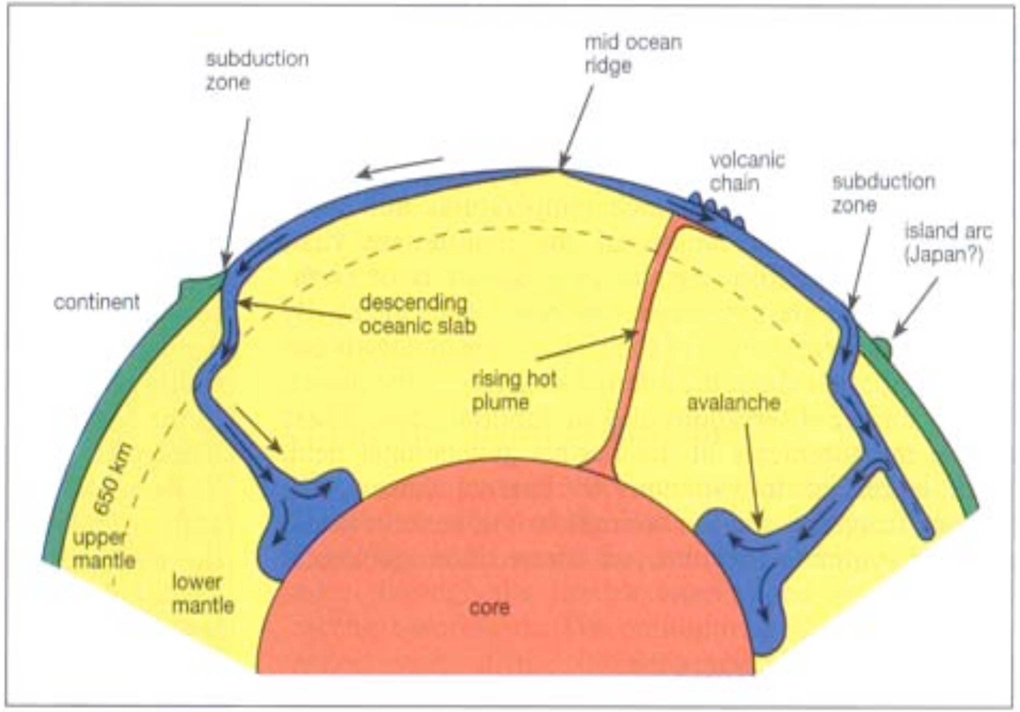
We don’t understand plate tectonics at a fundamental level

- Rock deformation is complex
  - Viscous, brittle, plastic, elastic, nonlinear
  - Dependent on grain size, composition (major and trace element, eg water)
- Multi-scale
  - Lengthscales from mm to 1000s km
  - Timescales from seconds - Gyr

Dynamical lengthscales

Global

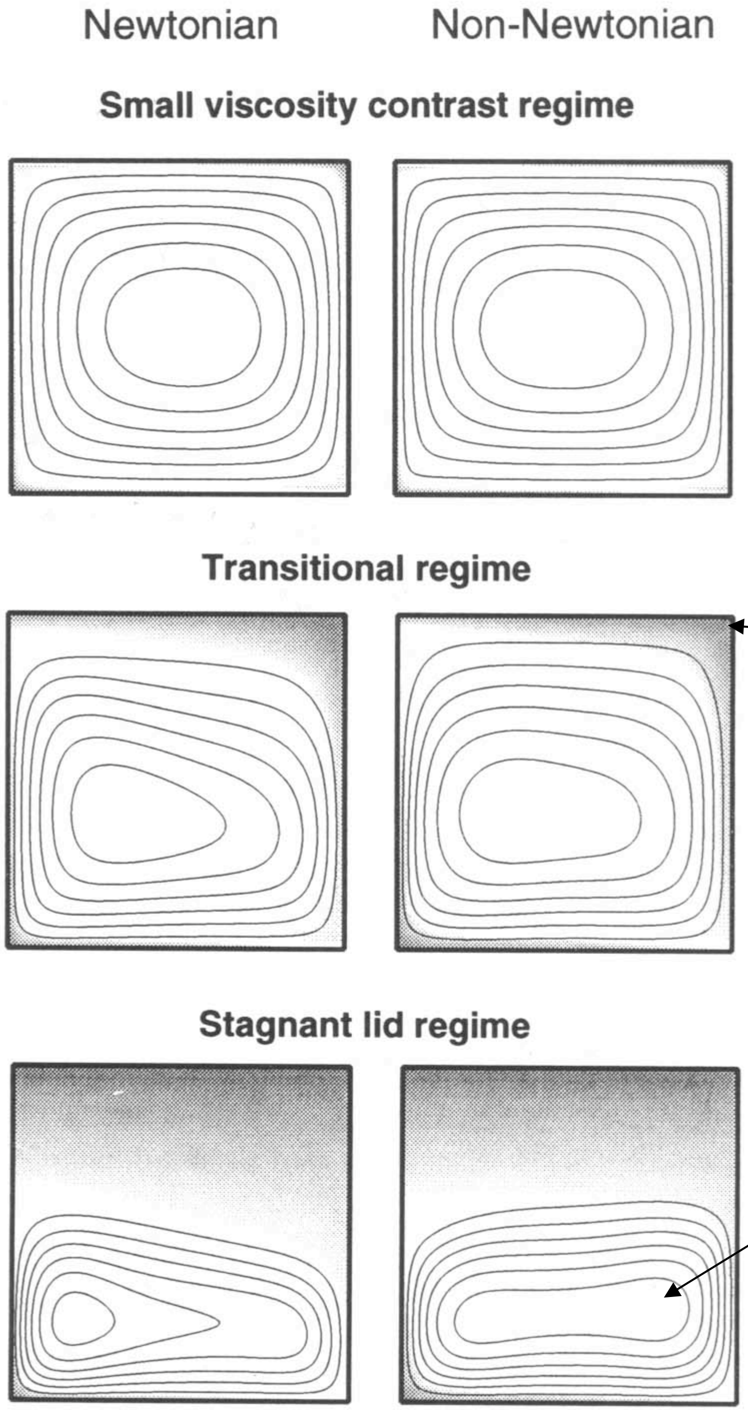
'Human' scale



1 Schematic diagram showing the processes that occur in the mantle. The lithosphere – the outermost layer of the Earth – is made up of tectonic plates that move relative to one another. Where two plates converge, the heavy oceanic plates (blue) sink into the mantle in a process known as subduction, which cools the mantle below. Continental plates (green), which are lighter, do not subduct – at the boundaries between these plates earthquakes and volcanoes occur, and mountain ranges are formed. Hot material rises from the base of the mantle in the form of "plumes", causing volcanoes to form.

The plate problem

- Viscous, T-dependent rheology appropriate for the mantle leads to a stagnant lid
- exp(E/kT) where E~340 kJ/mol
- T from 1600 -> 300 K
- =>**1.3×10^{48}** variation
- => RIGID or STAGNANT LID!

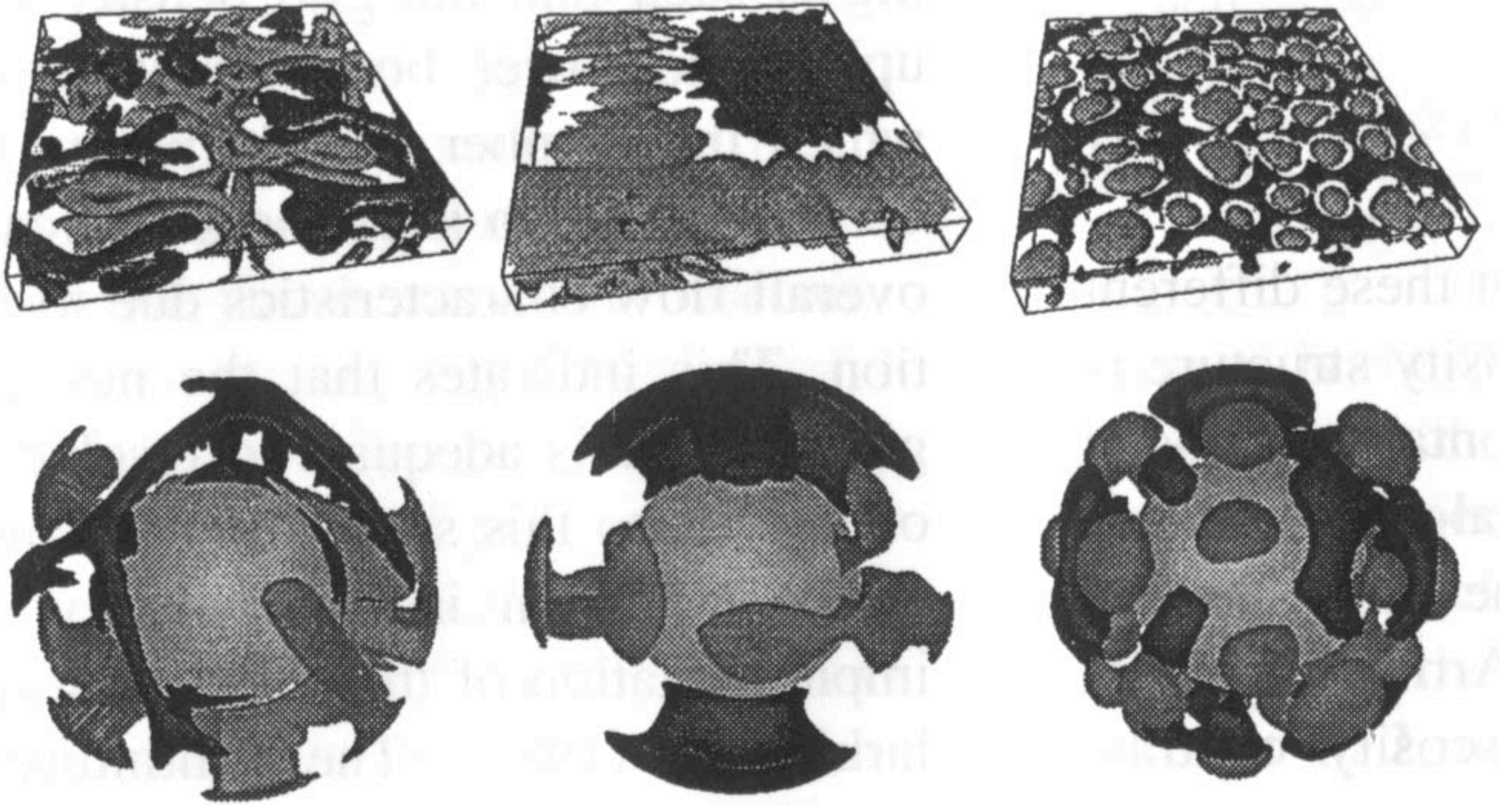


(figure from Solomatov + Moresi)

Most dissipation is in lid: this determines velocities

~constant viscosity convection below stagnant lid

The 3 regimes in 3D



Modelling Plates and Mantle

- ‘Traditional’ approach
    - 2 separate systems, insert by hand
    - plates ‘drive’ mantle (geologists/tectonicists)
    - mantle ‘drives’ plates (geodynamicists)
- Self-consistent approach
    - One system
    - same rheology applies everywhere: viscosity(T,p,e,C,history)

Rheology

- Typical mantle convection models:
  - temperature-dependent
  - Diffusion creep and dislocation creep
- Realistic:
  - as above plus:
  - elastic and brittle
  - plasticity/Peierls
  - dependent on grain-size, composition, volatile content...
  - history-dependent (e.g., strain weakening or hardening)
- Complicated: what is most important? What is the appropriate ‘large-scale’ rheology?

Strength of rocks

- Increases with confining pressure (depth) then saturates

Low-T deformation: Effect of P

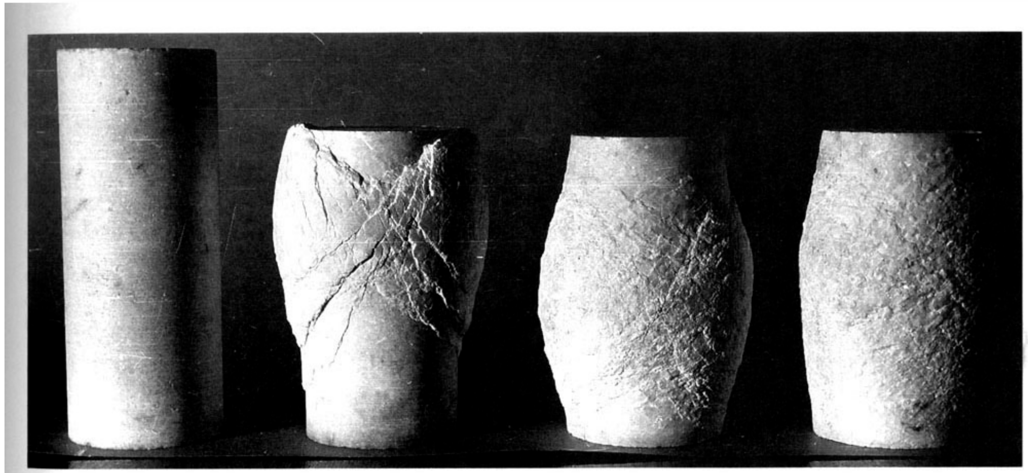


FIGURE 15.6
A marble cylinder deformed in the laboratory by applying thousands of pounds of load from above. Each sample was deformed in an environment

Low T: Effect of P

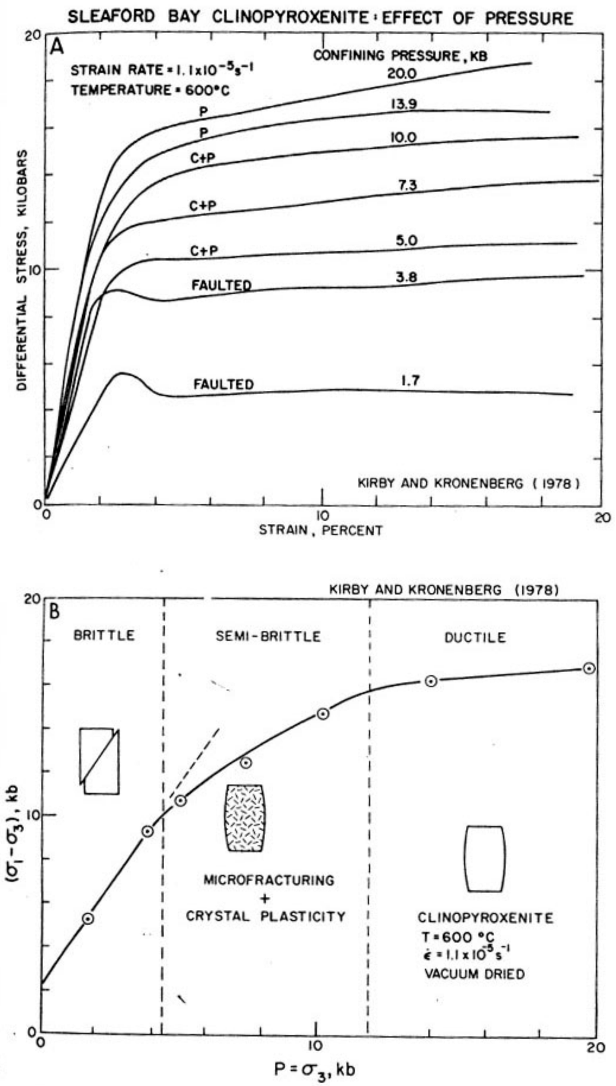
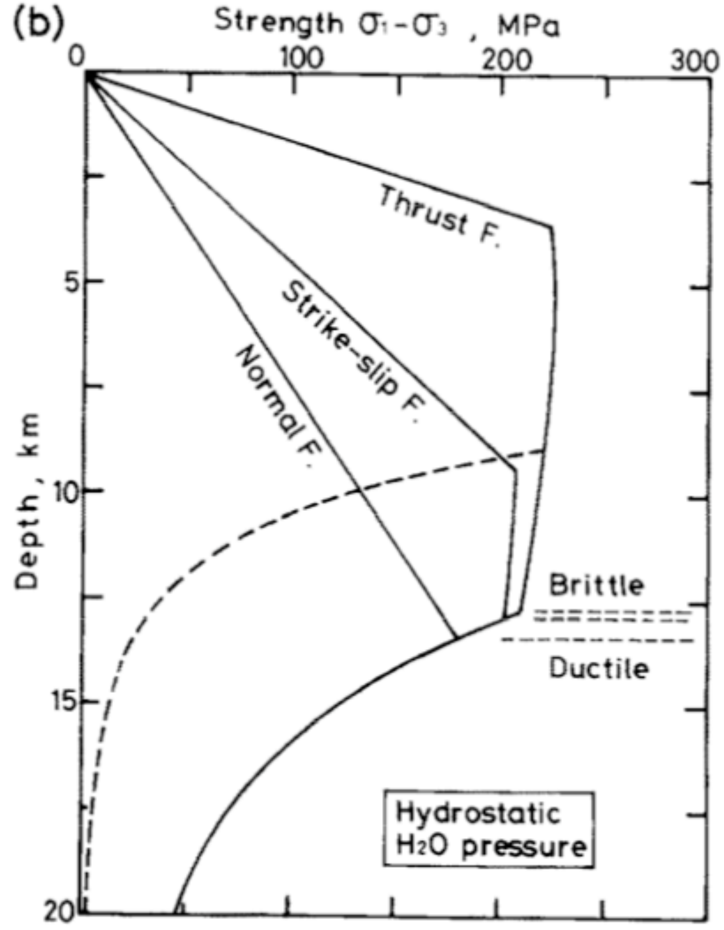


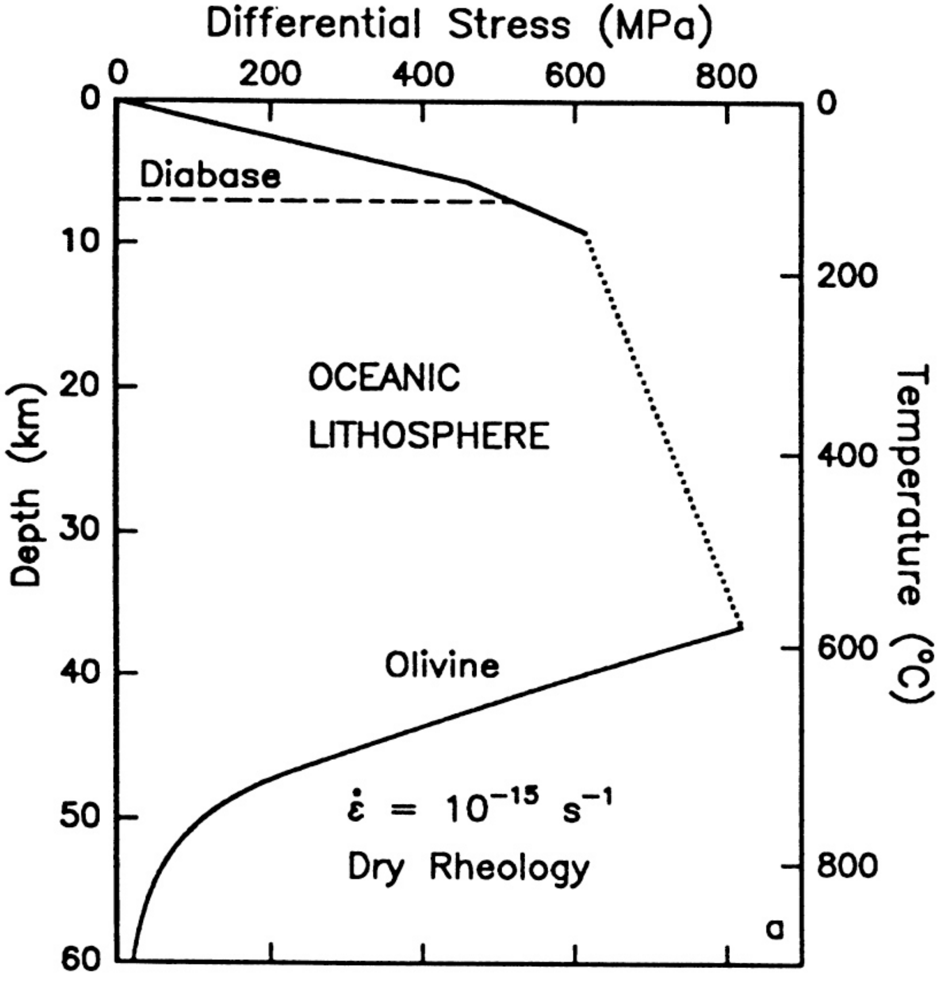
Fig. 6. Effect of confining pressure on the strength of Sleaford Bay clinopyroxenite tested in triaxial compression (S. H. Kirby and A. K. Kronenberg, unpublished data, 1978): (*a*) stress-strain curves, (*b*) ultimate strength or stress at 10% strain as a function of confining pressure.

Strength profile of lithosphere

Continental (granite): Shimada 1993



Oceanic: Kohlstedt 1995



Equations

- Boussinesq, infinite Prandtl number

$$\nabla \bullet \left(\eta\left(v_{i,j} + v_{j,i}\right)\right) - \nabla P = Ra.T\hat{z}$$

$$\eta_{eff} = \min\left\lfloor \eta(T), \frac{\sigma_{yield}}{2\dot{e}} \right\rfloor$$

$$\nabla \bullet \vec{v} = 0 \qquad \frac{\partial T}{\partial t} = \kappa \nabla^2 T - \vec{v}.\nabla T + H$$

Rayleigh number

$$Ra = \frac{\text{advection " velocity"}}{\text{diffusion "velocity"}}$$

$$= \left(\frac{\rho g \alpha \Delta T D^2}{\eta}\right) \Big/ \left(\frac{D}{\kappa}\right)$$

$$= \frac{\rho g \alpha \Delta T D^3}{\eta \kappa}$$

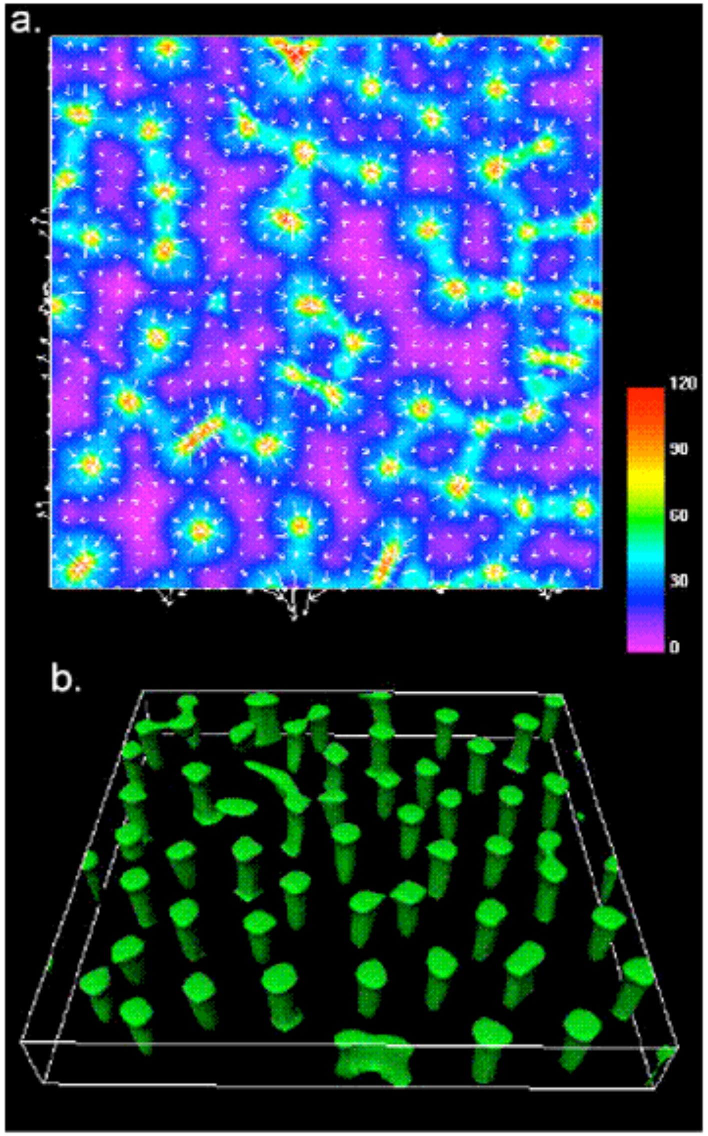
- As planet cools, Ra decreases mainly because h increases

Constant Viscosity convection

- Surface strain rate

- Cold temperature anomalies

Geochemistry Geophysics Geosystems G3



34 MPa

Continuous
”plates”

70 MPa

86 MPa

120 MPa

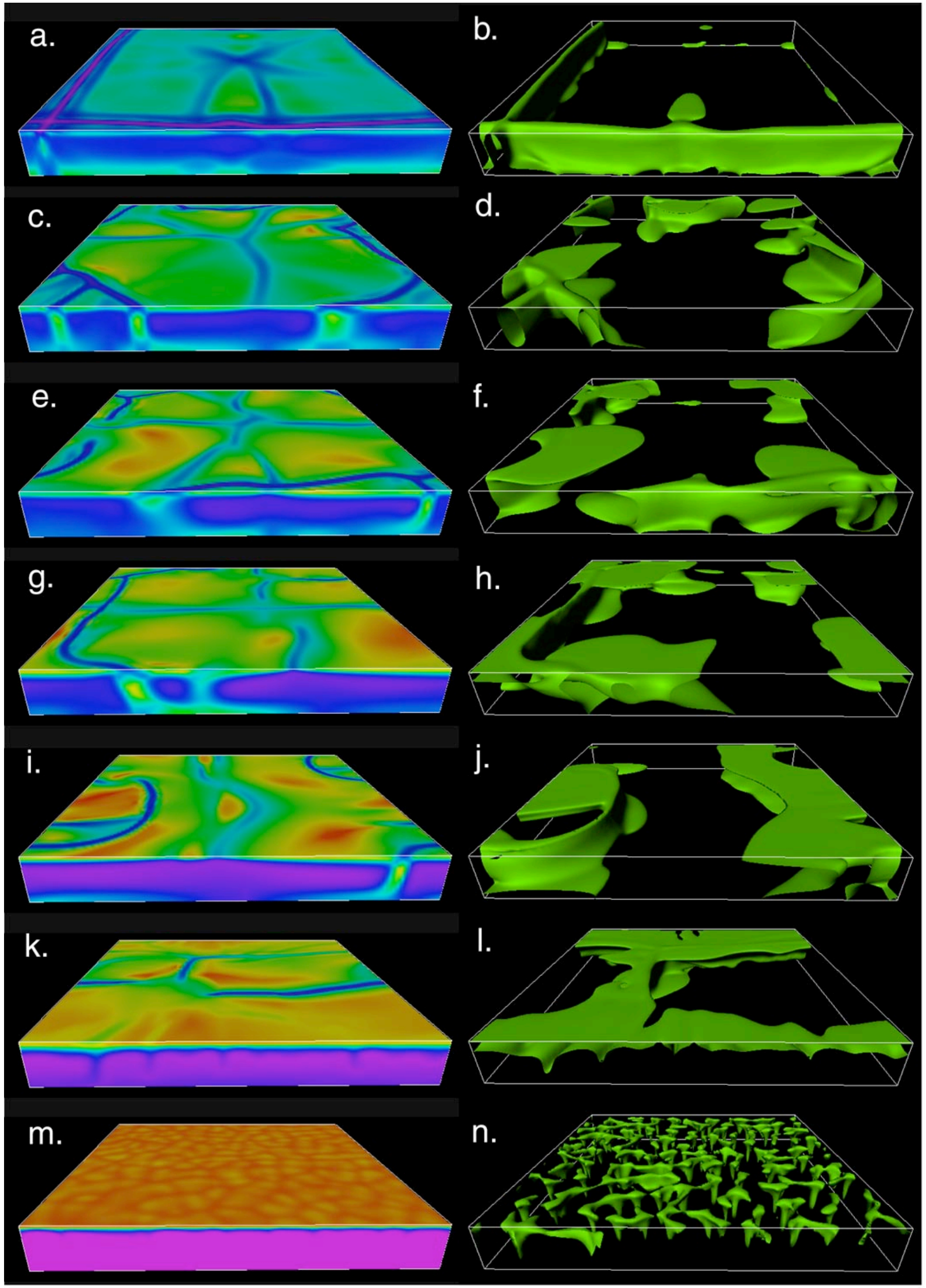
Episodic
plates

168 MPa

200 MPa

Rigid lid

340 MPa



Surface Strain Rate and Velocity

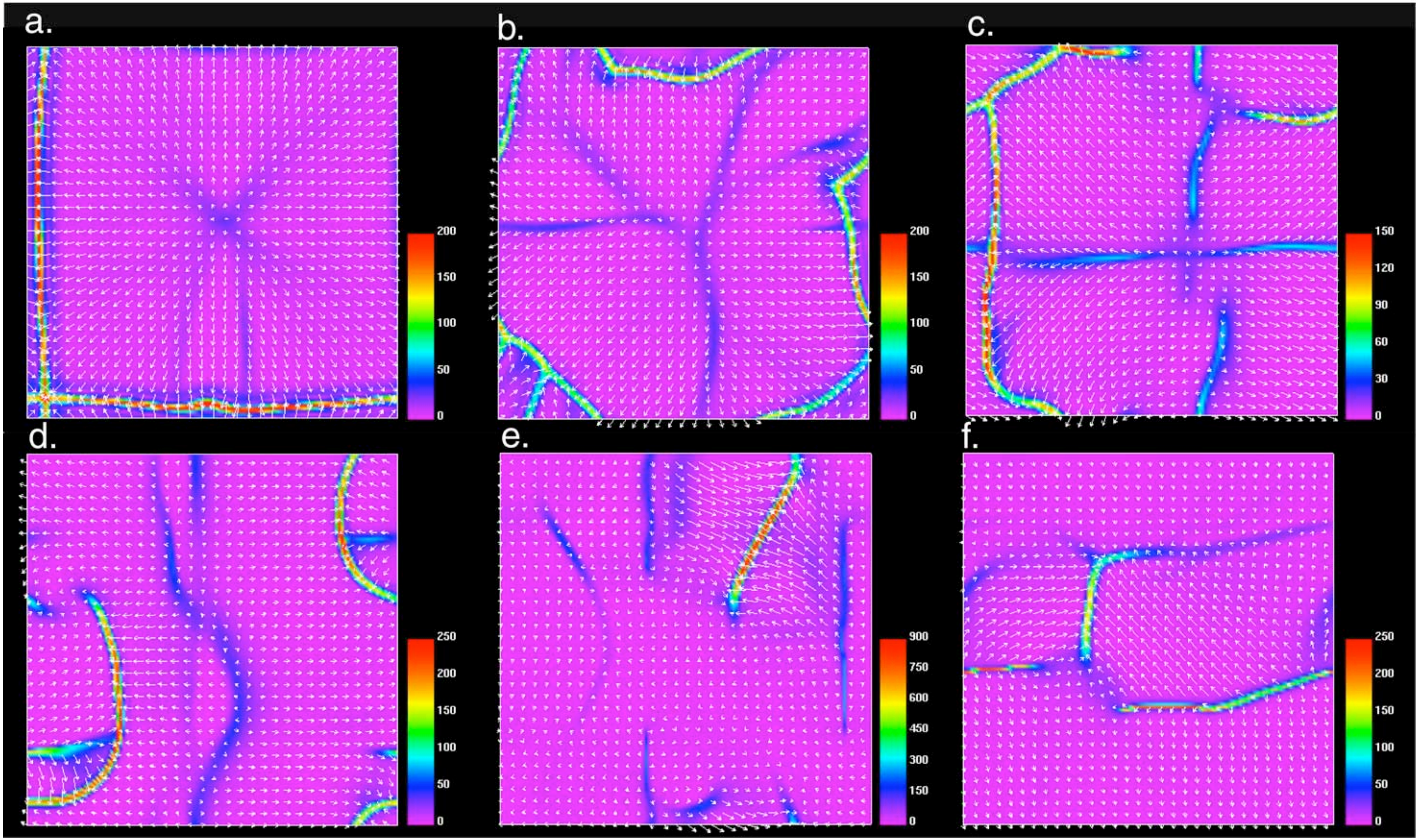
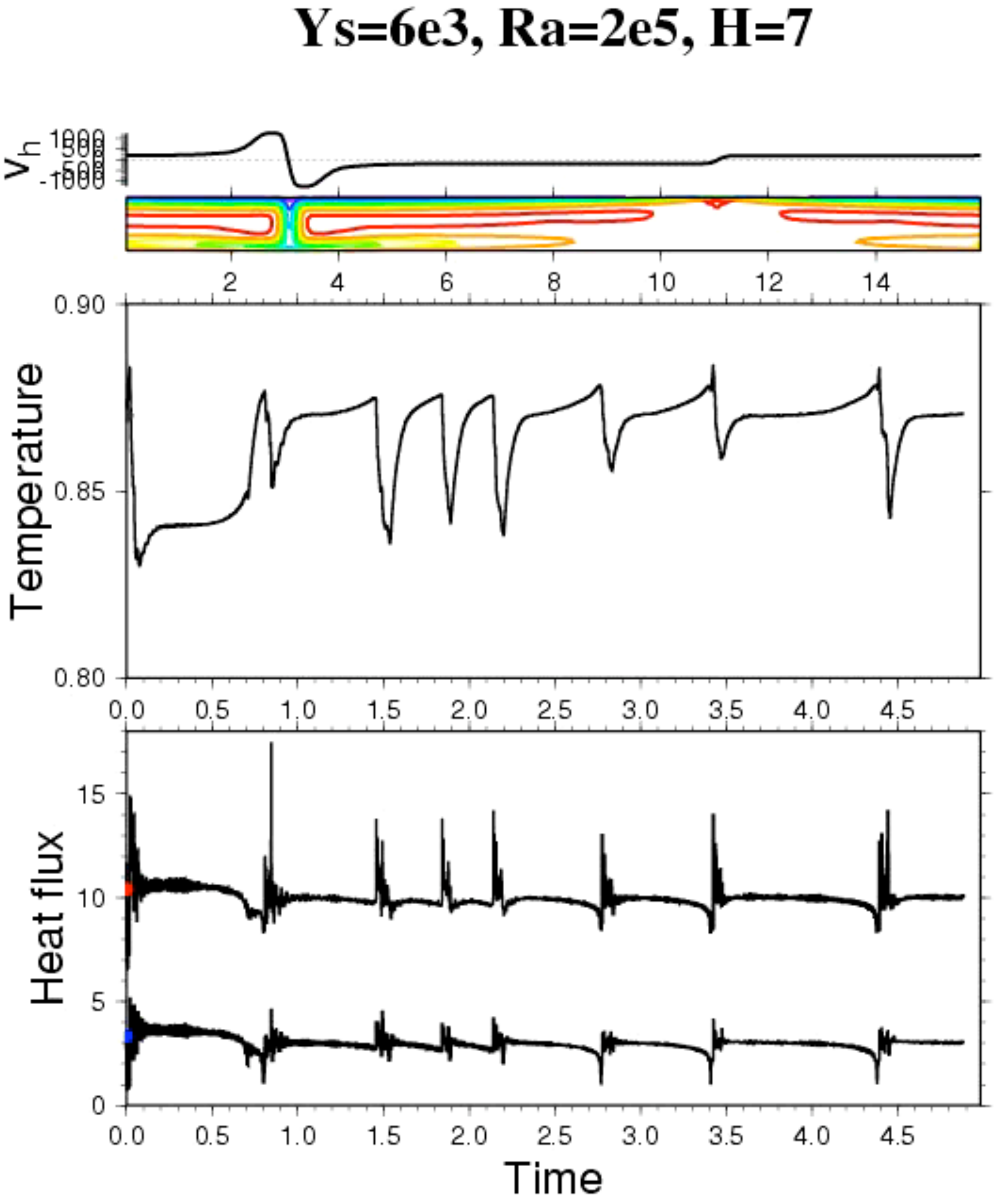
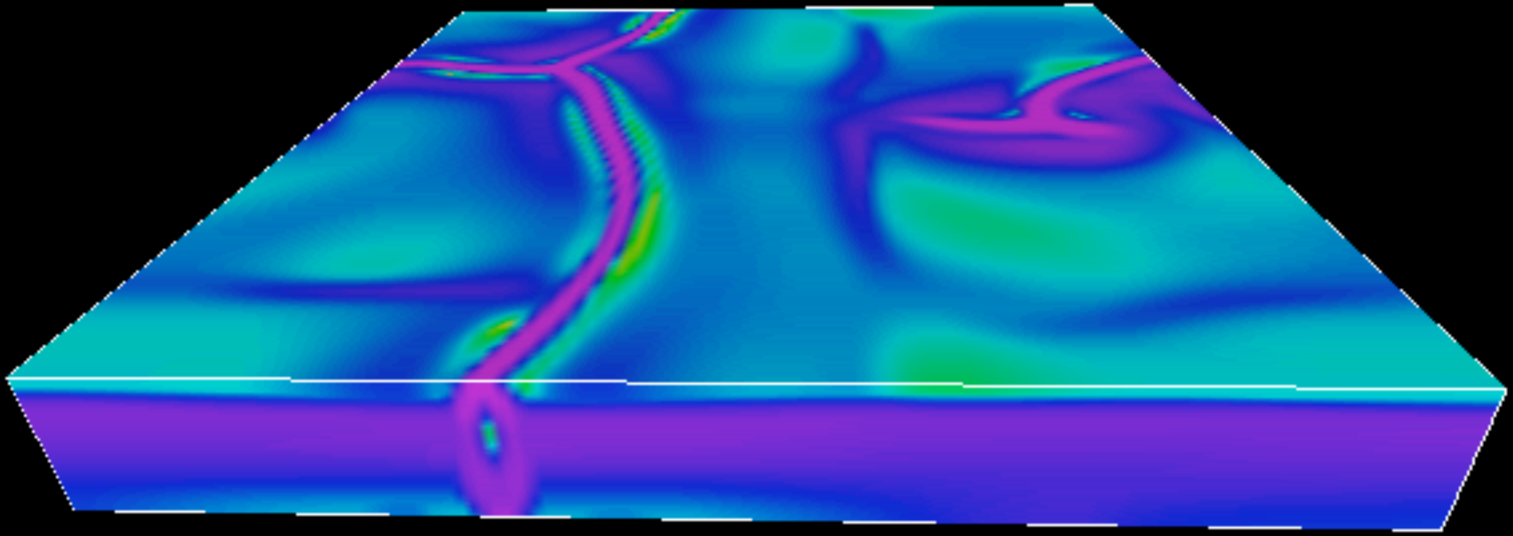


Plate boundary jumps

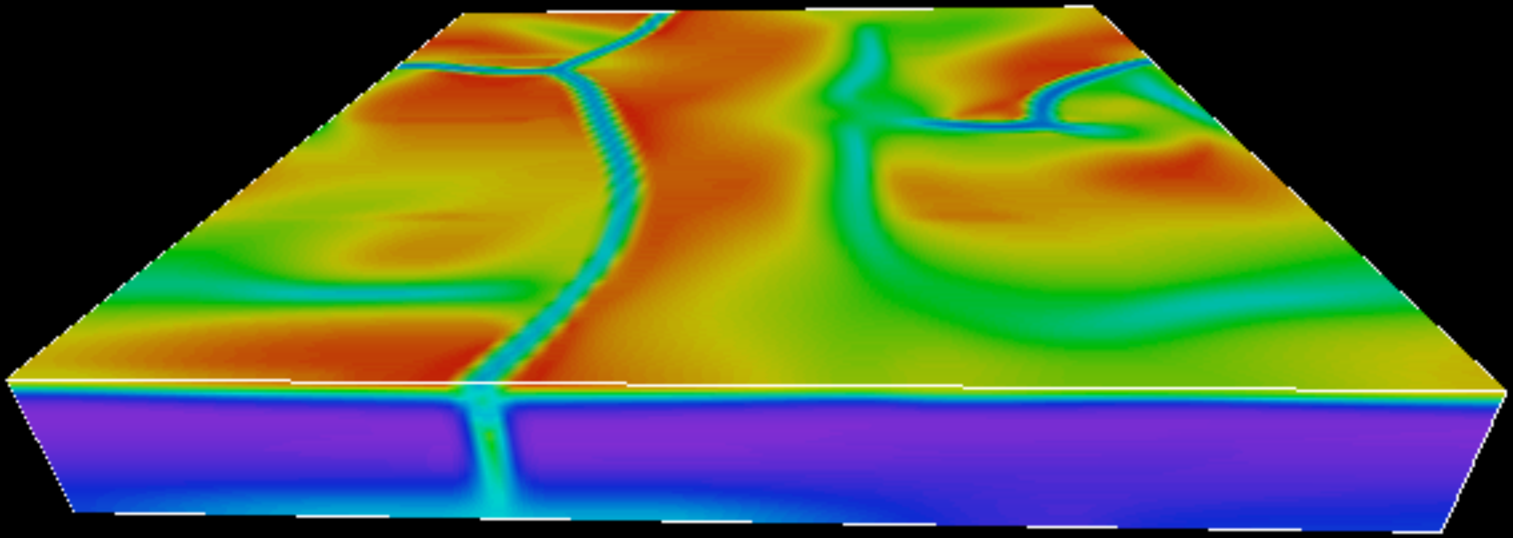
(movie by S. Labrosse)

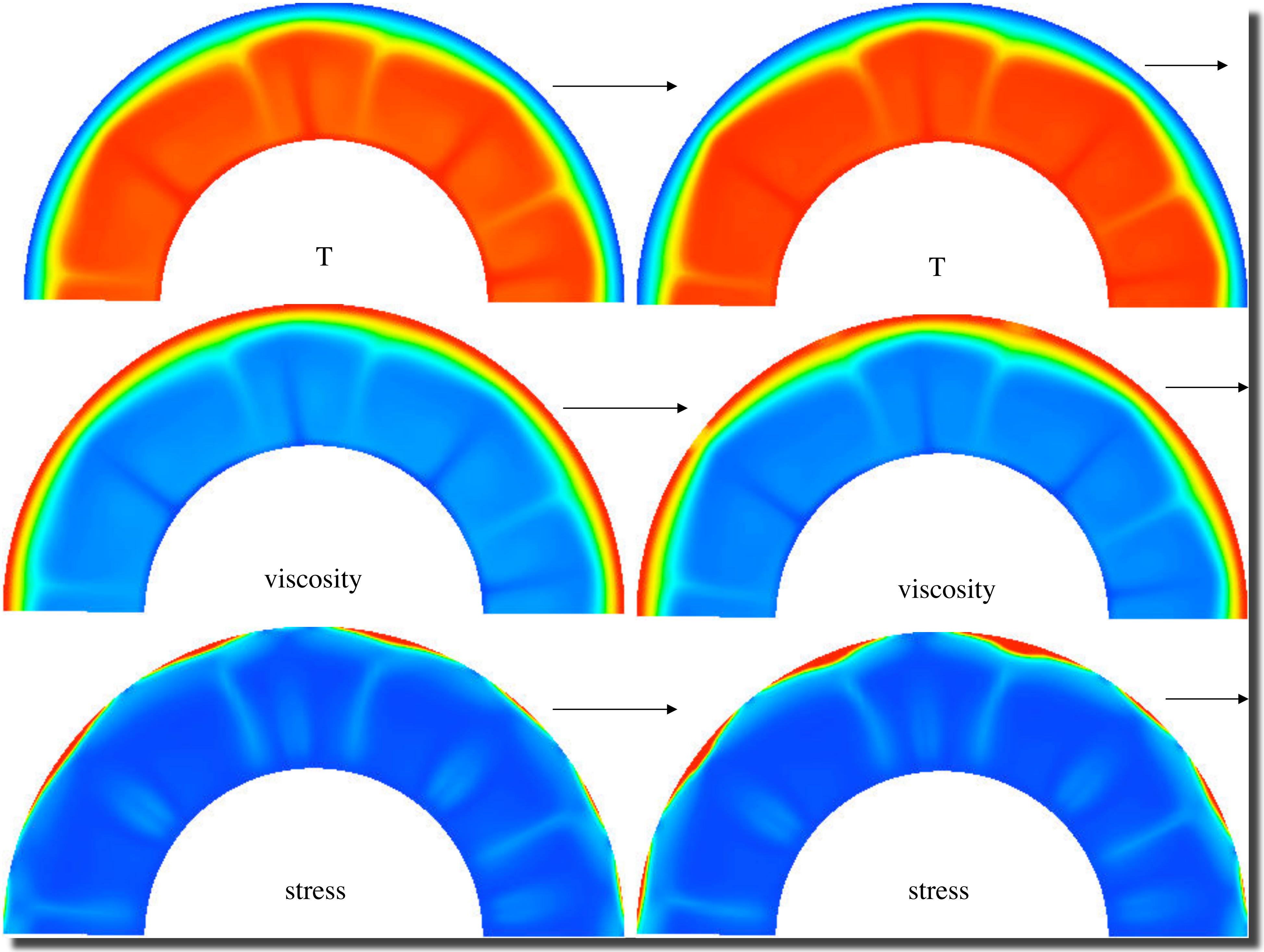


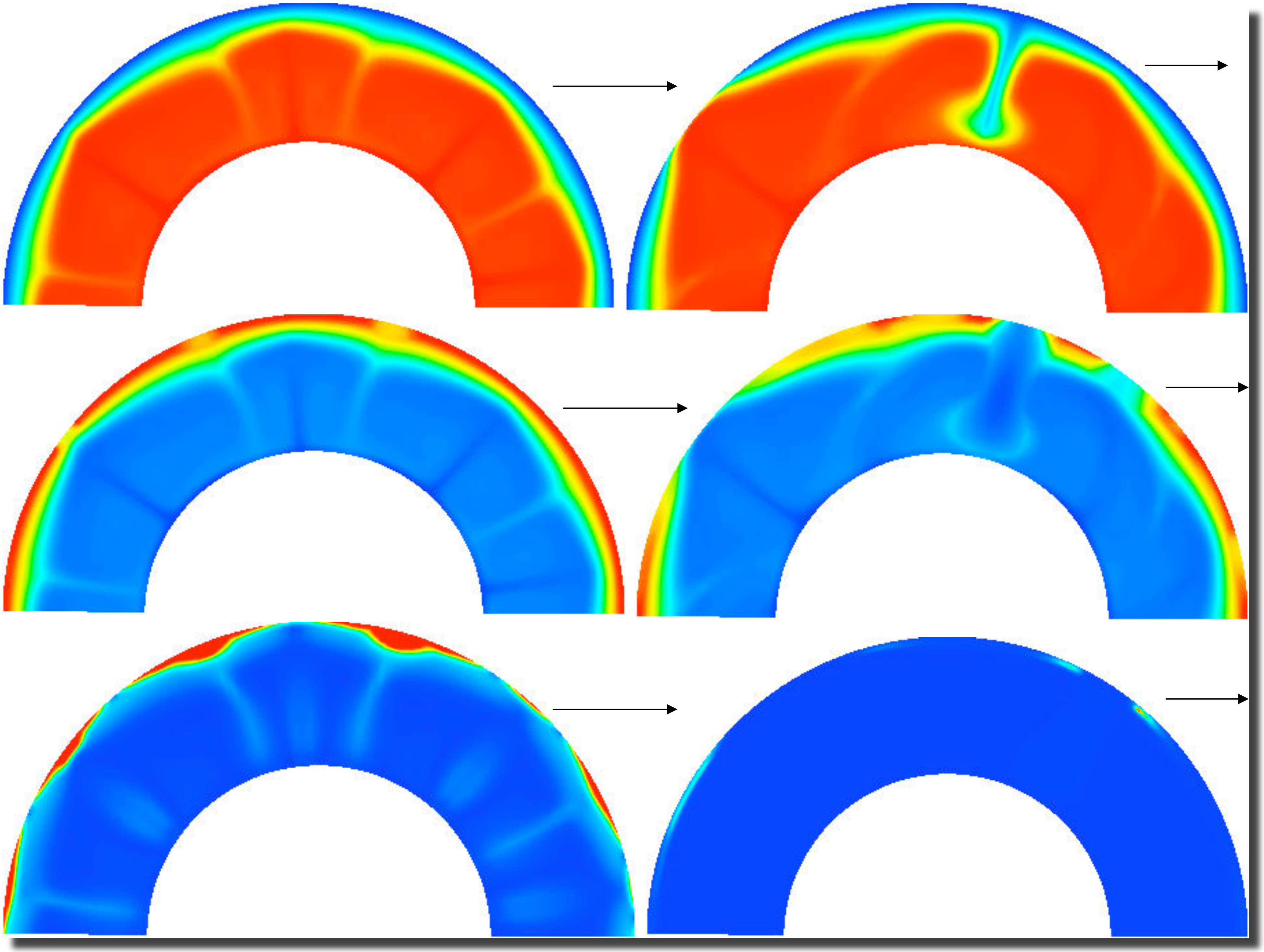
Smoothly-evolving plates

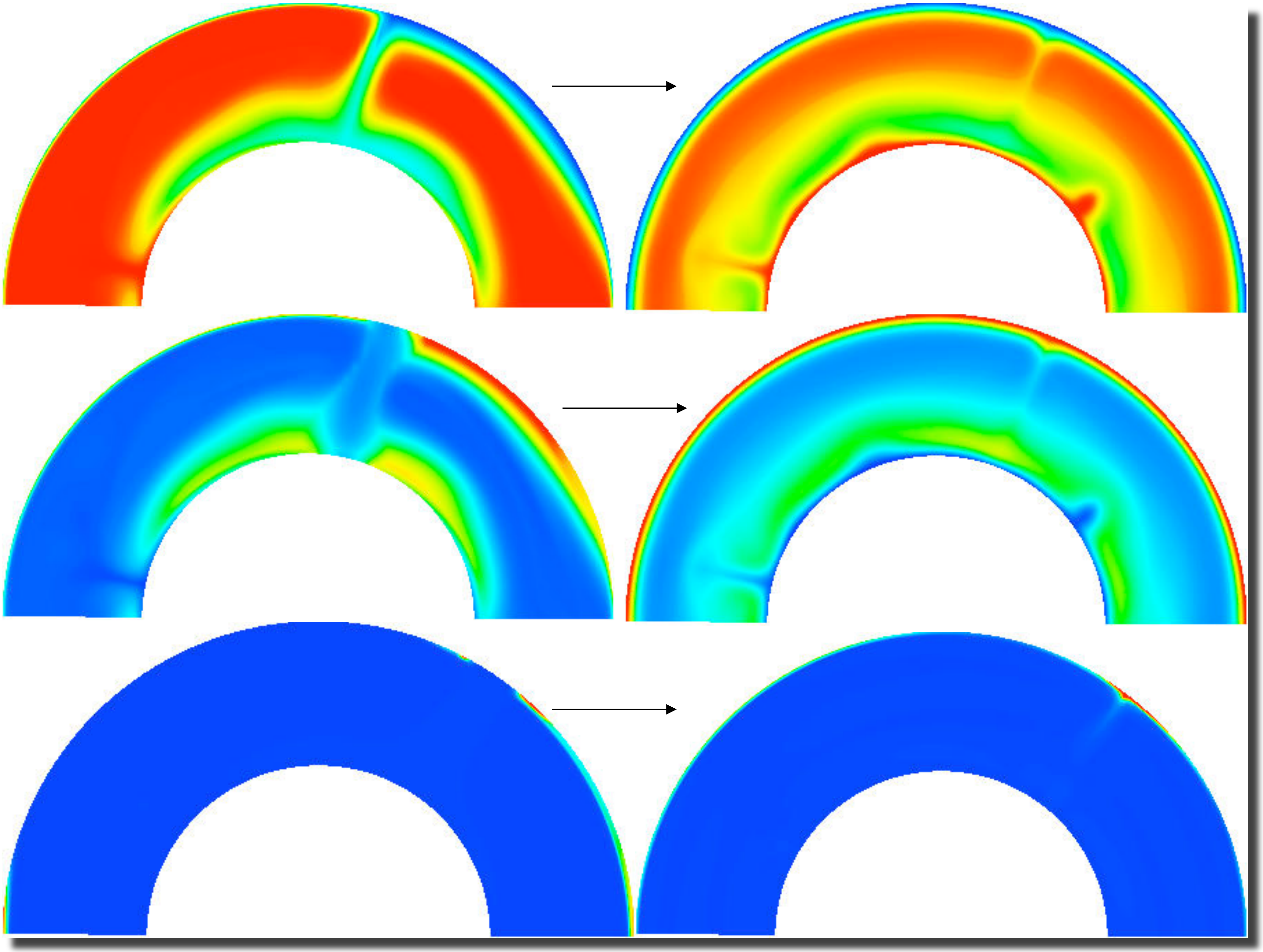


Episodic regime



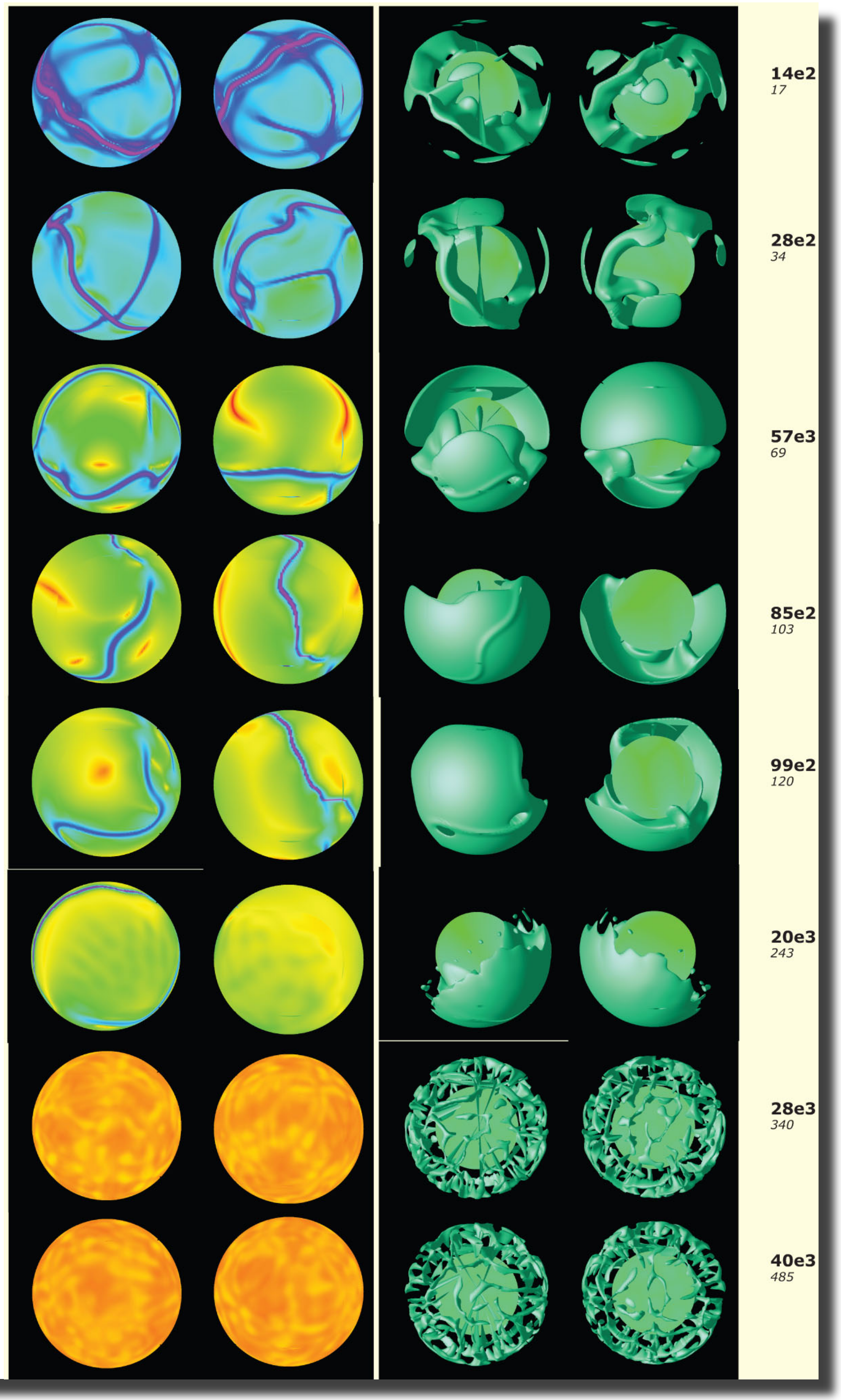

T
T
viscosity
viscosity
stress
stress

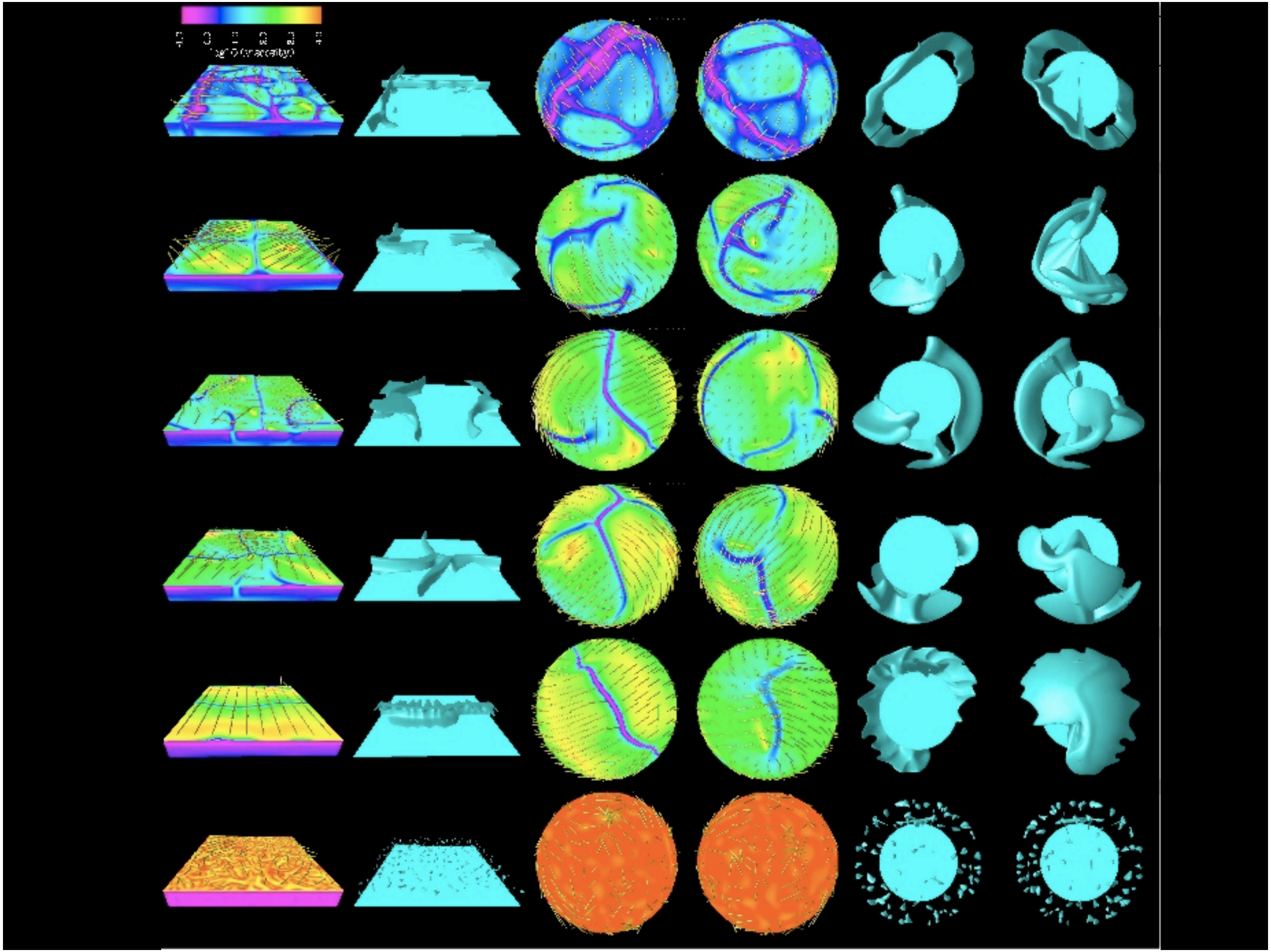


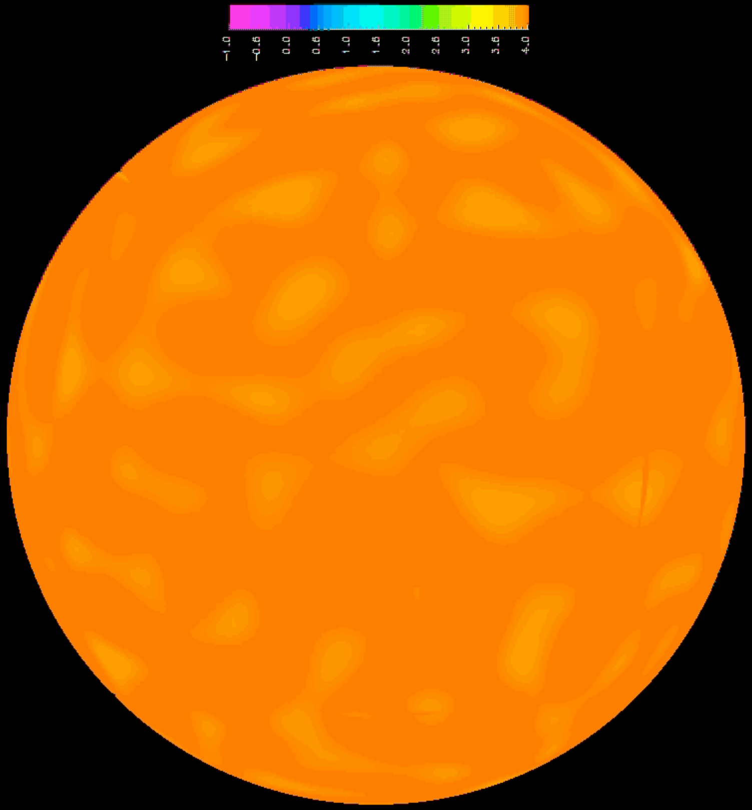


Spherical geometry versions

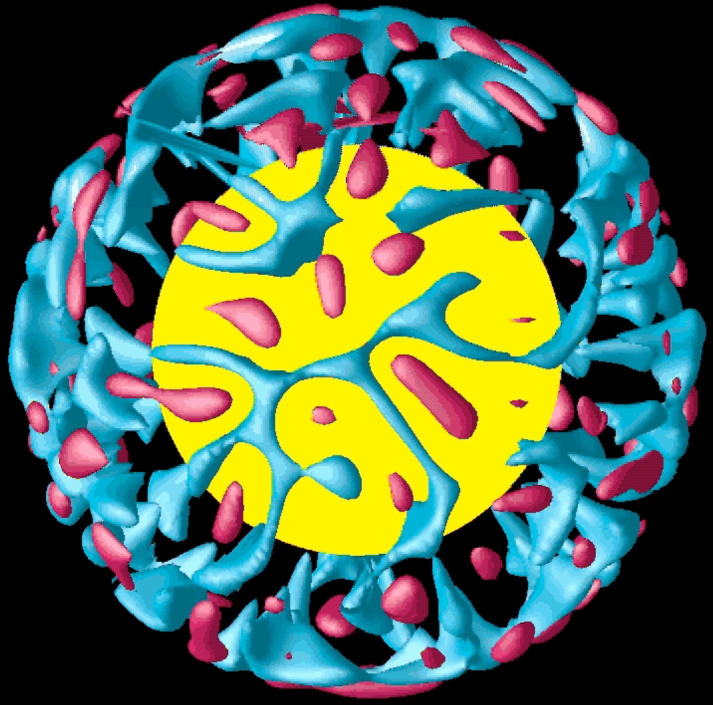
Hein van Heck & me, GRL 2008

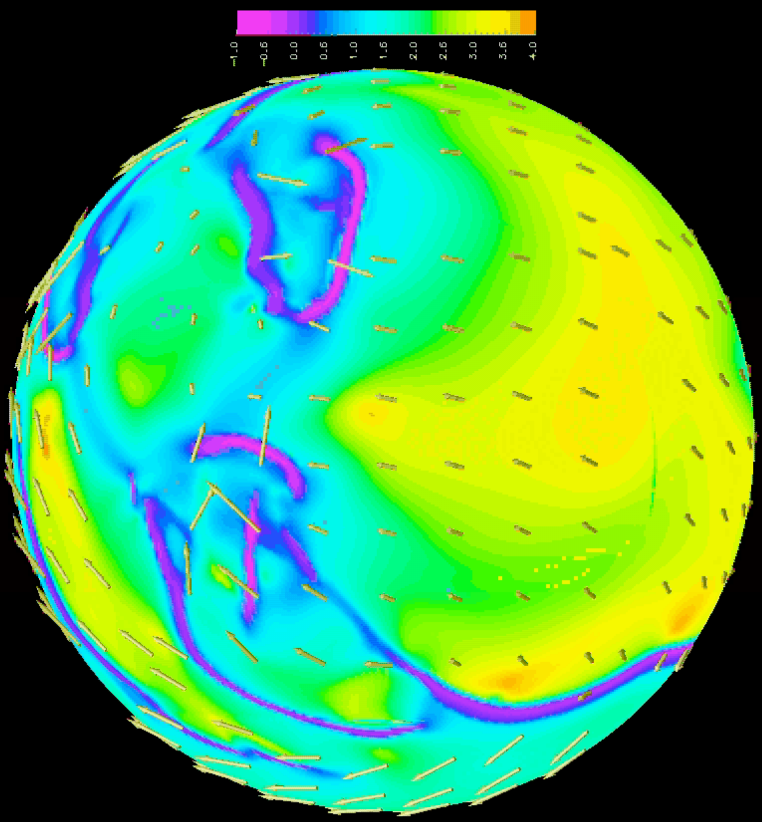


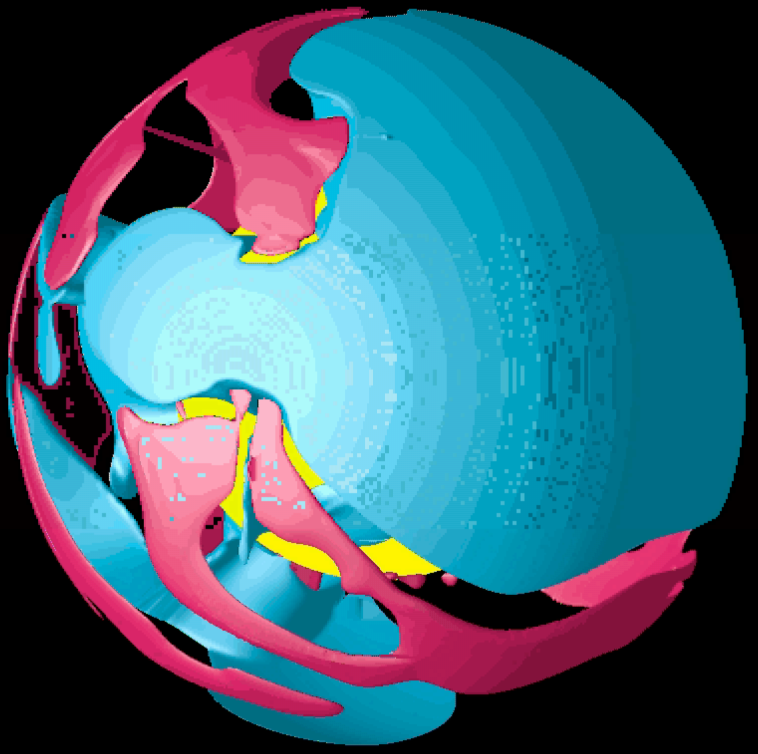

-1.0
0.0
1.0
2.0
3.0
4.0
log10 (viscosity)

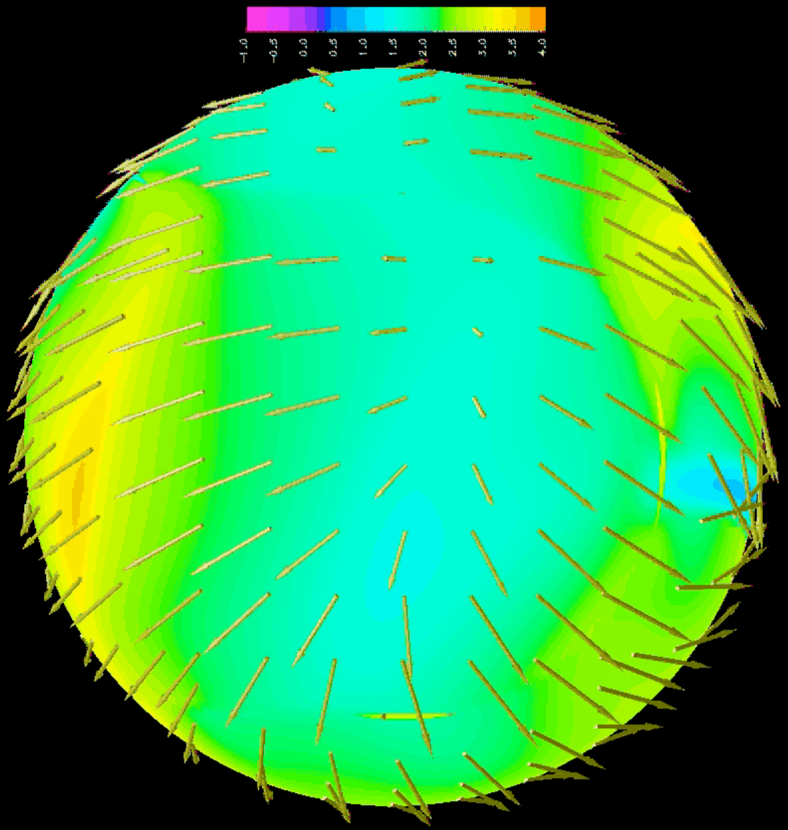


Yield Stress = 3.5*10000 (420 MPa)




-1.0
-0.5
0.0
0.5
1.0
1.5
2.0
2.5
3.0
3.5
4.0

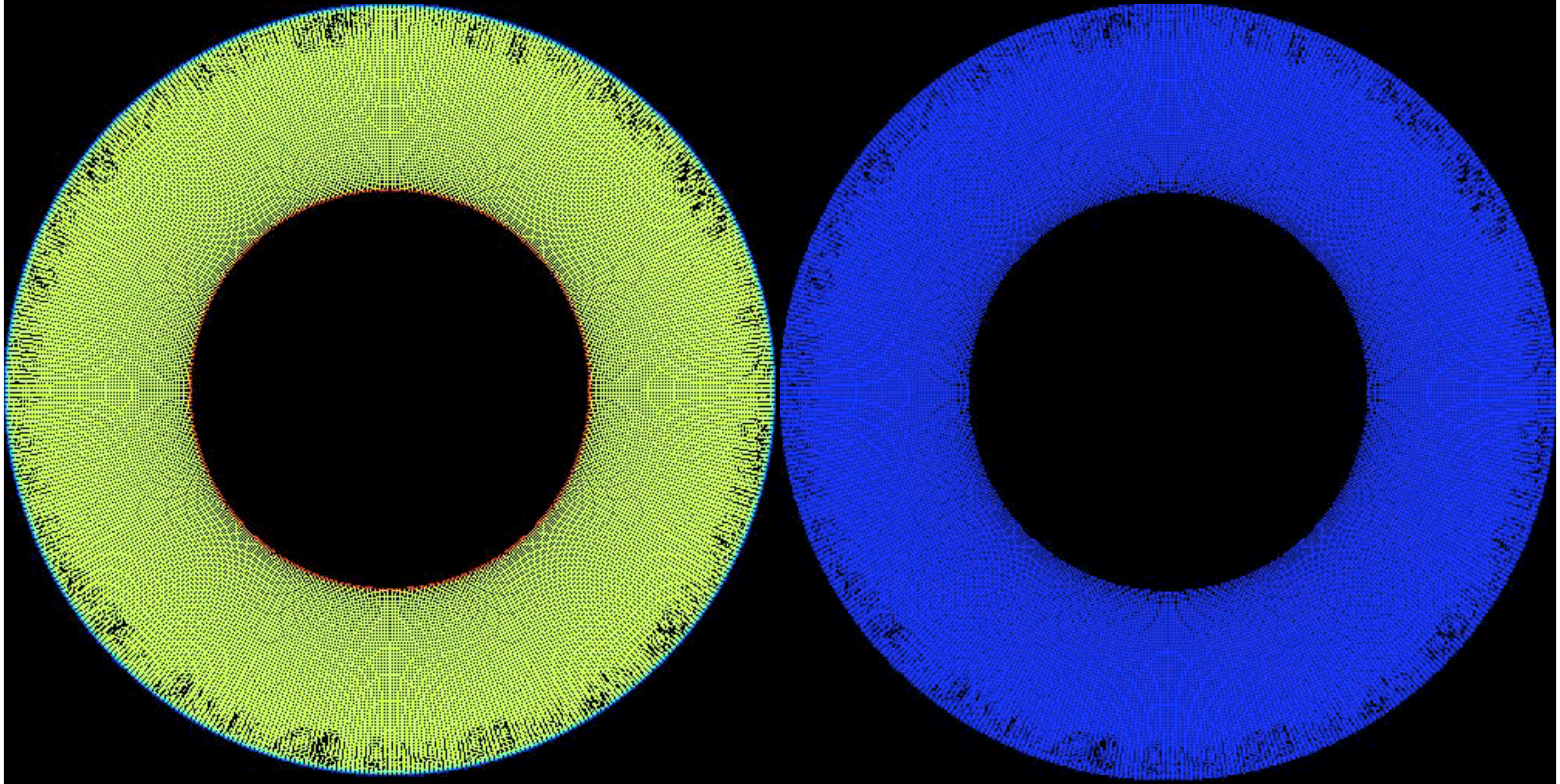




Yield Stress = 8.5*1000 (102 MPa)

Application to Venus

Episodic subduction with crustal production



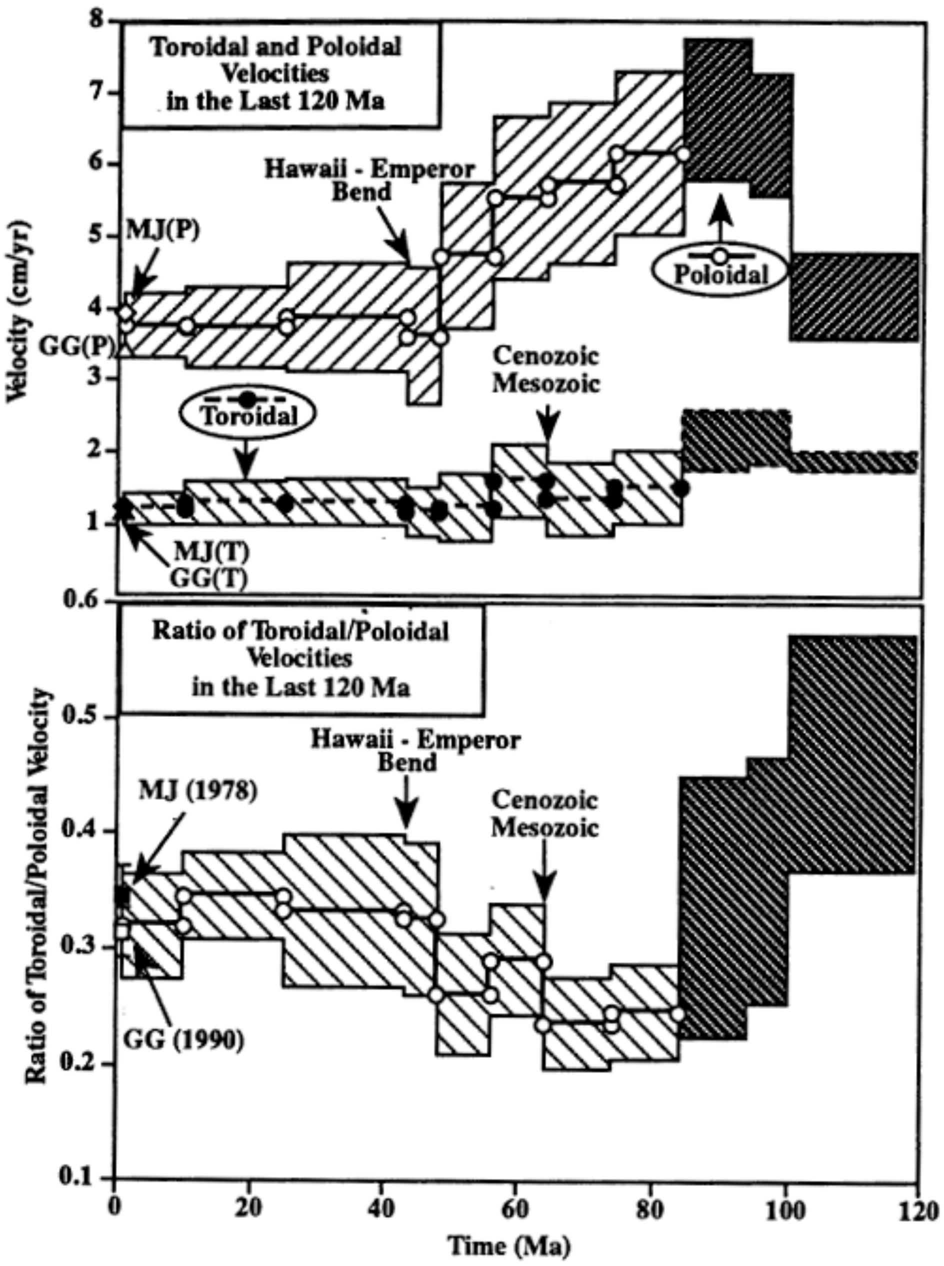
by M. Armann

Quantifying plate tectonics

- 'Plateness': Most deformation focused in narrow zones ~15% of surface area (Stein)
- Significant **toroidal** motion
- Plates are **mobile** (velocity similar to underlying mantle)
- Spreading centers: passive, symmetric
- Subduction: single-sided
- Strike-slip boundaries

- Earth’s Tor/Pol ratio ~0.3-0.5 (excluding net rotation)
- **Toroidal**: rotation in horizontal plane - associated with strike-slip plate boundaries & plate rotation
- **Poloidal**: divergence or vertical in a horizontal plane- driven by convection

Lithgow-Bertelloni et al., 1993



Observed plate divergence and vorticity

- Divergence: Poloidal field
- Vorticity: Toroidal field

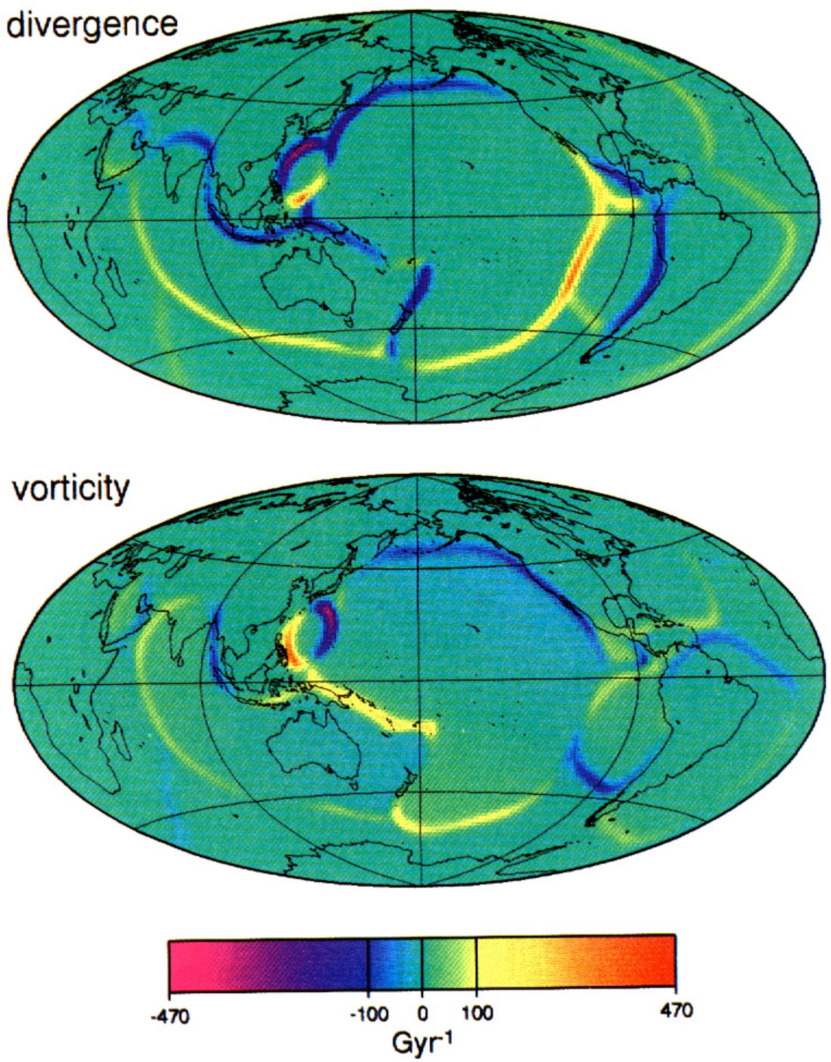
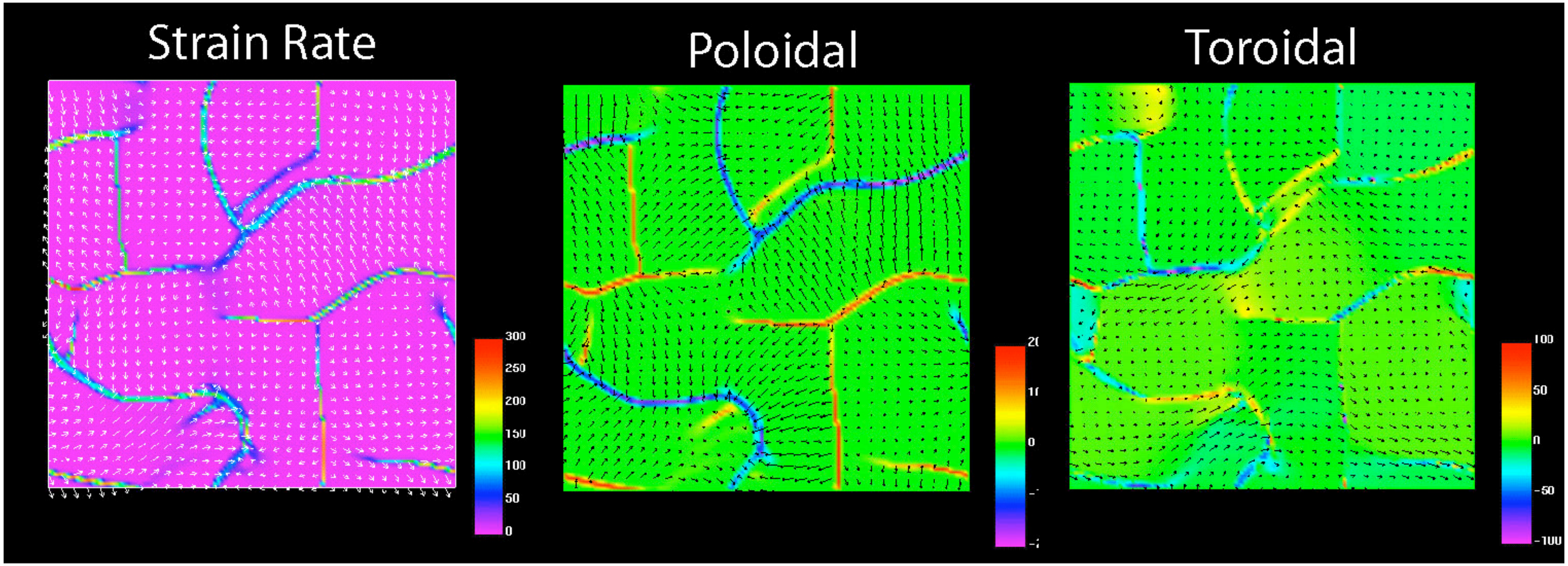


Figure 5. Horizontal divergence and radial vorticity for the continuous plate model with margin width $R\delta = 200$ km.

Bercovici and Wessel, 1994

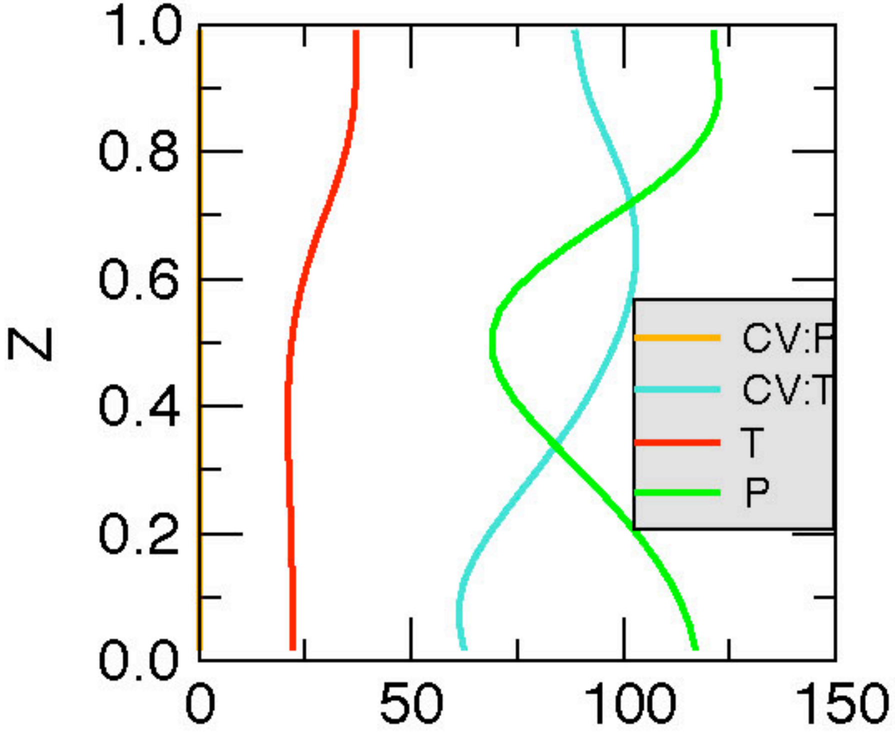
Surface Strain rate

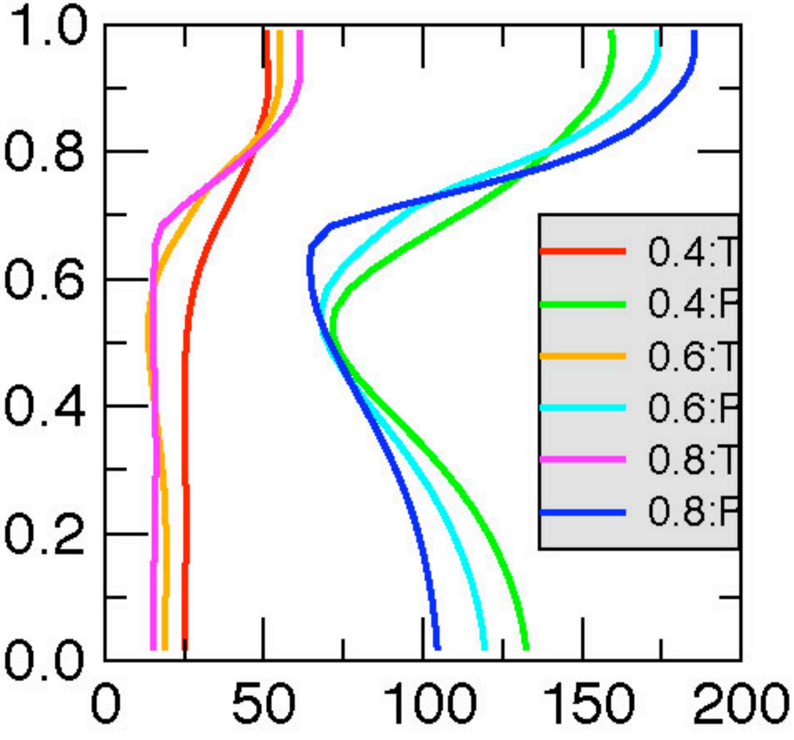


Divergence

Vorticity

Pol- & Toroidal with depth


Simple yield stress and constant viscosity


Increasing melting

Diagnostics

Plateness: $P = (0.6 - f_{80})/0.6$

f_{80}=fraction of surface area in which highest 80% of integrated strain rate occurs

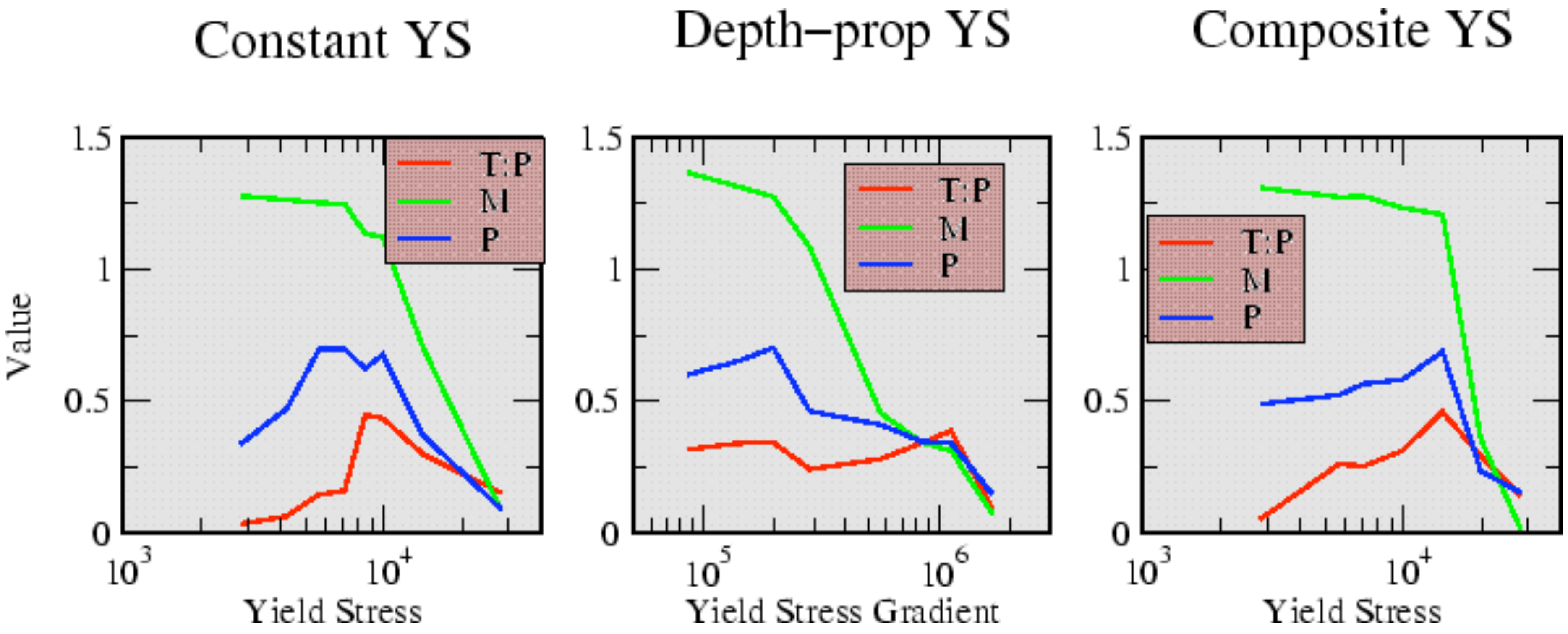
$P \approx 0$ for constant-viscosity, internally-heated convection.

$P=1$ for surface deformation localized into infinitely-narrow zones.

Surface **T:P ratio**: $T : P = \sqrt{\dfrac{< v_{tor}^2 >}{< v_{pol}^2 >}}$

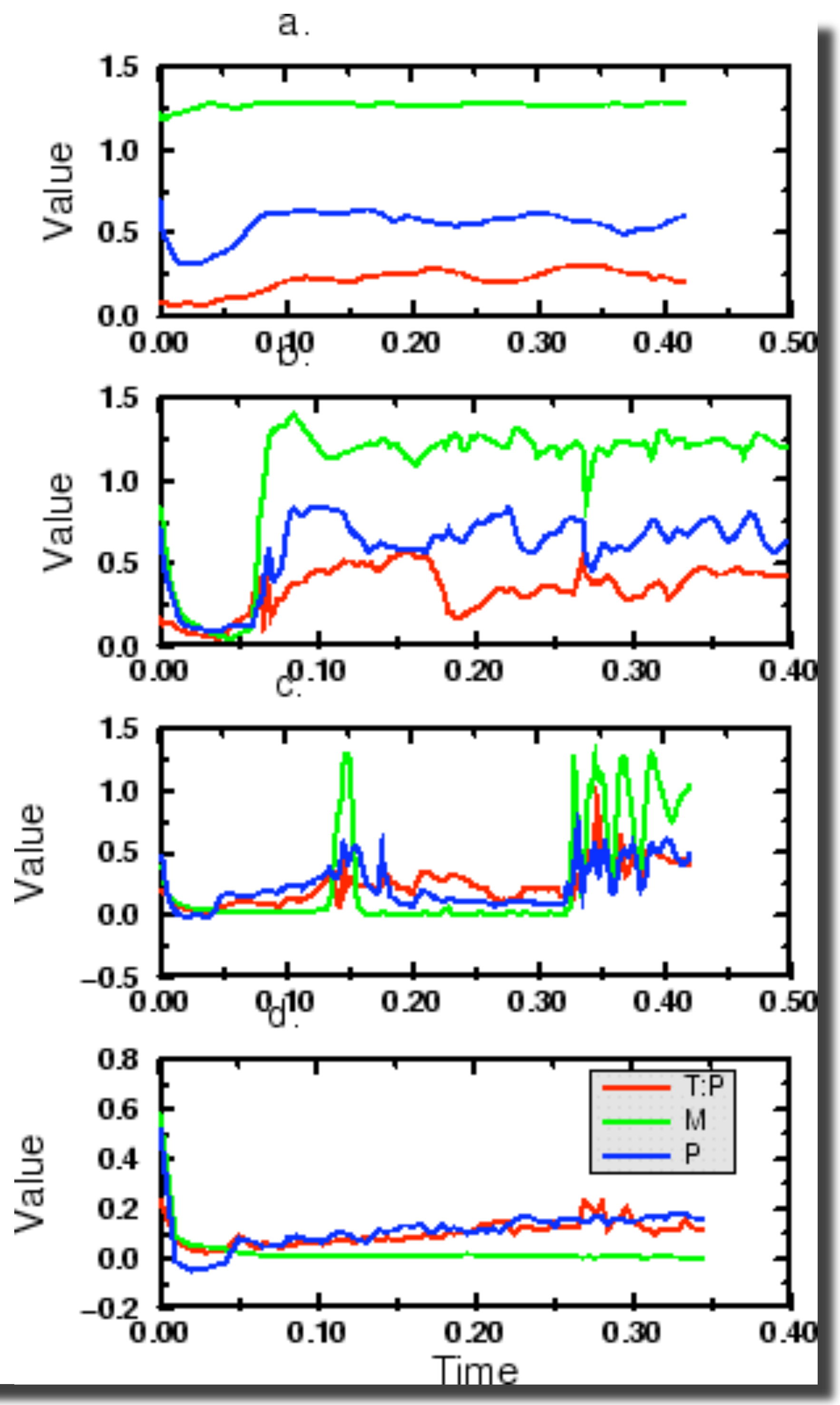
Surface **Mobility**: $M = \dfrac{(v_{rms})_{surface}}{(v_{rms})_{domain}}$

Scaling of plate diagnostics with Yield Stress



Time-Dependence

- Yield stress increases top to bottom

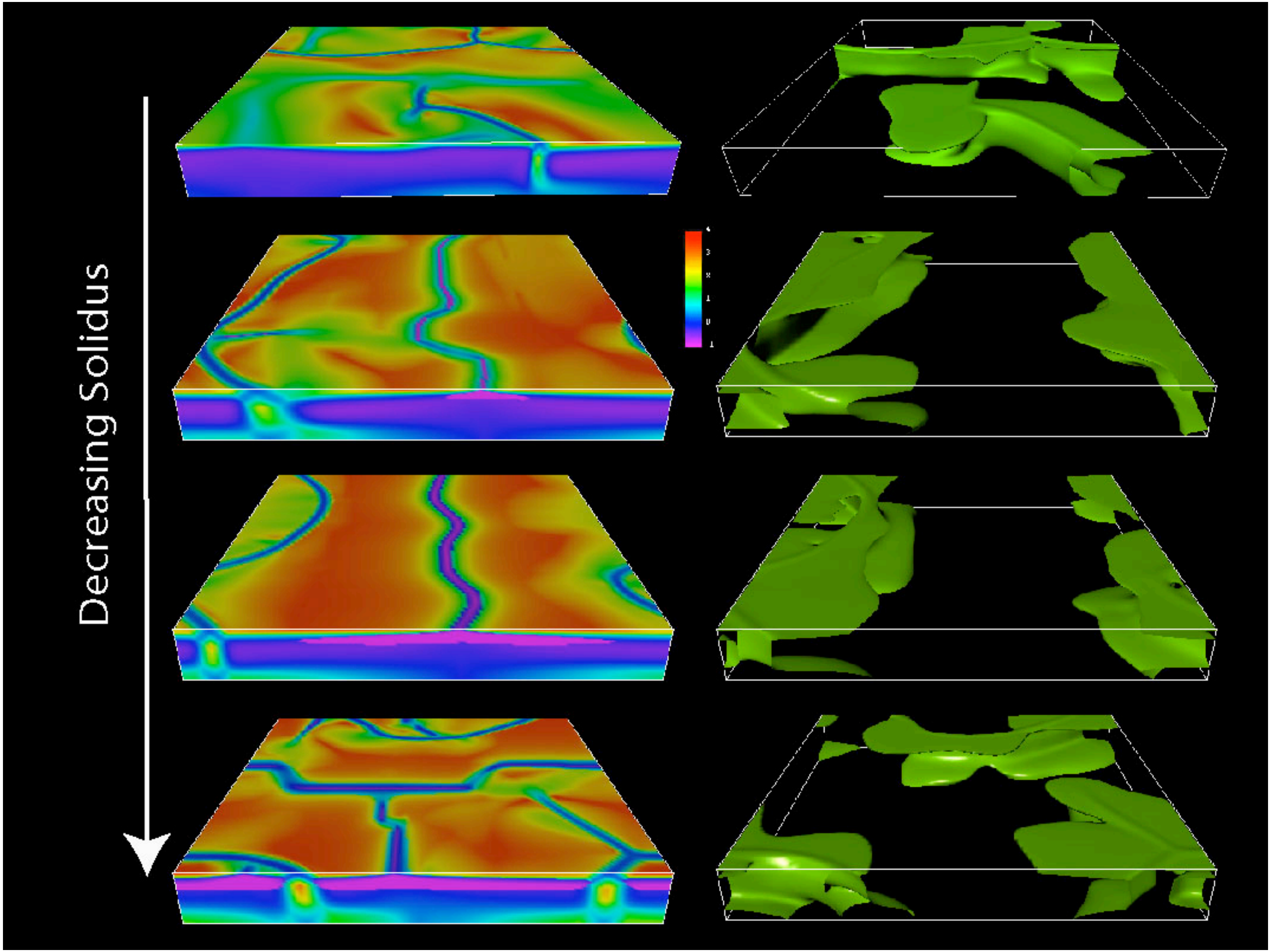


Helpful complexities?

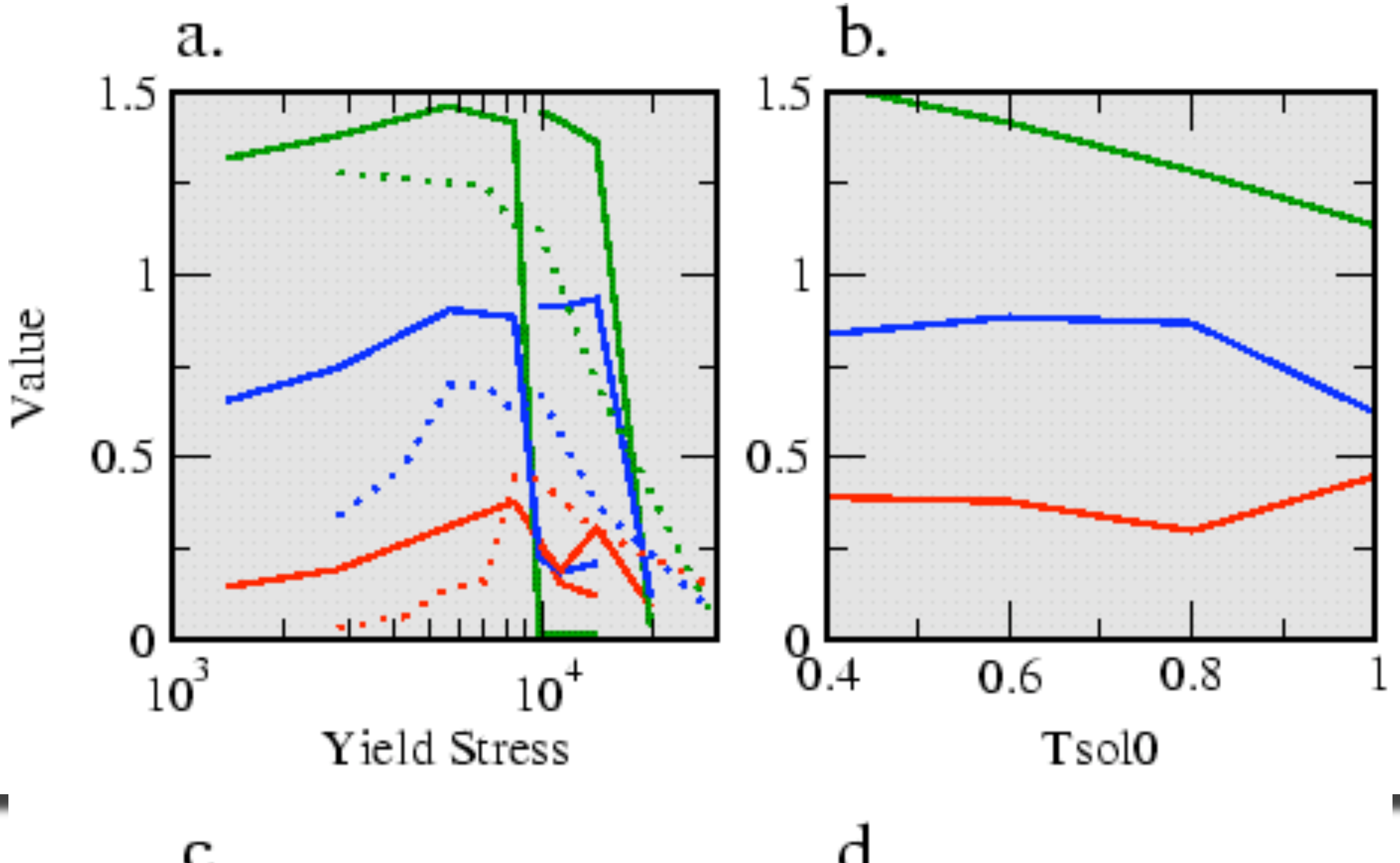
- Low viscosity asthenosphere
- Strain weakening

Does low viscosity beneath the lithosphere help?

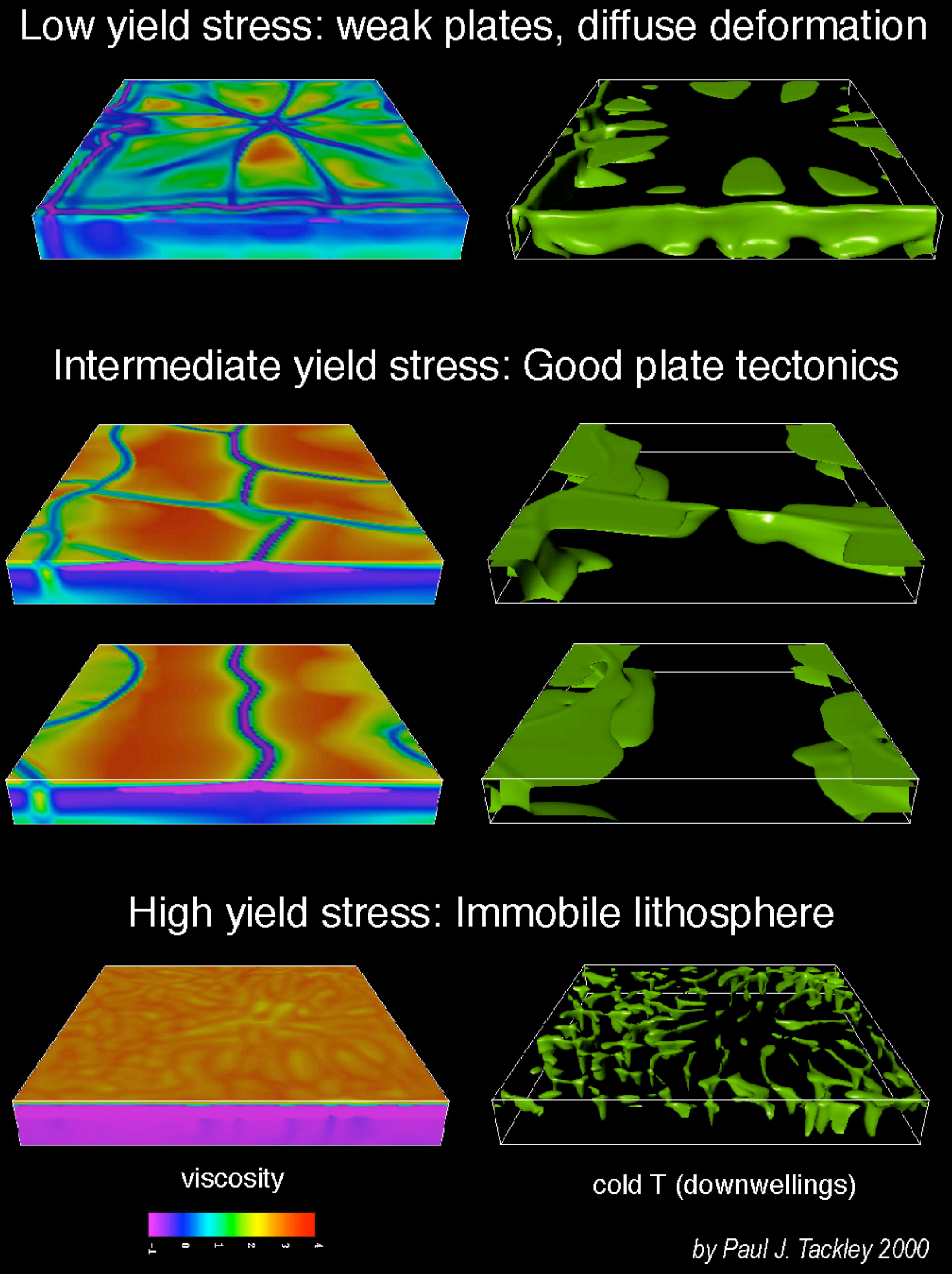
- ‘Asthenosphere’
- Decouples piecewise continuous plate motion from distributed mantle deformation ?
- Want to add in such a way that viscosity is unchanged elsewhere
- Define ‘solidus’ $T=T_0+A$*depth, decrease h by factor 10 when T reaches solidus
- (in reality getting close to solidus is sufficient)


Decreasing Solidus
4
3
2
1
0
-1

Greatly improves plate quality



- Varying yield strength, including asthenosph.



Low yield stress: weak plates, diffuse deformation

Intermediate yield stress: Good plate tectonics

High yield stress: Immobile lithosphere

by Paul J. Tackley 2000

So far…instantaneous rheology

Isn’t history dependence important? i.e., Strain weakening, and healing

Strain weakening?

- Observed in laboratory
- Expected in theory
- Evidenced in the field
- Mechanisms:
  - Dynamic recrystallization => small grains
  - Volatile infiltration + hydration reactions (Bercovici)
  - Viscous dissipation (shear heating)
- Provides positive feedback leading to **strain localization** and narrow shear zones
- Models proposed by Bercovici and Bercovici +Ricard

Shear instability
Constant stress experiment

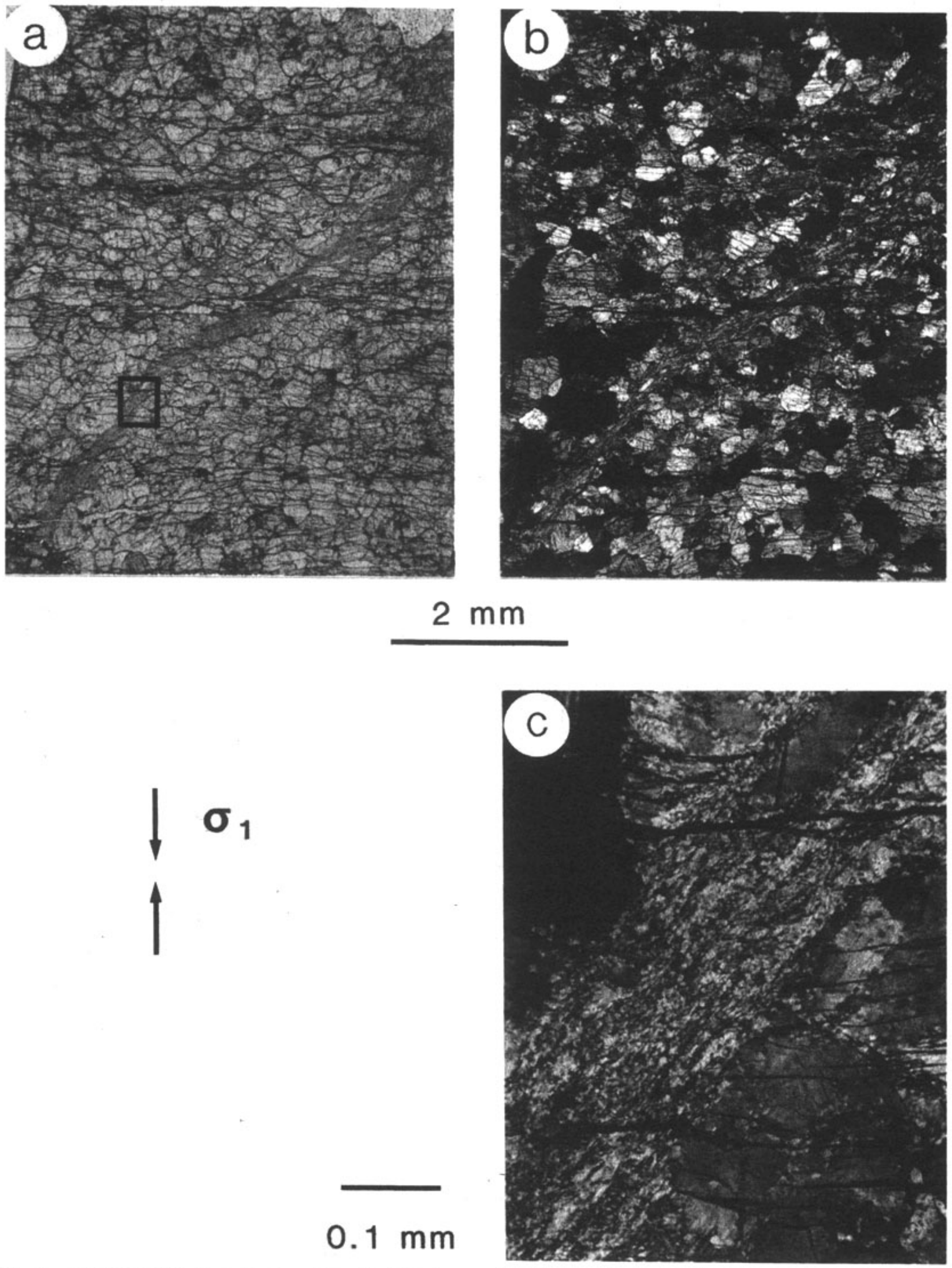


Fig. 14. Ductile shear zone developed in the high-temperature/low strain rate regime. N-318, $T = 1100°C$, $\dot{\varepsilon} = 1.1 \times 10^{-5}$ s^{-1}. (*a*) Plane polarized light. (*b*) Crossed polars. (*c*) High magnification micrograph of region shown in Figure 14*a*. Crossed polars.

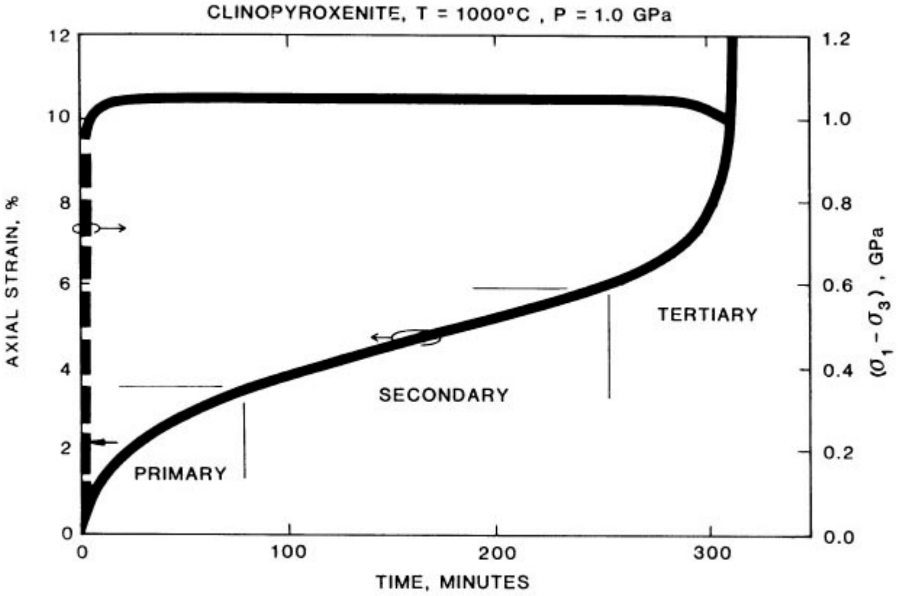


Fig. 9. Creep instability developed in polycrystalline diopside Ca(Mg, Fe)Si$_2$O$_7$. The typical prima
secondary creep stages are followed by a tertiary stage in which strain rates rapid accelerate until th
motor was no longer able to maintain the 1.03 GPa differential stress.

History-dependence: 'Damage' evolution based on Bercovici’s work

$$\frac{dD}{dt} = A\sigma : \dot{e} - R(T)\,D$$

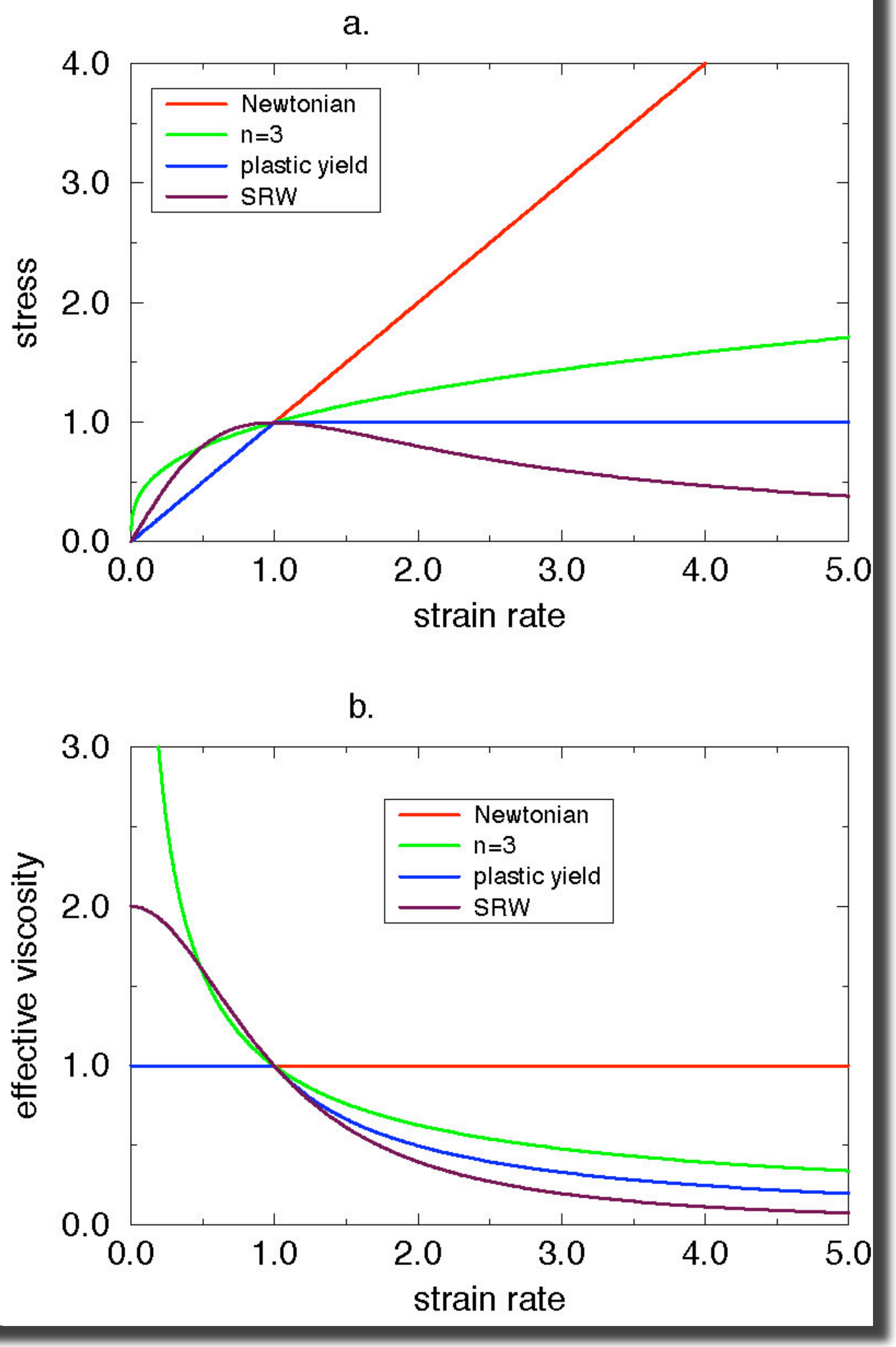
$$\eta = \eta_{undamaged}(1 - D)$$

$$e.g.,\ \ R(T) \propto 1/\sqrt{\eta(\mathrm{T})}$$

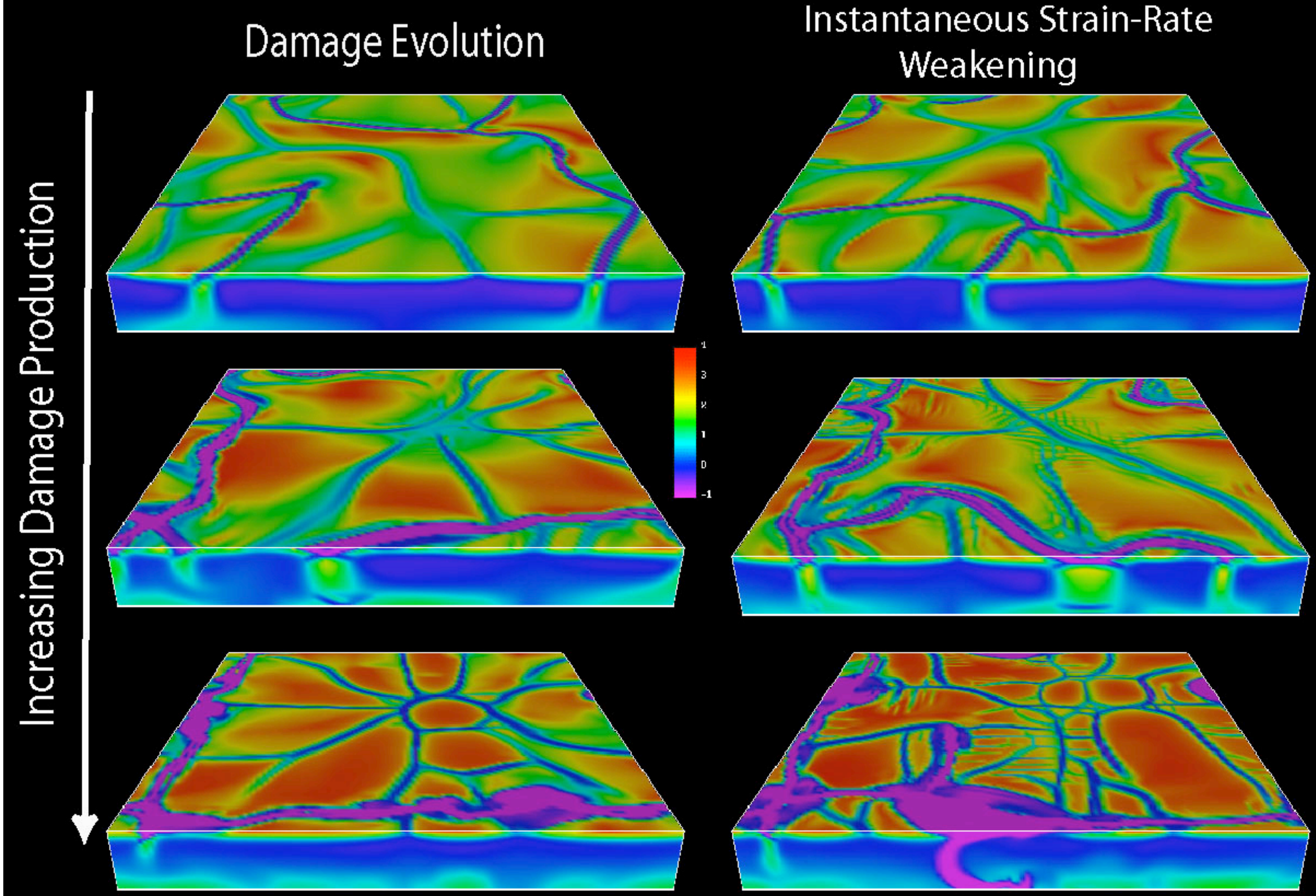
$$\eta = \eta\left(T, z, \dot{e}, history\right)$$

If A and R very large => **strain-rate weakening**

Comparison of various rheologies



Add SW to yielding models



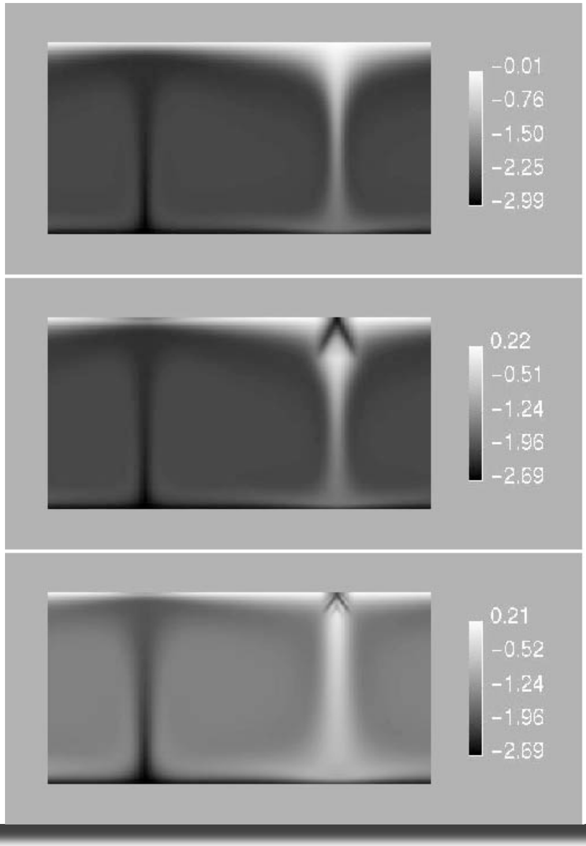
Auth, Bercovici, Christensen (GJI 2003)

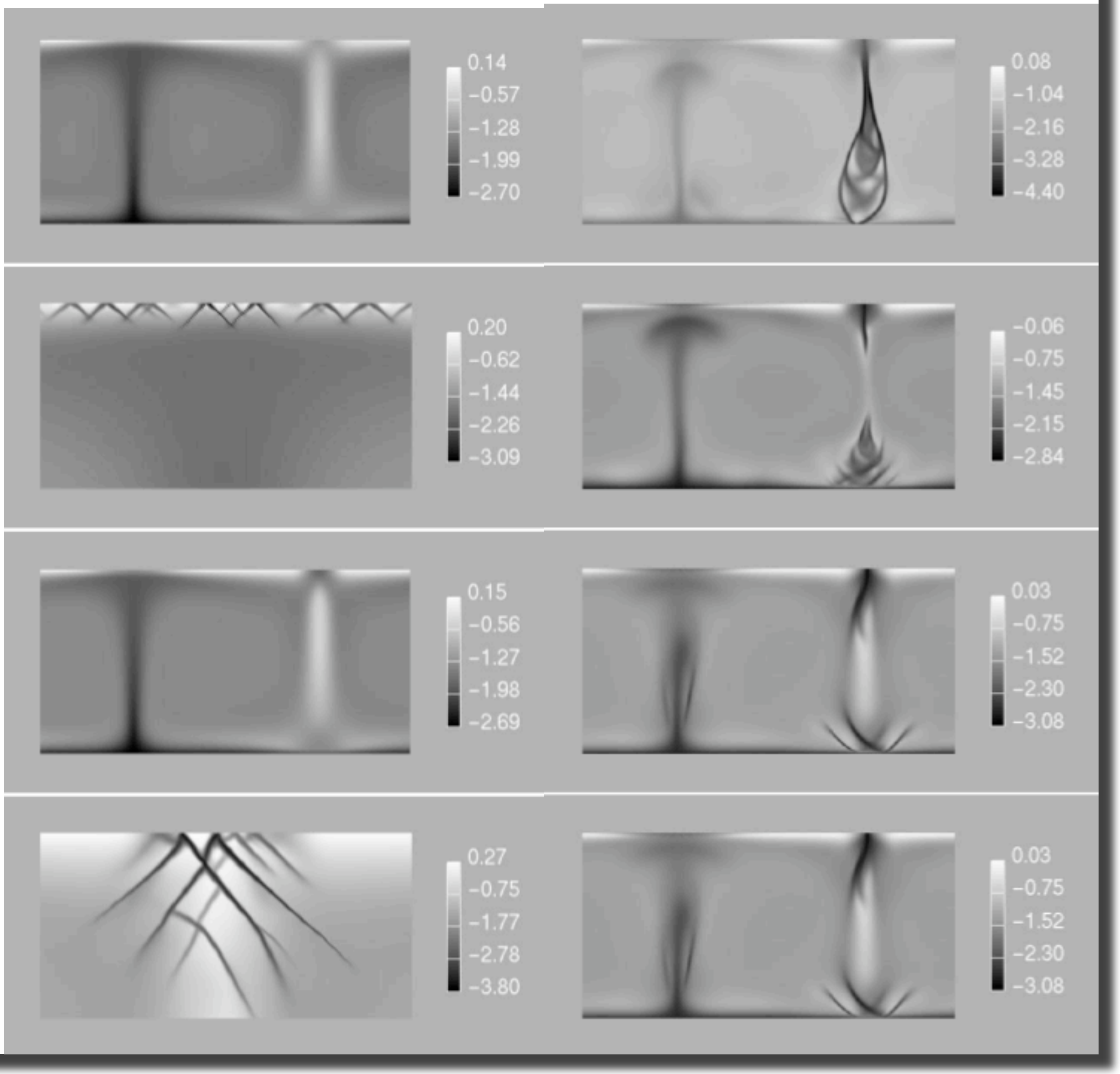
Viscosity(T) + damage

$$\eta = \frac{2}{1+d^m}\, e^{-\gamma T}$$

$$\frac{\partial d}{\partial t} + \boldsymbol{u}\cdot\boldsymbol{\nabla} d = a\sigma\dot{\varepsilon} - b e^{\gamma T} d$$

Produces ‘plates’ but
still double-sided



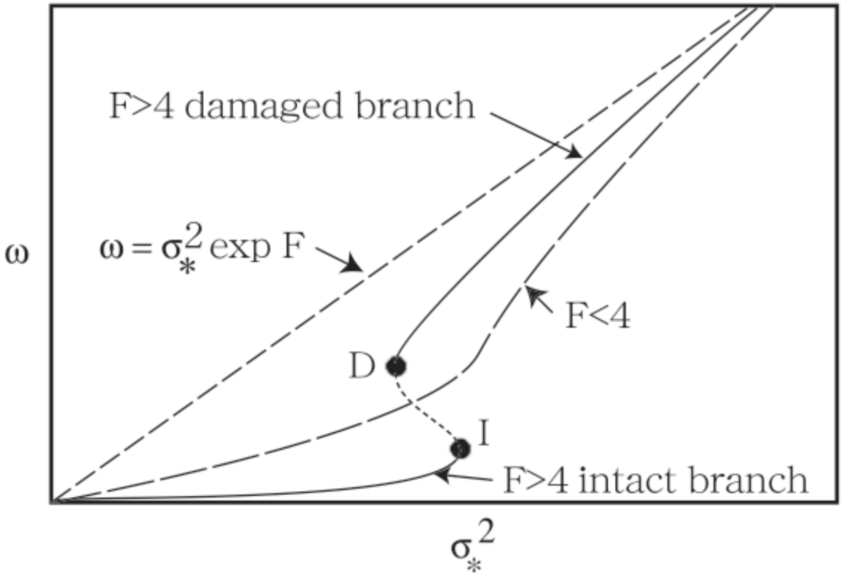


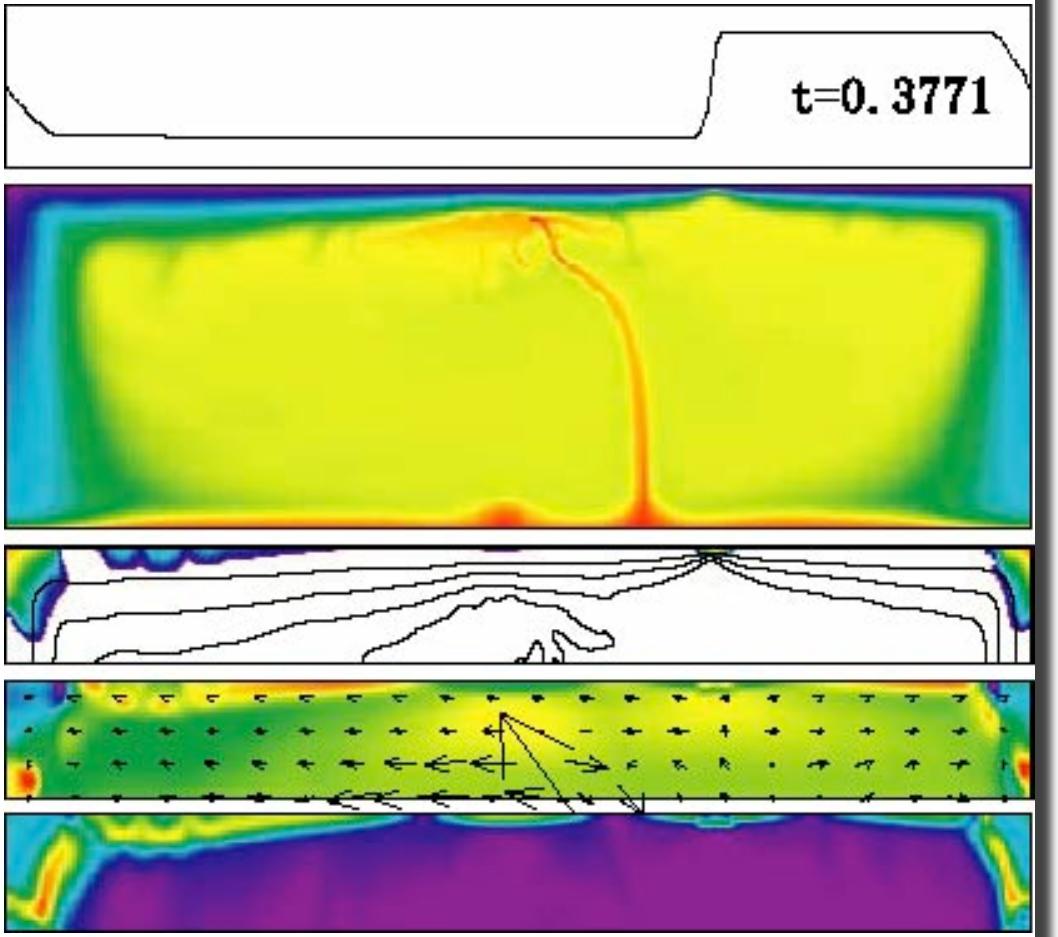
Ogawa (2003 JGR): Damage equation gives hysteresis (weak & strong brances)

$$\eta = \exp[E(T_{\mathrm{ref}} - T) + Vz - F\omega/(1+\omega)]$$

$$\frac{\partial \omega}{\partial t} + \mathbf{u} \cdot \nabla\omega = \Gamma\sigma_{ij}\, \dot{e}_{ij} - \lambda\omega$$

$$\lambda = \lambda_0 \exp(ET)$$

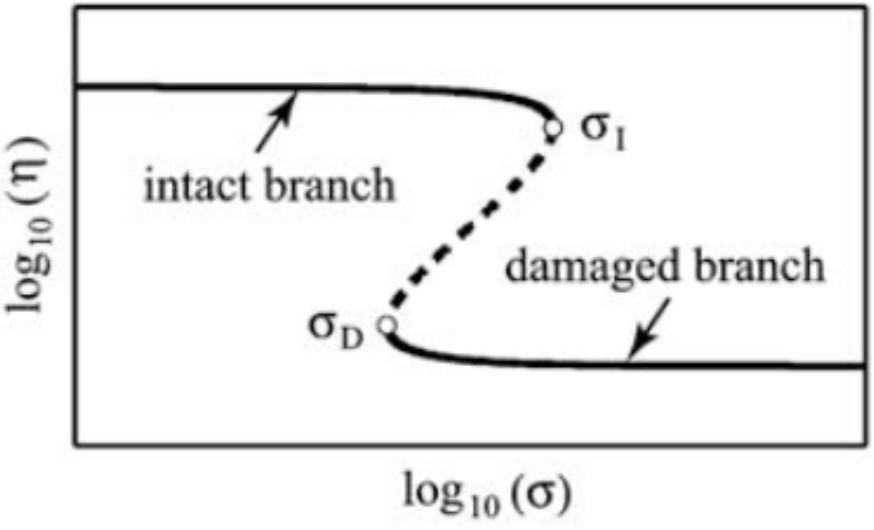


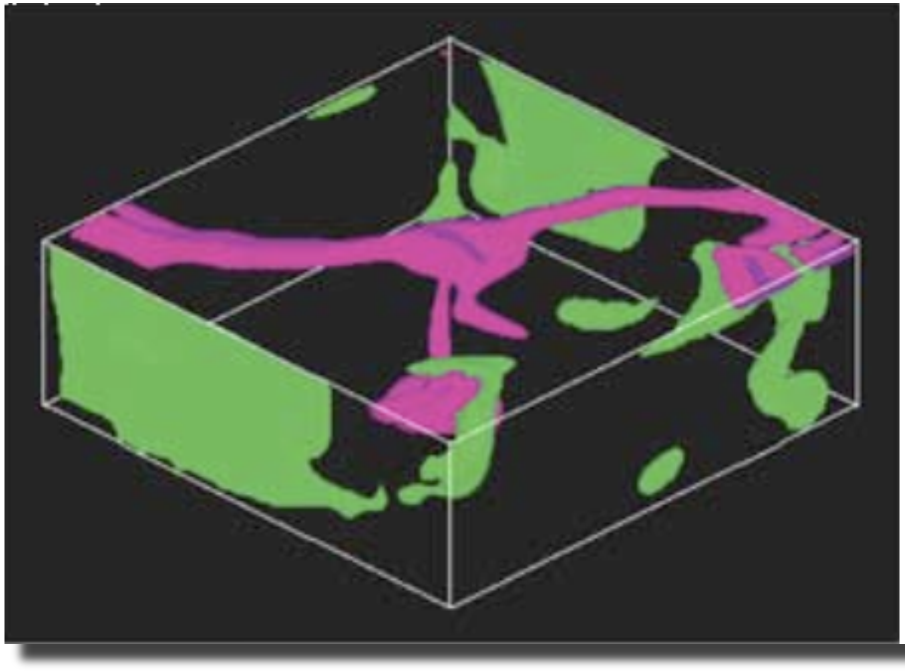


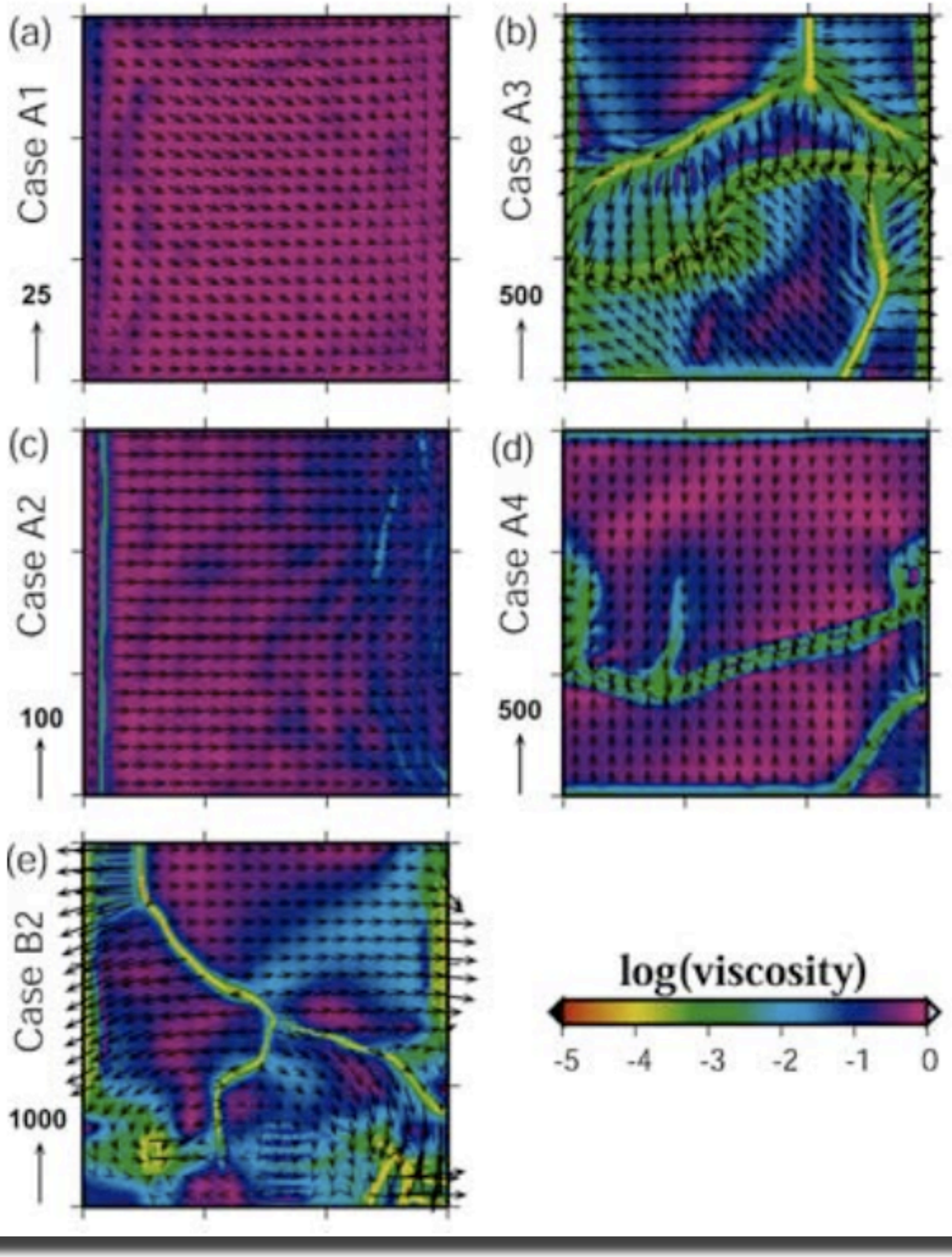
Plumes are needed to ‘break’ the lithosphere

Yoshida & Ogawa (GRL 2004)

Extend this to 3-D



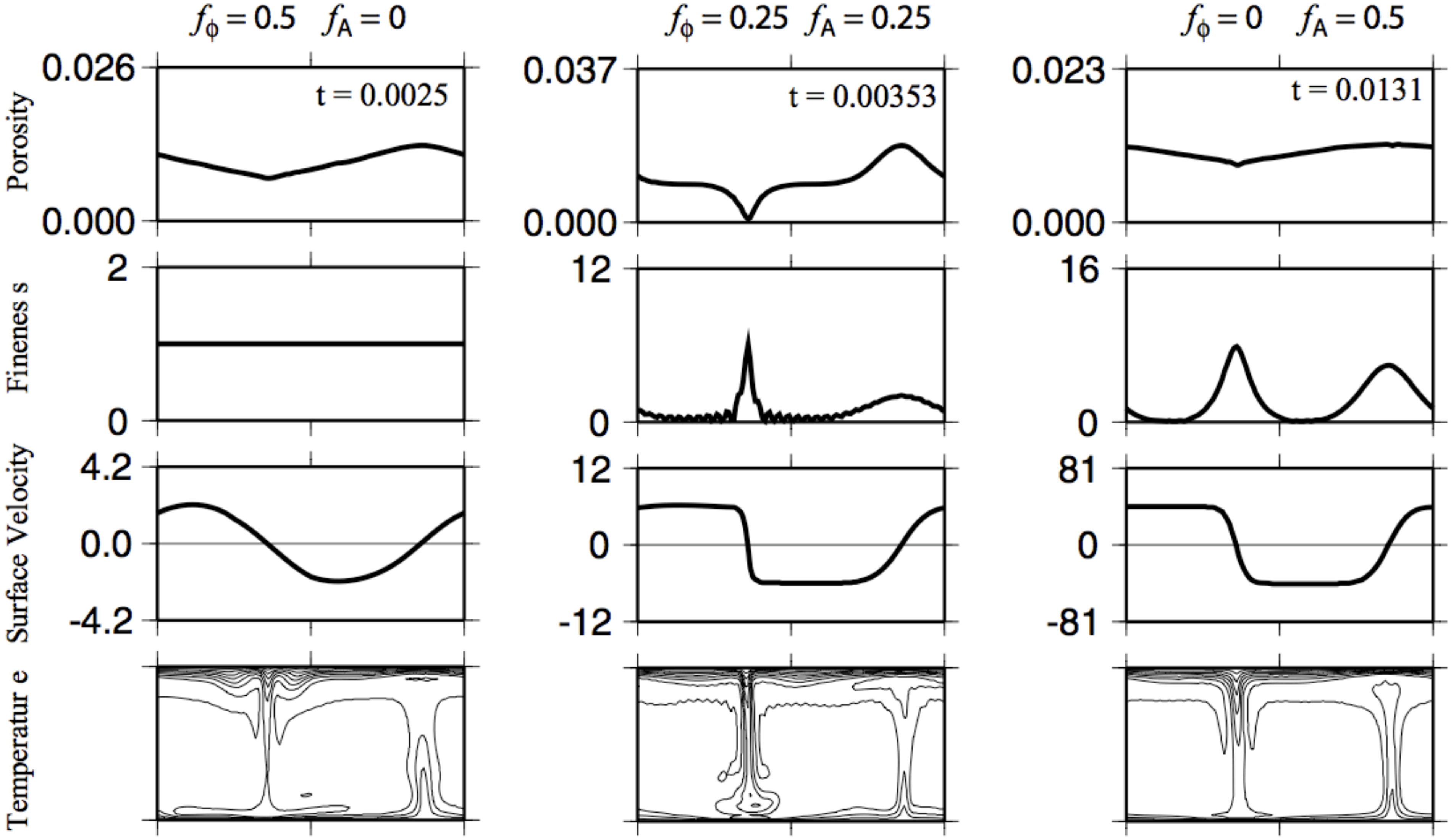




Landuyt et al. (GJI 2008)

2-phase damage theory + ‘fineness’ (grain size reduction)

‘Thin sheet’ high viscosity lithosphere



Summary so far

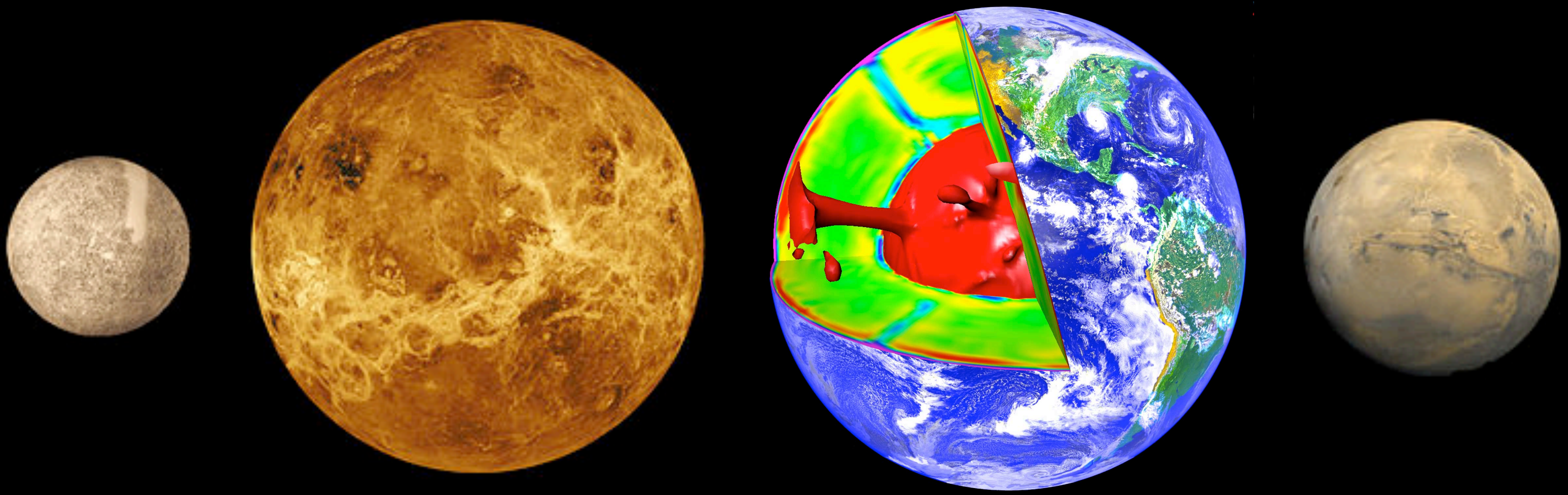
- ‘Effective’ strength of the lithosphere ~100 MPa (or similarly, effective friction coefficient <0.1). Lower than ‘laboratory’ values.
- Weak ‘asthenosphere’ increases ‘plateness’
- Both plastic yielding and ‘damage’ can cause plate boundaries
- Successes
  - Linear 'subduction'
  - Linear passive spreading centers+rifts
  - Toroidal:Poloidal ratio realistic (sometimes)
- Failures
  - Subduction double-sided
  - No pure strike-slip margins
  - Yield stress too low

One-sided Subduction in Self-Consistent Models of Global Mantle Convection: The Importance of a Free Surface and Weak Crustal Layer

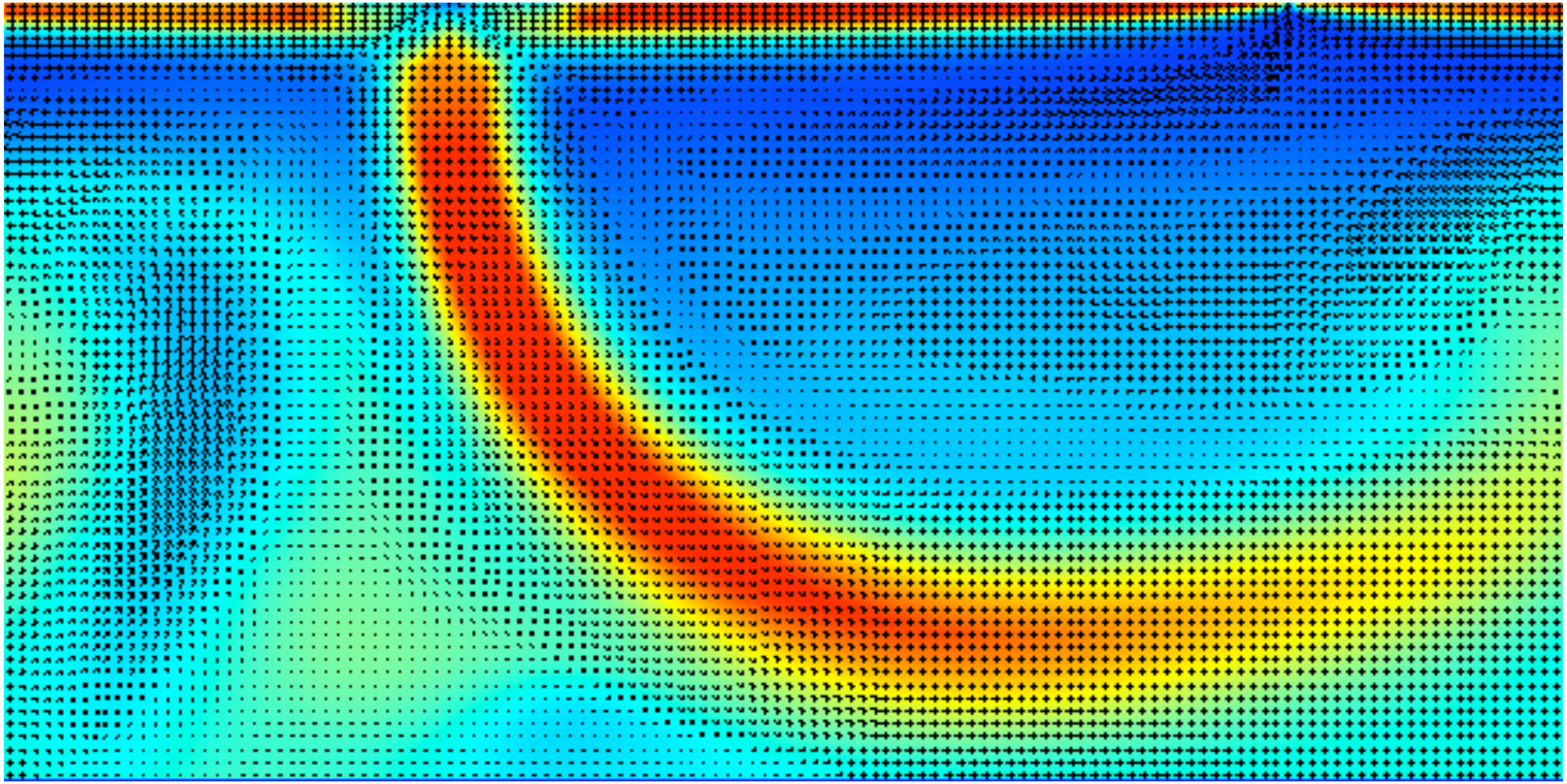
Paul Tackley,

F. Meilick, F. Crameri, T. V. Gerya, B. Kaus, T. Keller

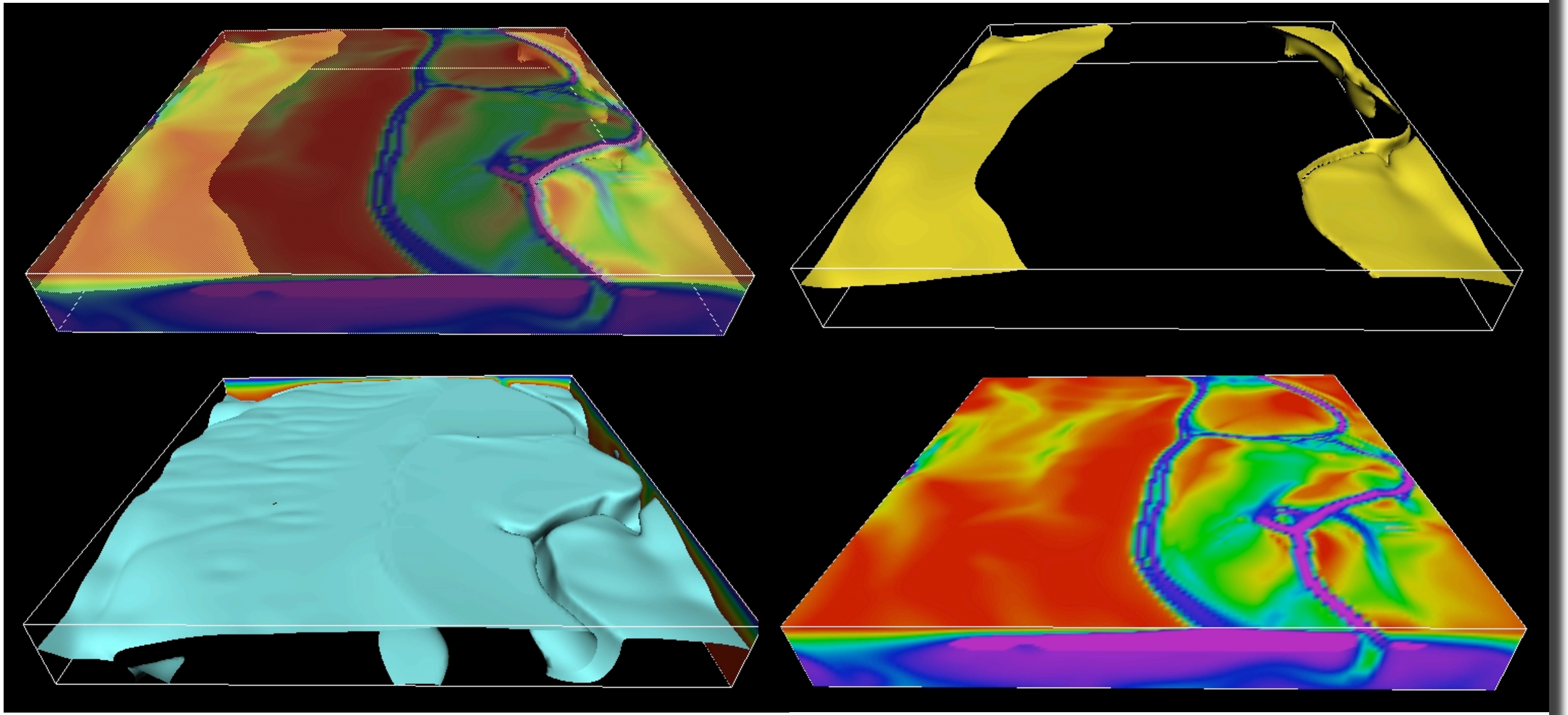
Institut für Geophysik, ETH Zürich



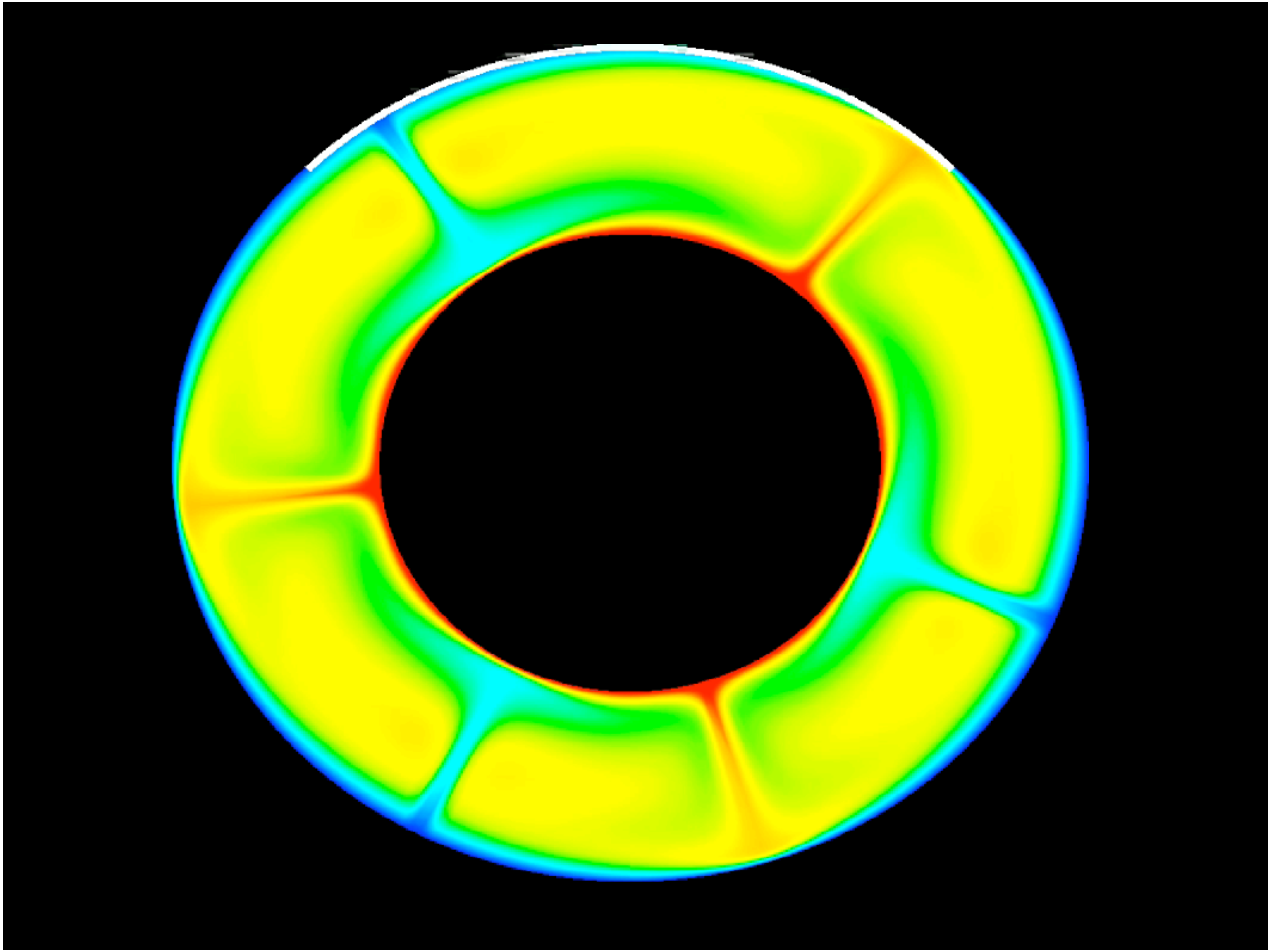
The problem: 2-sided subduction!



One solution: asymmetry due to continents



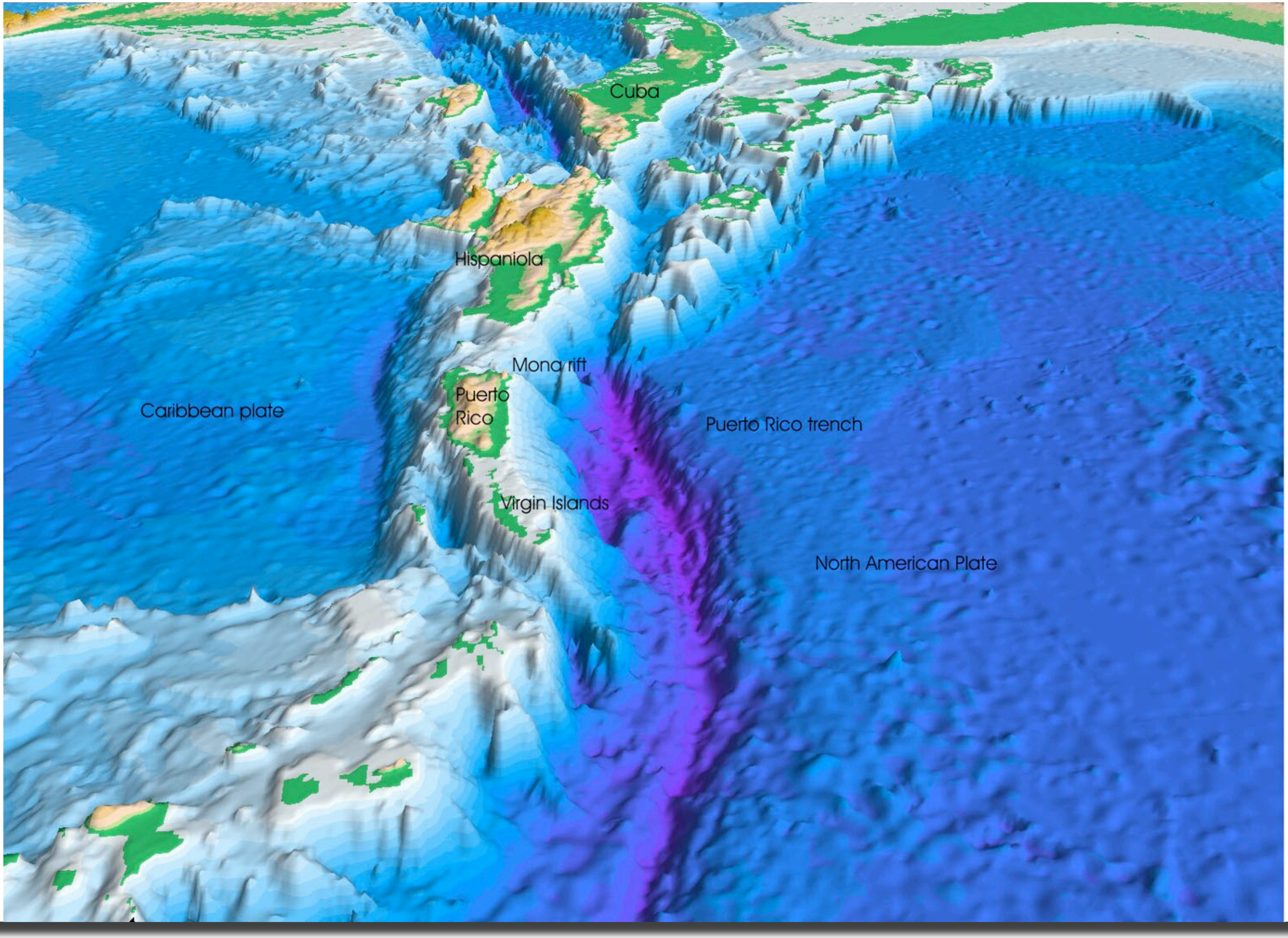
But on Earth, 1-sided ocean-ocean subduction also exists



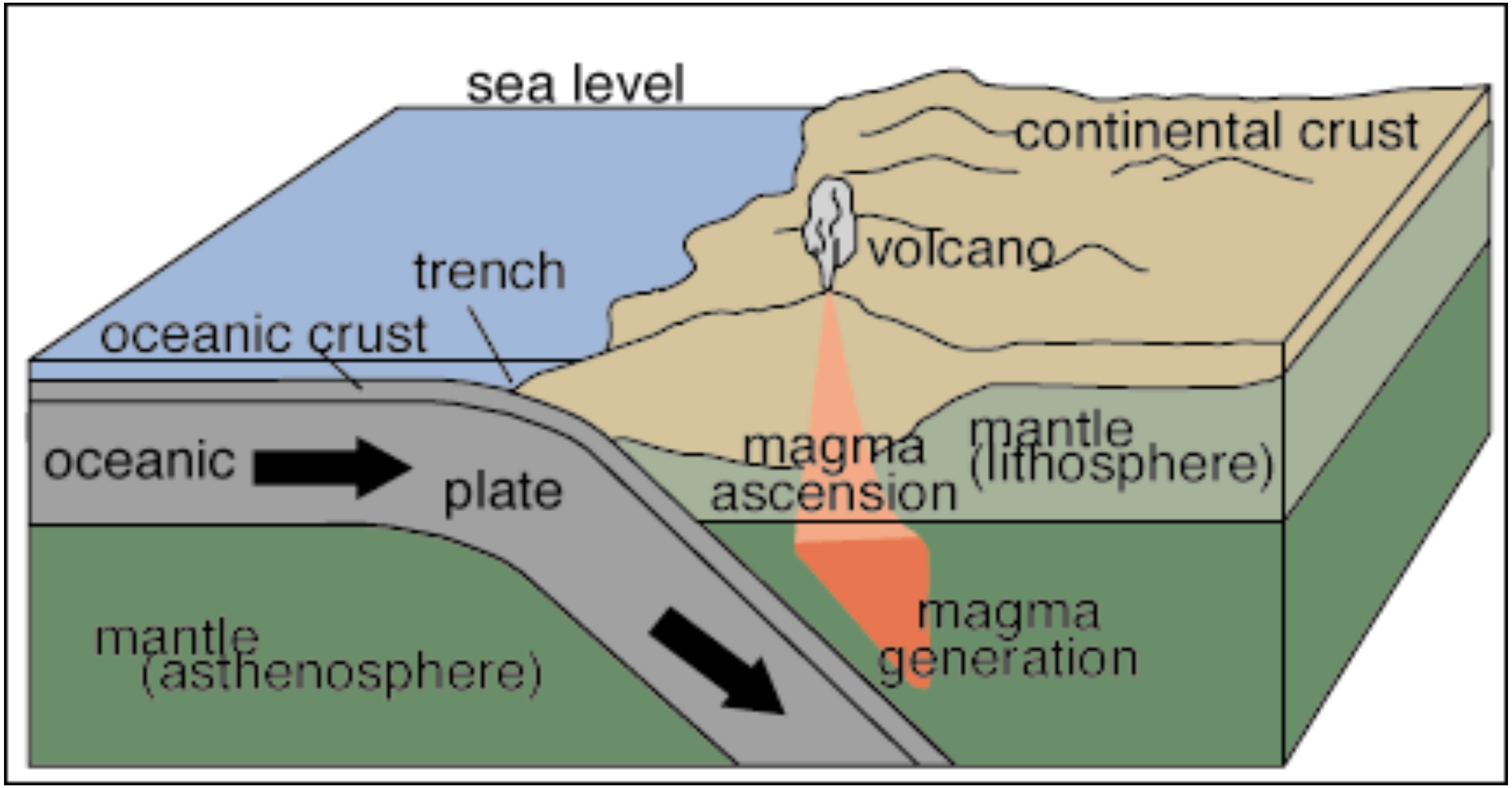
Mantle convection codes assume a free-slip boundary condition: surface is FLAT

- Zero shear stress but finite normal stress, proportional to what the topography would be if allowed.
- But this may create unnatural geometries at subduction zones….

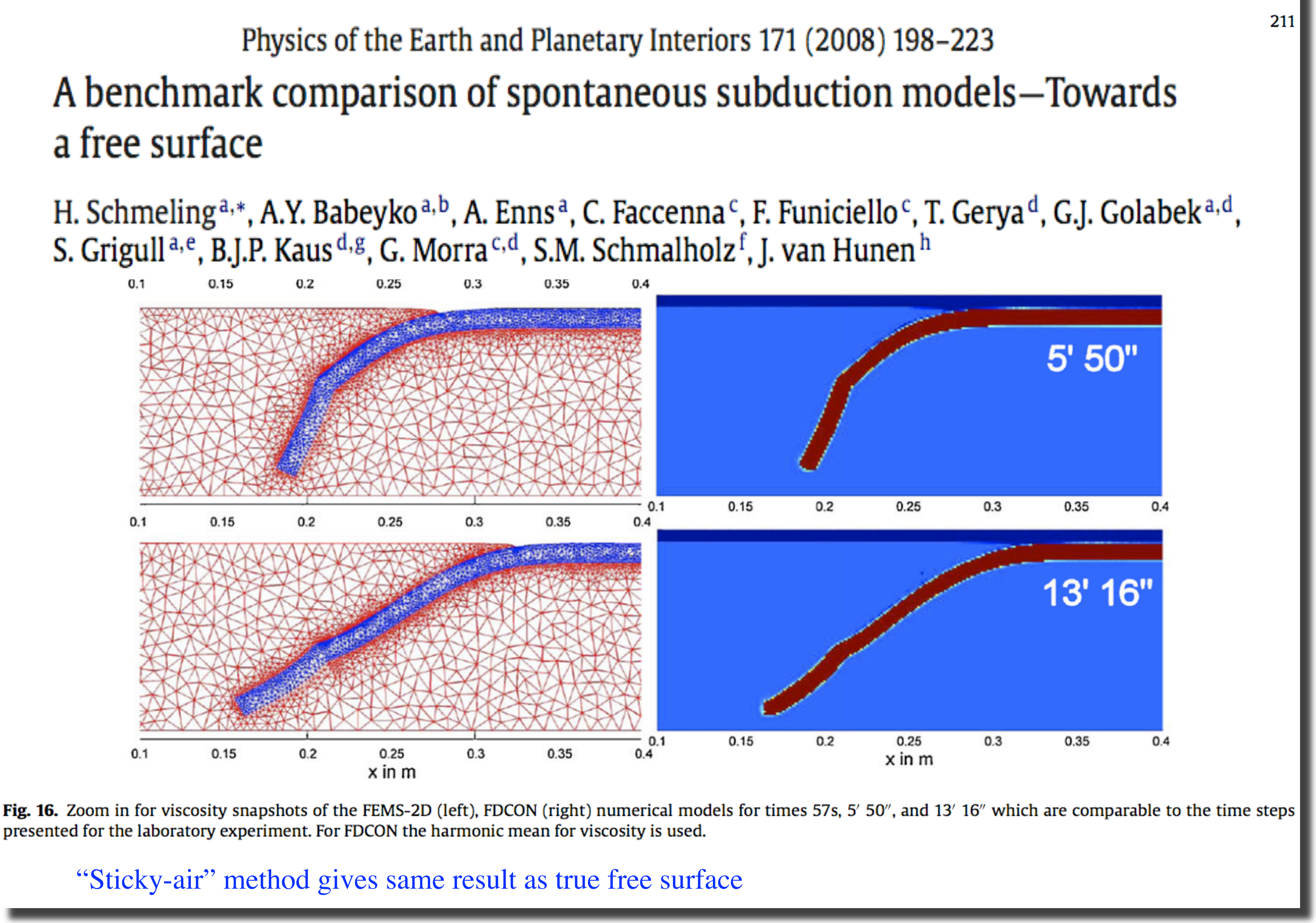
Real subduction zone: NOT FLAT



Trench due to bending



Numerical models with a free surface: also get a trench



Physics of the Earth and Planetary Interiors 171 (2008) 198–223

A benchmark comparison of spontaneous subduction models—Towards a free surface

H. Schmeling[a,*], A.Y. Babeyko[a,b], A. Enns[a], C. Faccenna[c], F. Funiciello[c], T. Gerya[d], G.J. Golabek[a,d], S. Grigull[a,e], B.J.P. Kaus[d,g], G. Morra[c,d], S.M. Schmalholz[f], J. van Hunen[h]

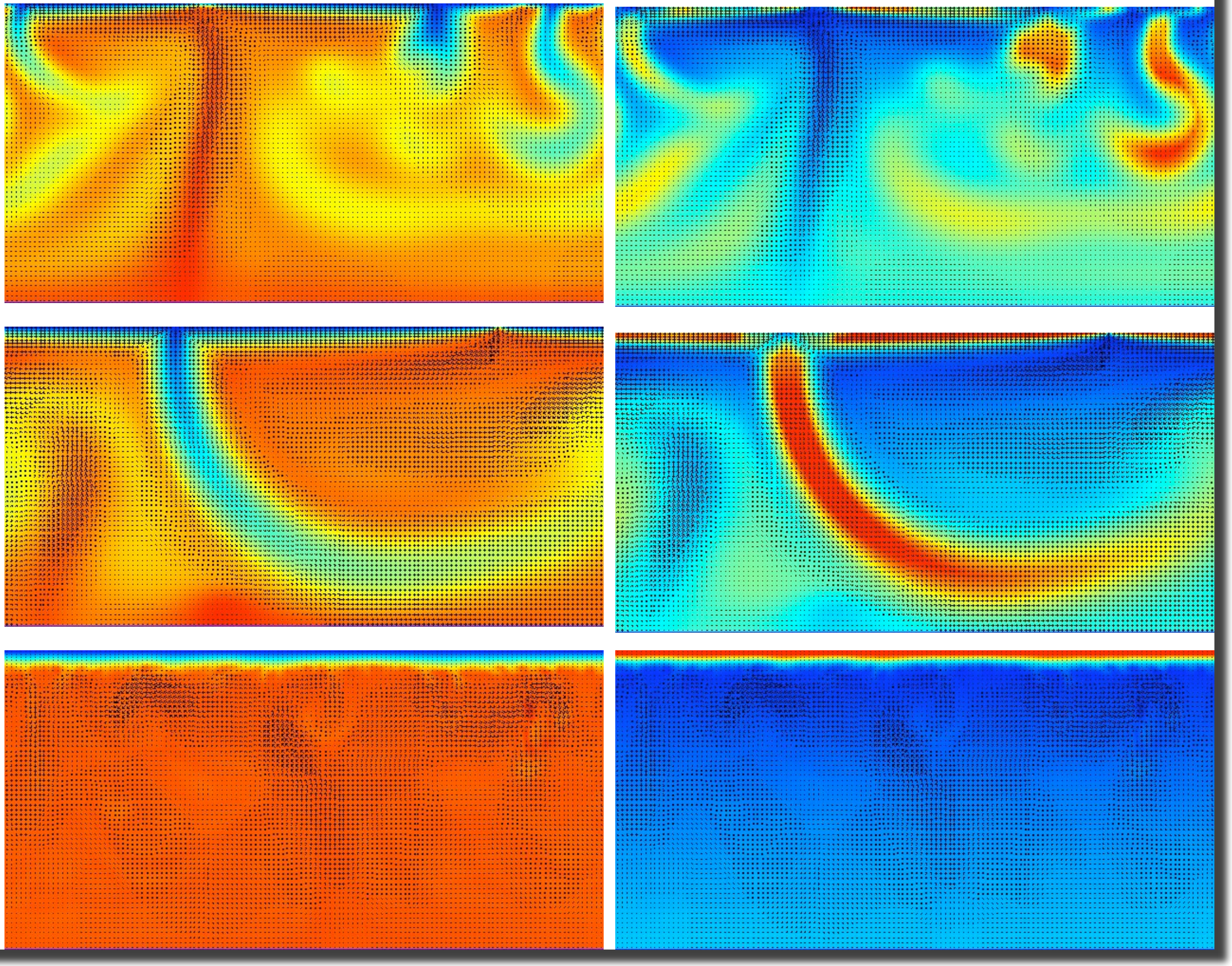
Fig. 16. Zoom in for viscosity snapshots of the FEMS-2D (left), FDCON (right) numerical models for times 57s, 5′ 50″, and 13′ 16″ which are comparable to the time steps presented for the laboratory experiment. For FDCON the harmonic mean for viscosity is used.

“Sticky-air” method gives same result as true free surface

What effect does a free surface have on free convection with “self-consistent” plate tectonics?

- Run simple, Boussinesq convection models with strongly T-dependent viscosity
  - E_act for wet olivine, plus variable V_act
- …and depth-dependent (Byerlee’s law-type) plastic yield stress,
  - Drucker-Prager yield criterion (2nd invariant)
  - Specify friction coefficient
- Truncate to 9 orders of magnitude variation

Free-slip (flat) upper boundary



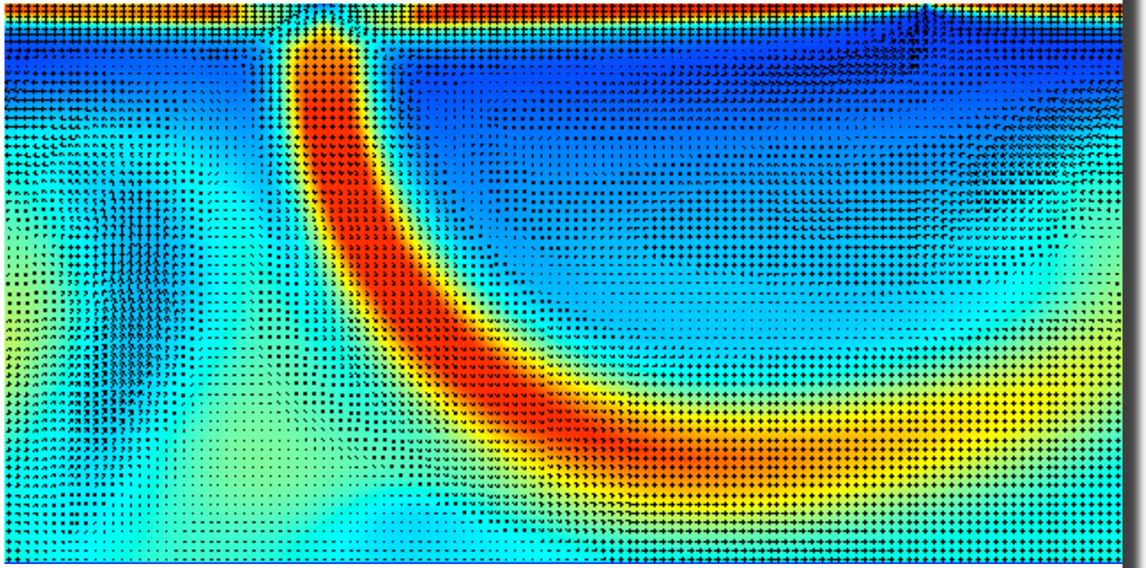
Friction=
0.03

0.09

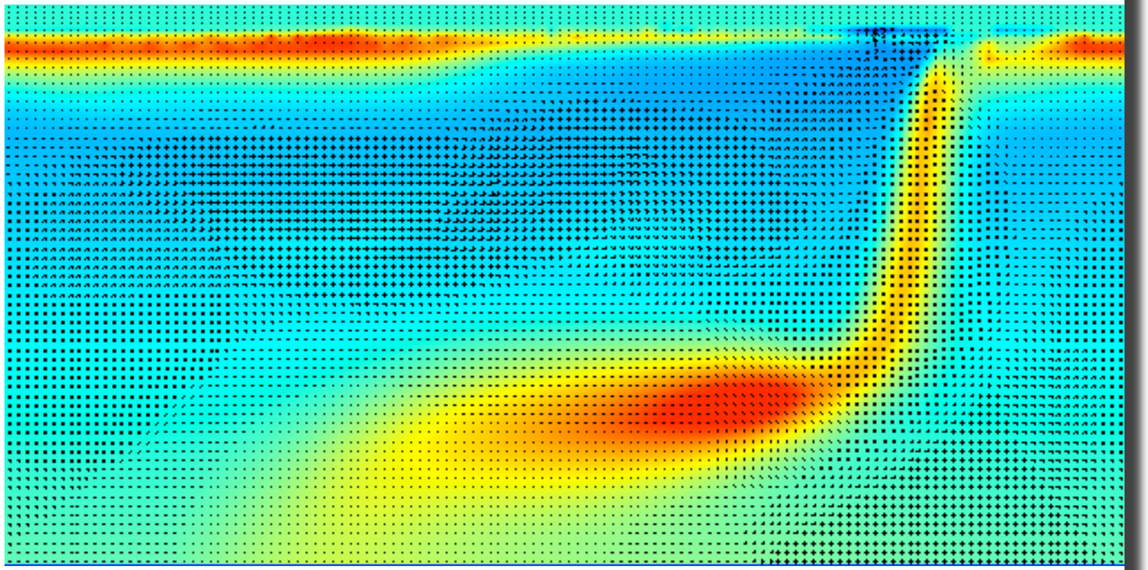
0.15

Free-slip to free comparison

Free-slip



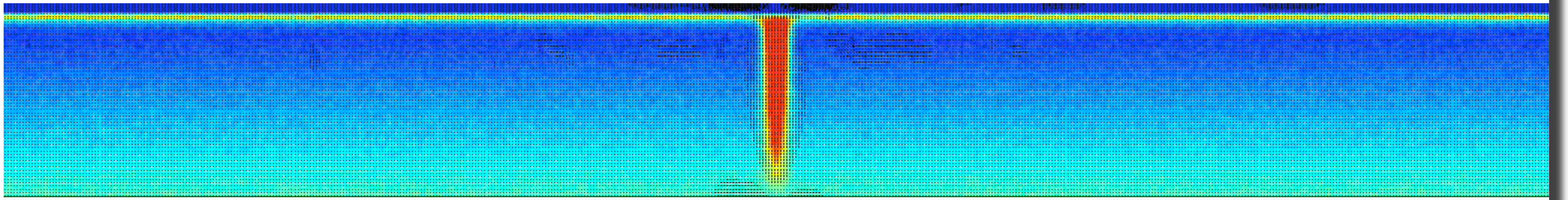
Free surface



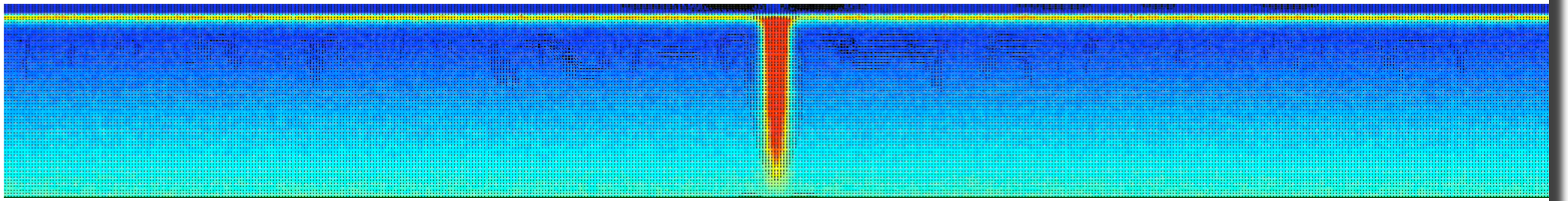
Single sided subduction!

Movies

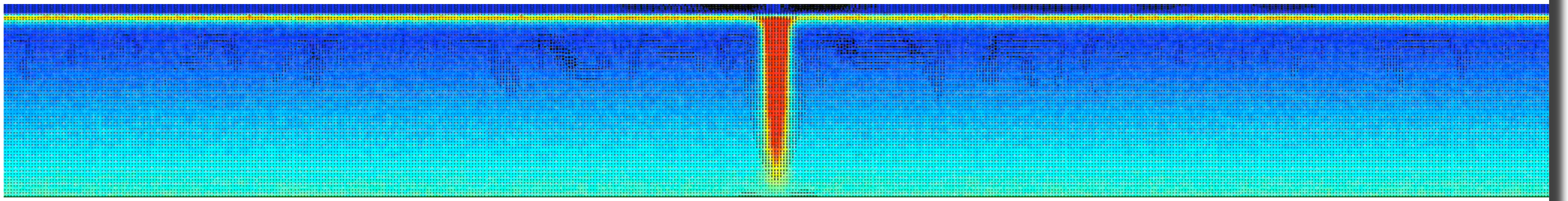
Friction coeff = 0.05



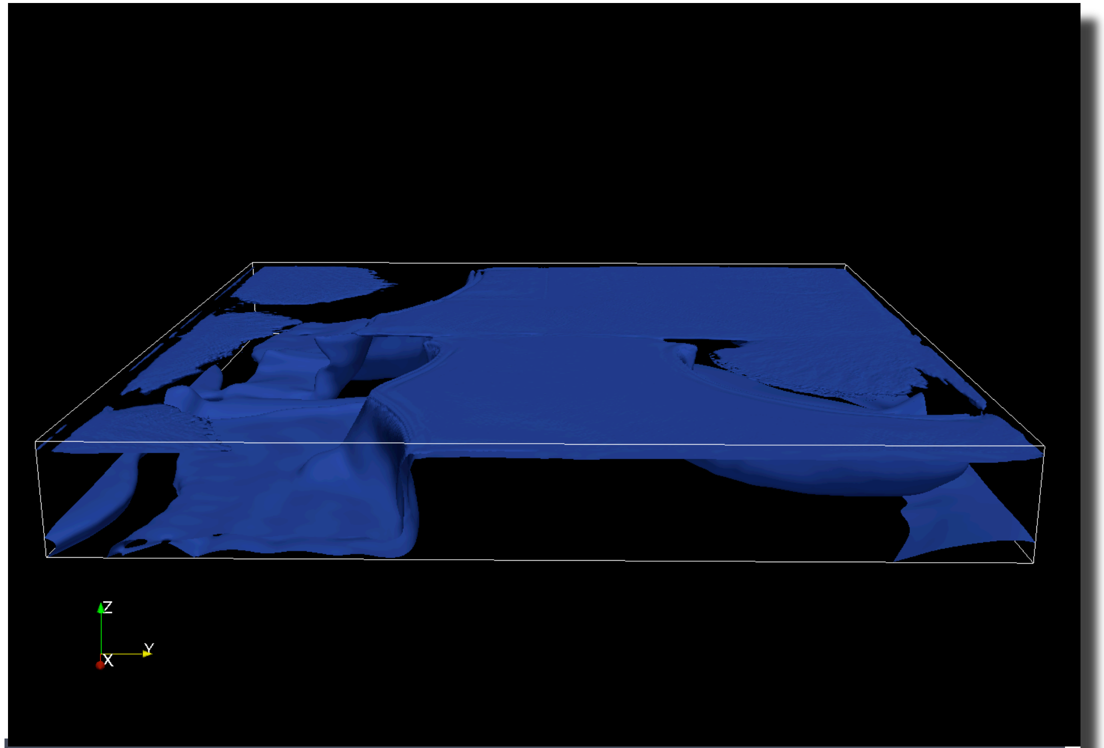
Friction coeff = 0.1

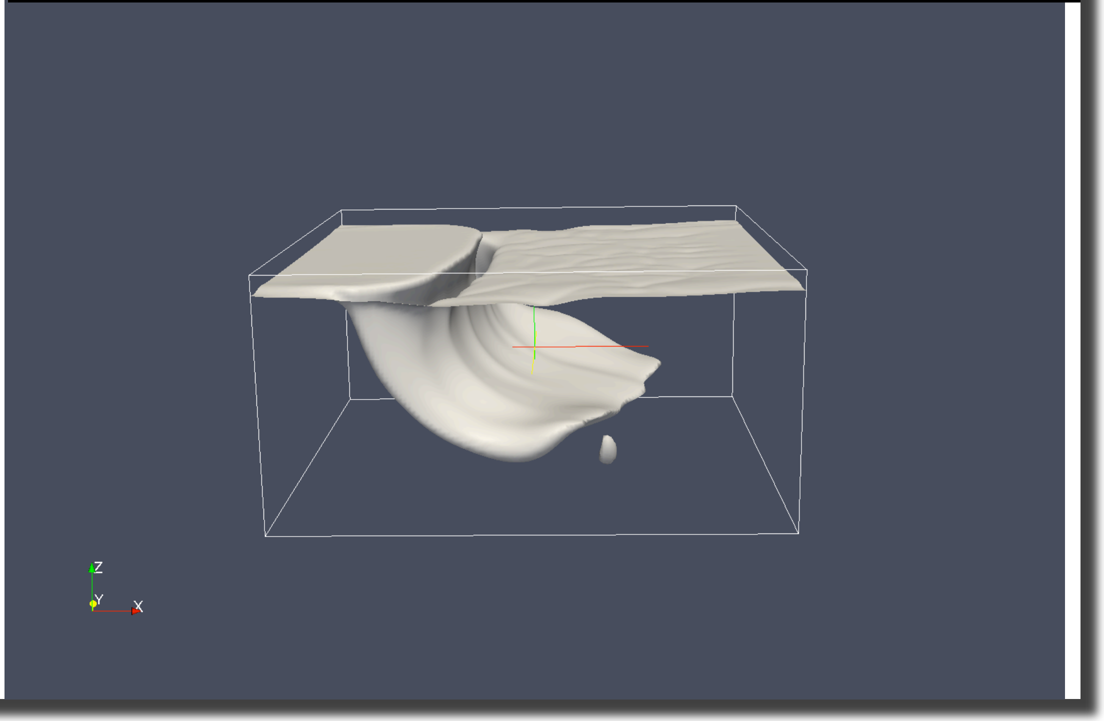


Friction coeff = 0.11



3-D cases

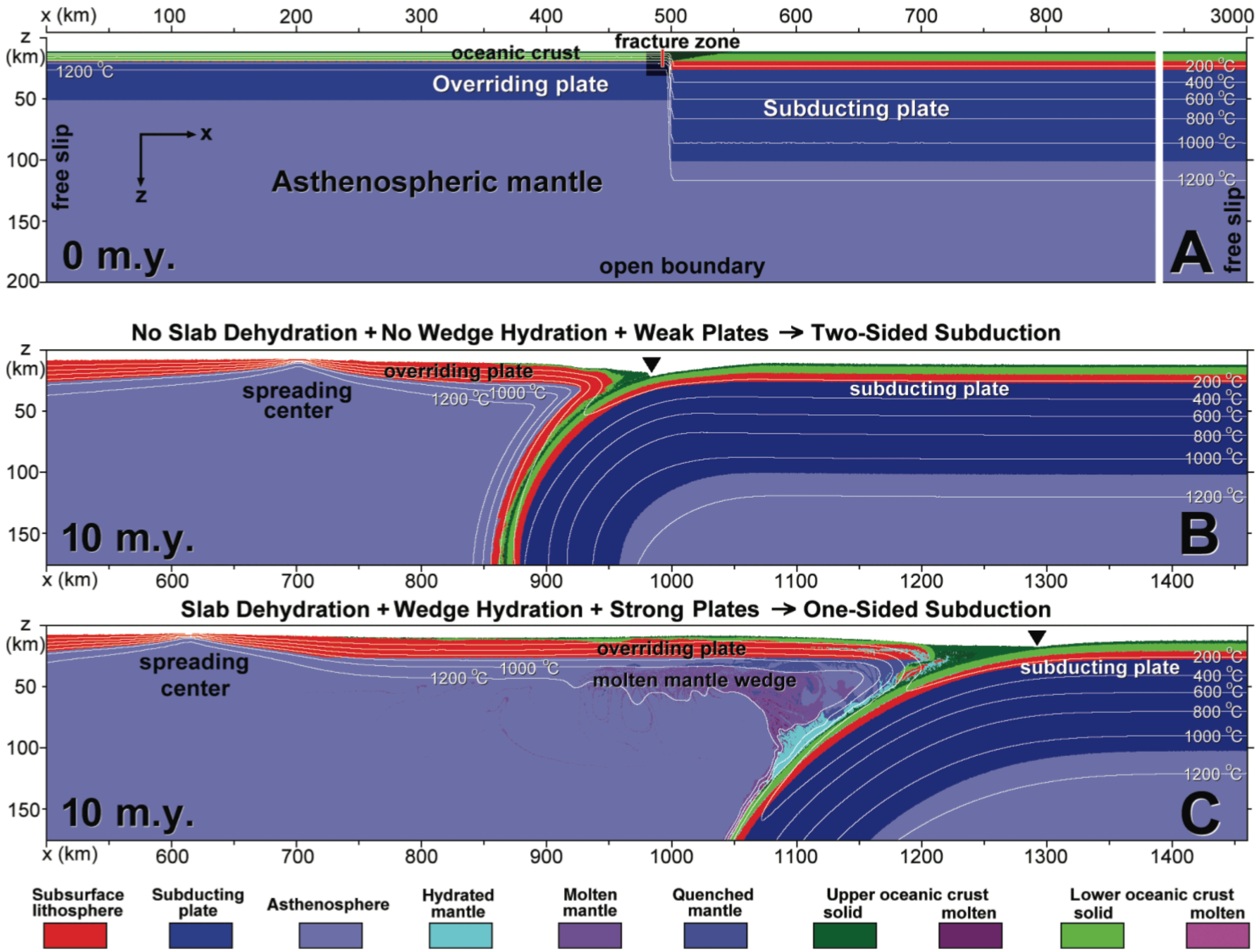




Findings

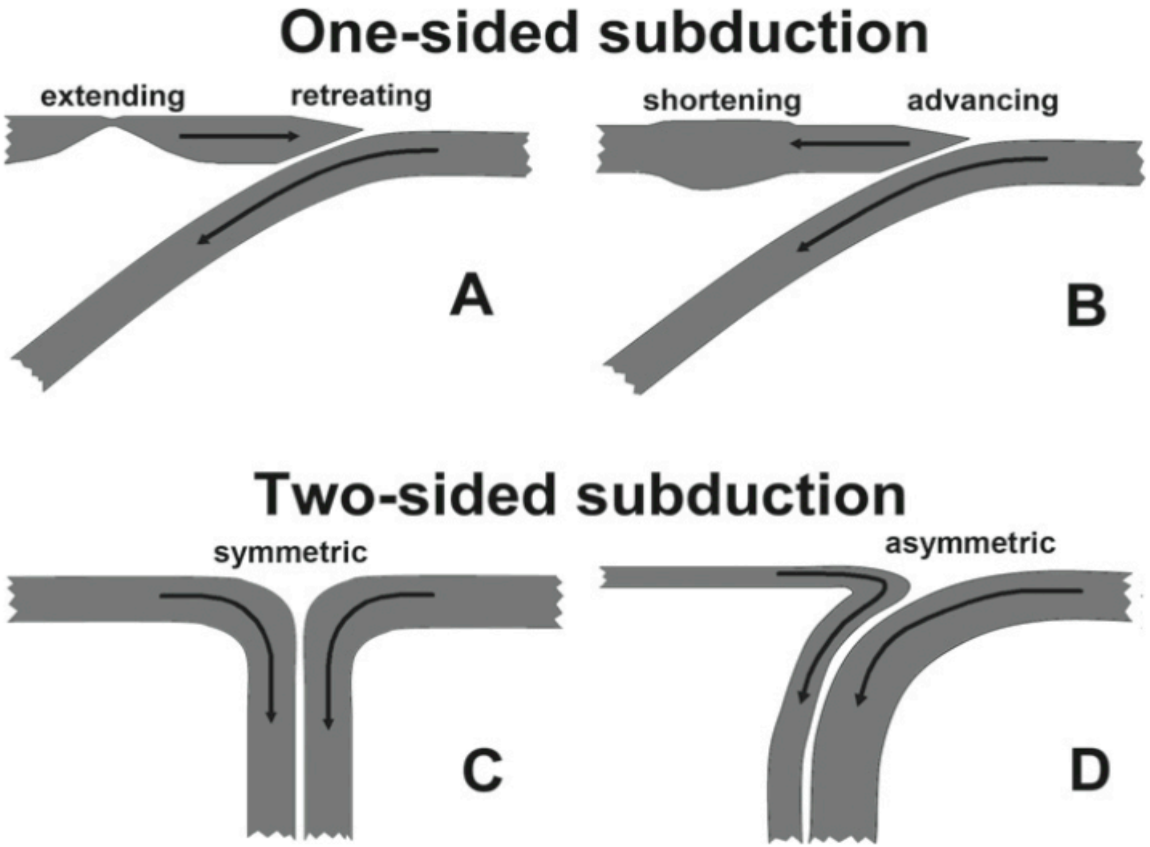
- Free surface leads to (thermally) single-sided subduction over a wide parameter range
- But so far, eventually a rigid lid is obtained, even for parameters that lead to stable “plate tectonics” with a free-slip surface
- Research is ongoing…

Weak hydrated sediments could be important: Gerya et al., Geology 2008



For free convection, we have found that they can have an importance effect but a free surface is needed as well

Modes

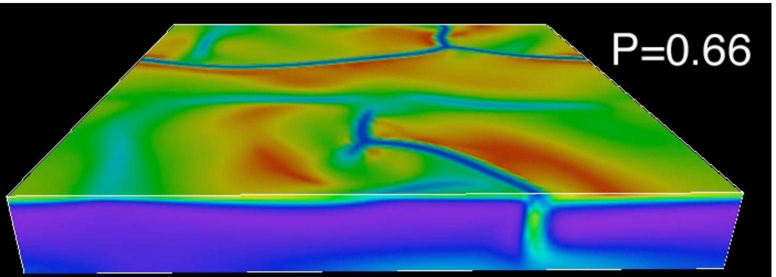

One-sided subduction
extending
retreating
A
shortening
advancing
B
Two-sided subduction
symmetric
C
asymmetric
D

Conclusions

- Free surface leads to (thermally) single-sided subduction over a wide parameter range
- But so far, eventually a rigid lid is obtained, even for parameters that lead to stable “plate tectonics” with a free-slip surface
- Weak sediments important, but don’t cause 1-sided subduction without a free surface
- Research is ongoing…

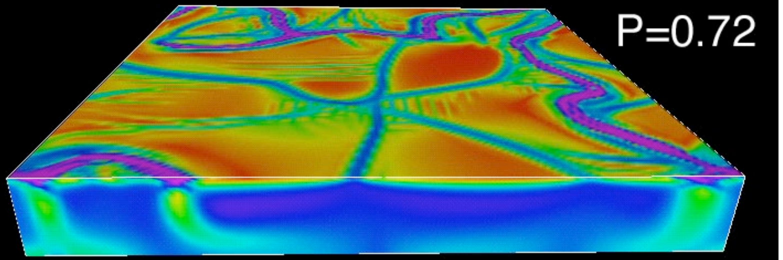
Summary

Constant Yield
Stress=8.4e3



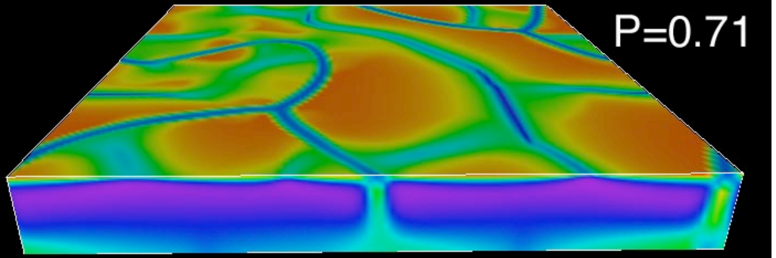
P=0.66

+strain-rate
weakening



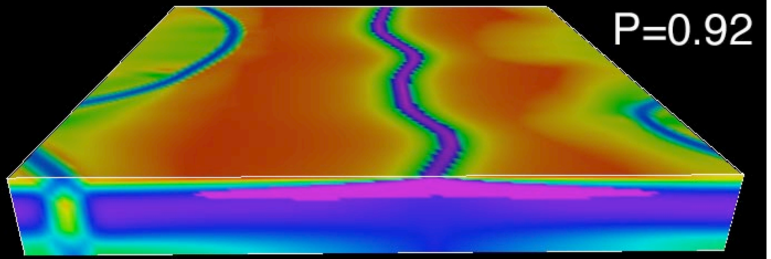
P=0.72

+visc(z) 10*



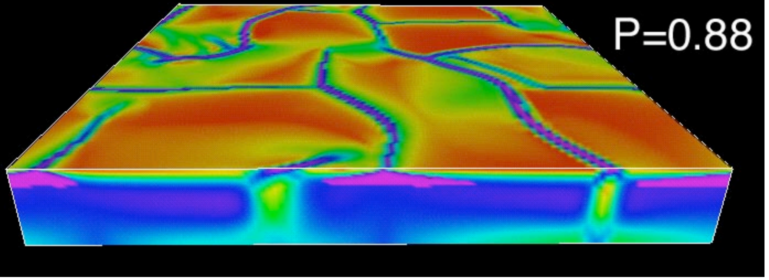
P=0.71

+visc(melt)
0.1*



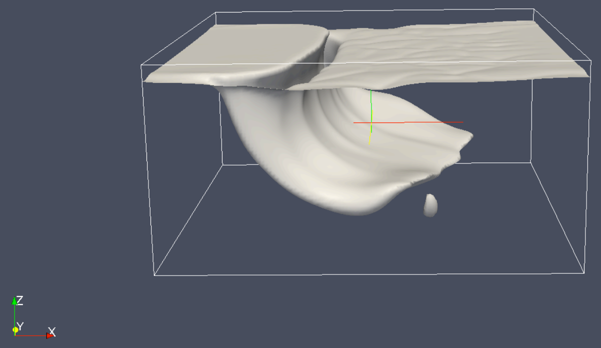
P=0.92

+SRW
+visc(melt)



P=0.88

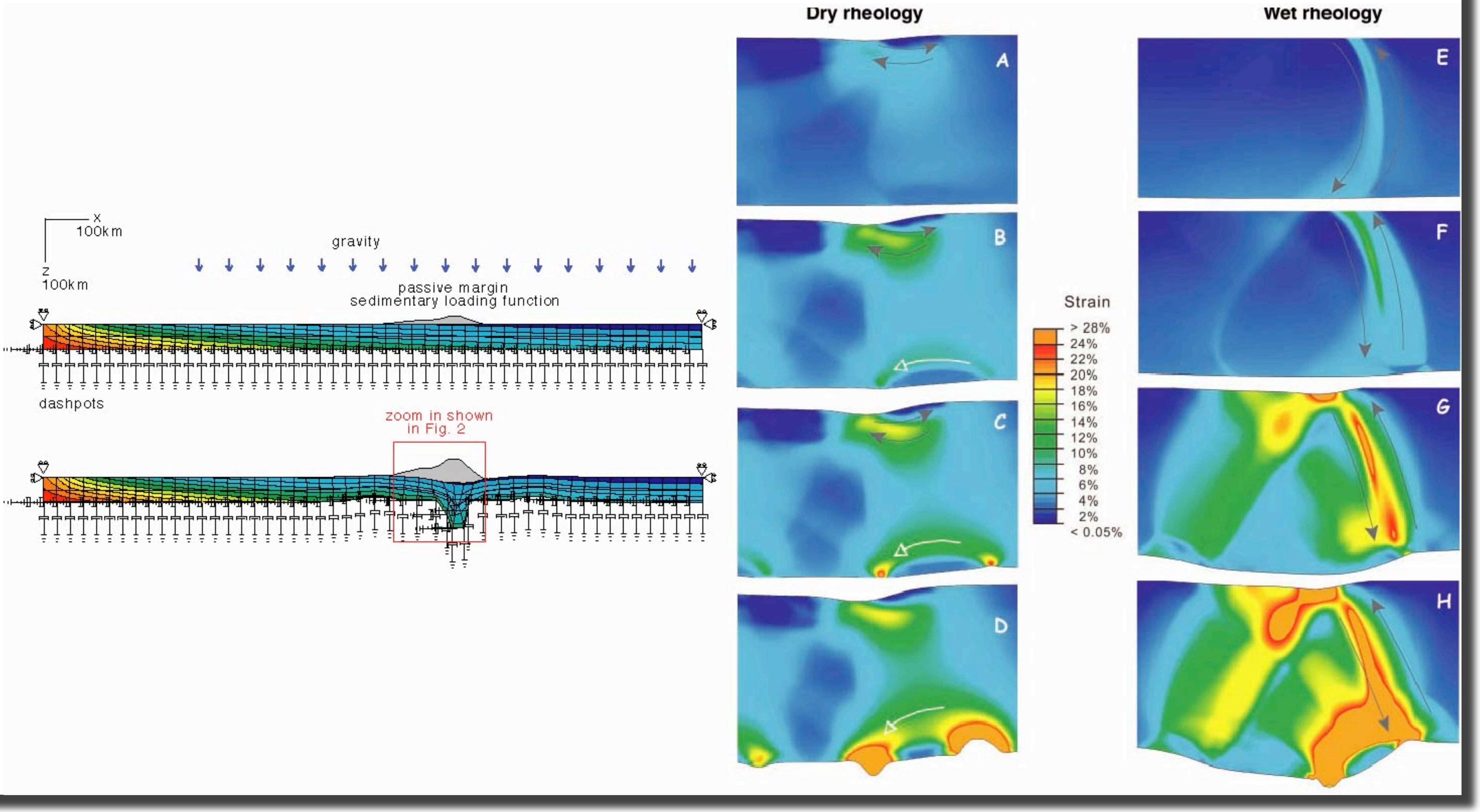
+free surface



Open questions

- Why are plate boundaries so weak?
- How is subduction initiated?
- What is a ‘realistic’ large-scale rheology?
  - How do small-scale processes influence the large scale?
- How important is history-dependence, anisotropy, …?

Regenauer-Lieb et al.: Full visco-elasto-plastic lithosphere models. Forms subduction zones



THE END!