The role of melt induced lithospheric weakening on the dynamics of continental rifting

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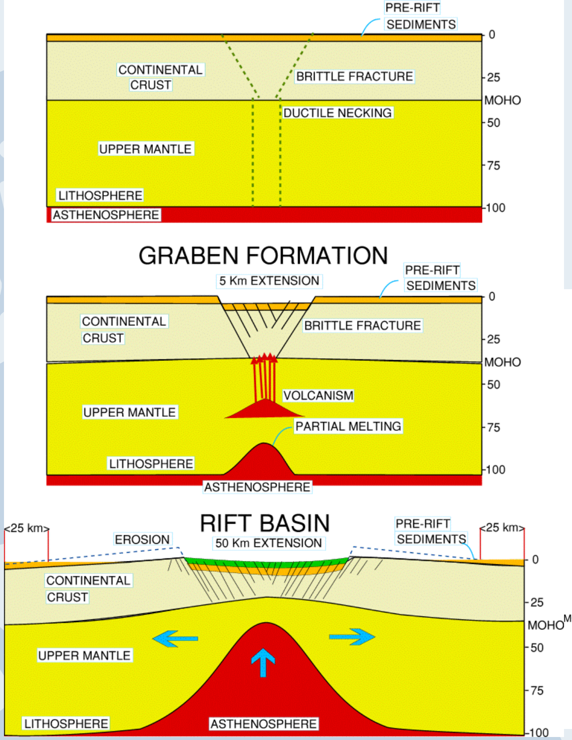
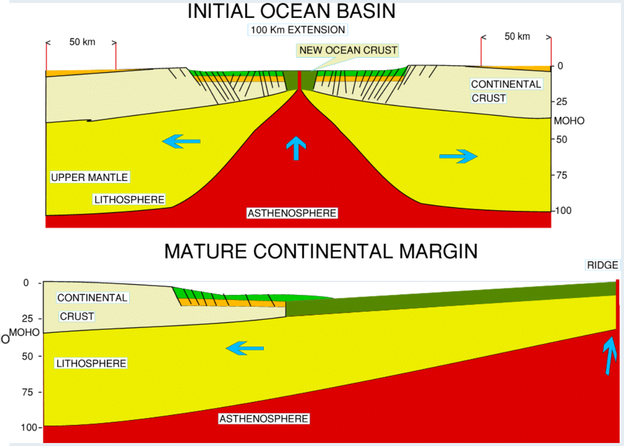
Aims

- Rifting scenario: Melt extraction and magma intrusion
  → enhanced heat transport → heating of shallow lithosphere
- 1D kinematic thinning model → T-increase
- 2D dynamic rifting model* → lithopheric weakening
- Effect on strength and feed back on magma production

* Schmeling, H., 2010: Dynamic models of continental rifting with melt generation. Tectonophysics,Volume 480, Issues 1-4, 5 January 2010, Pages 33-47

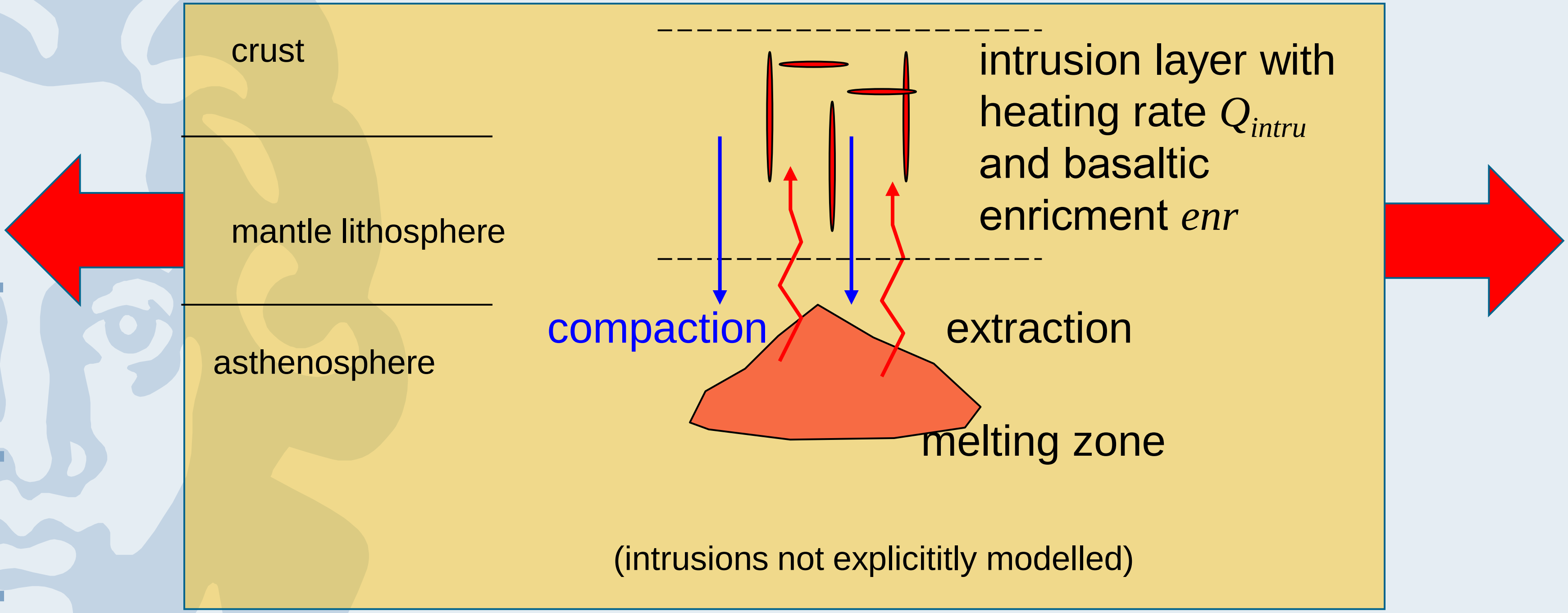




Source: J. Tarney

Model concept

Lithosphere under extension



crust

mantle lithosphere

asthenosphere

intrusion layer with heating rate Q_{intru} and basaltic enricment *enr*

compaction

extraction

melting zone

(intrusions not explicititly modelled)

$$Q_{intru} = \rho c_p \left(T_{source} - T_{ambient} + \left(1 - f_m(T_{ambient})\right) \cdot \frac{L}{c_p} \right) \cdot q_{intru}$$

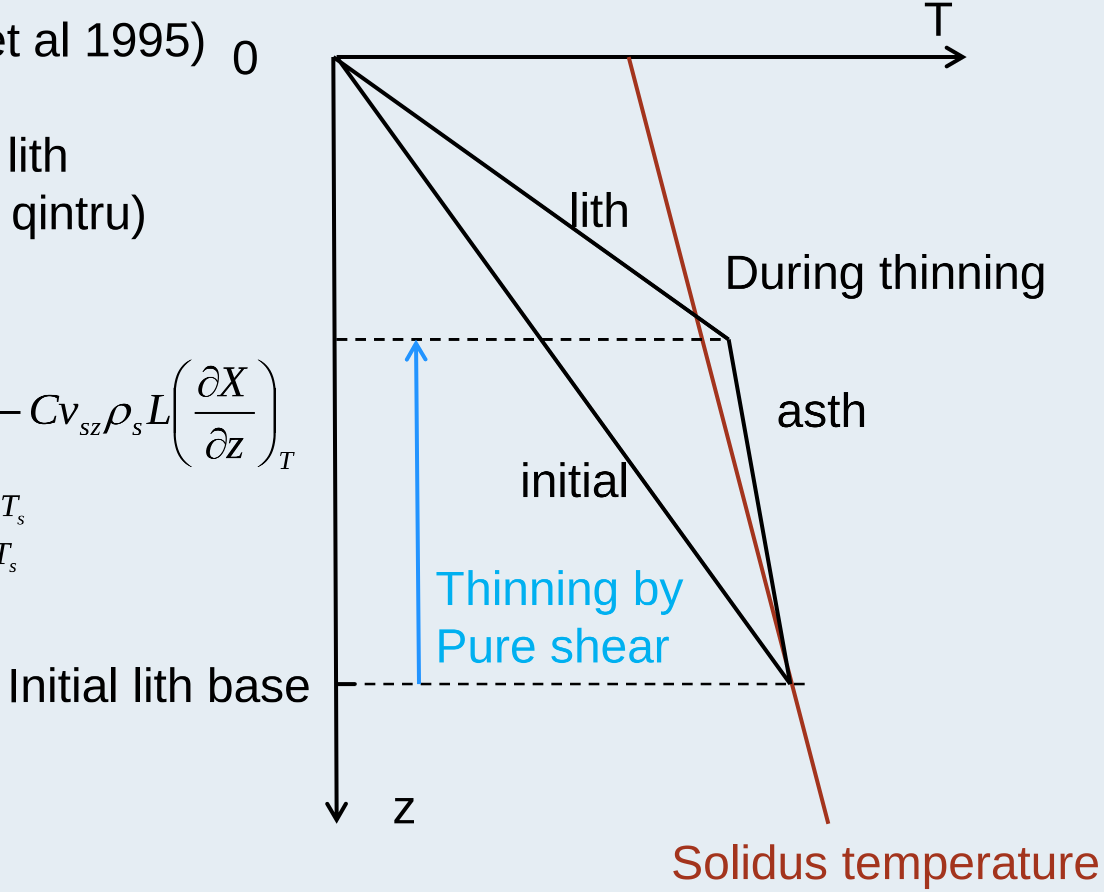
q_{intru} – volumetric intrusion rate

1D kinematic model

- Thinning by pure shear e_0 $v_{z0} = -\dot{e}_0 z_0$
- Fractional melting (Iwamori et al 1995)
- Melt extraction rate s in asth
- Magma intrusion rate q_{intru} in lith
- Mass conservation: $v_{sz}(e_0, s, qintru)$
- Energy equation

$$\rho_s \tilde{c}_p \left(\frac{\partial T}{\partial t} + \vec{v}_s \cdot \vec{\nabla} T \right) = k \nabla^2 T + Q_{intru} - C v_{sz} \rho_s L \left(\frac{\partial X}{\partial z} \right)_T$$

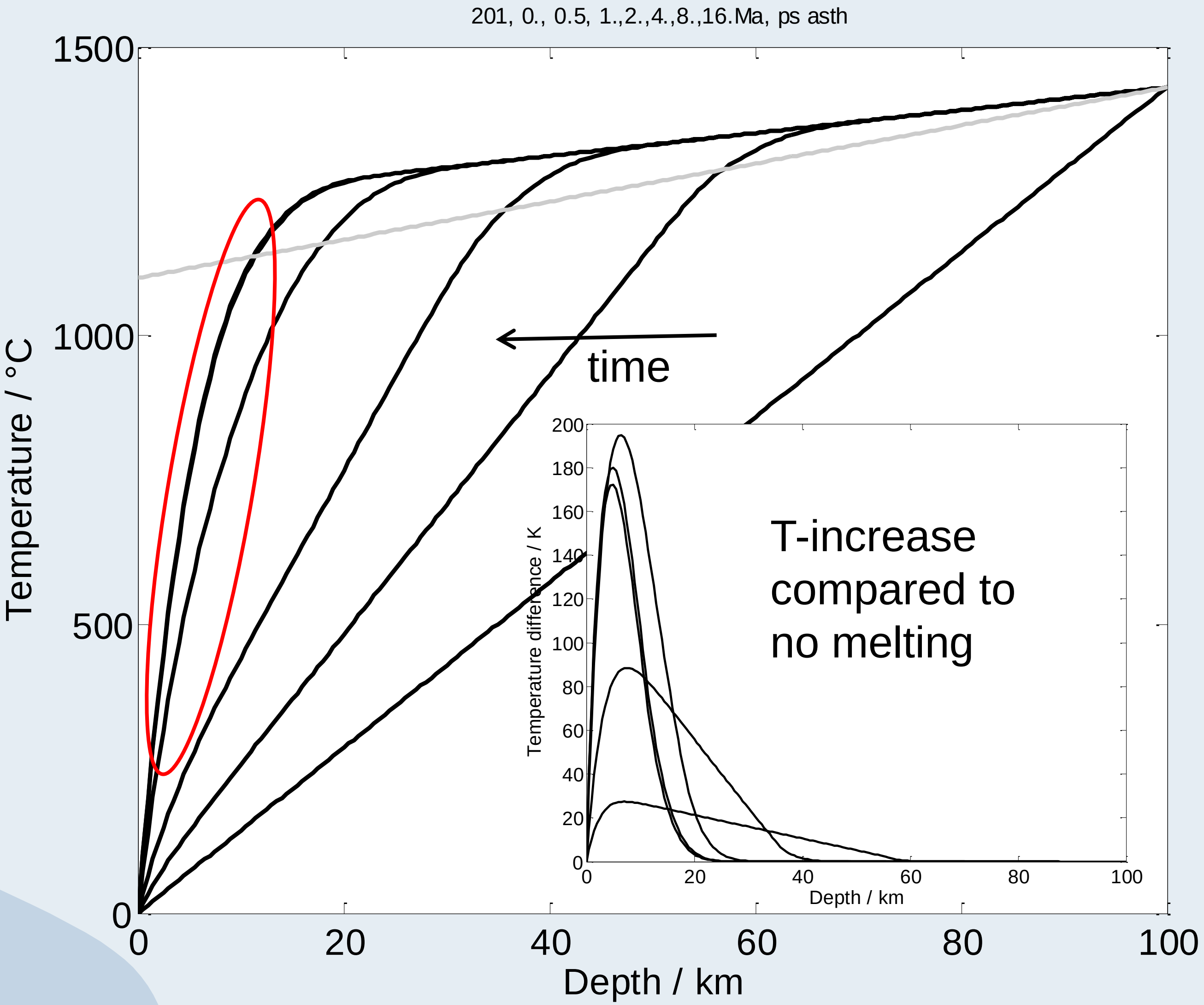
$$\tilde{c}_p = c_p + CL \left(\frac{\partial X}{\partial T} \right)_P , \qquad C = \begin{cases} 1 & for\ T > T_s \\ 0 & or\ T \le T_s \end{cases}$$



Result 1D model

→ No more straight lines due to strong intrusional heating

→ Several 100 K magmatic heating



T-increase compared to no melting

Result 1D model

Generalisation

- Thinning rate → Peclet number

$$Pe = \frac{t_{diffusion}}{t_{thinning}} = \frac{\dot{e}_0 h_0^{\,2}}{\kappa}$$

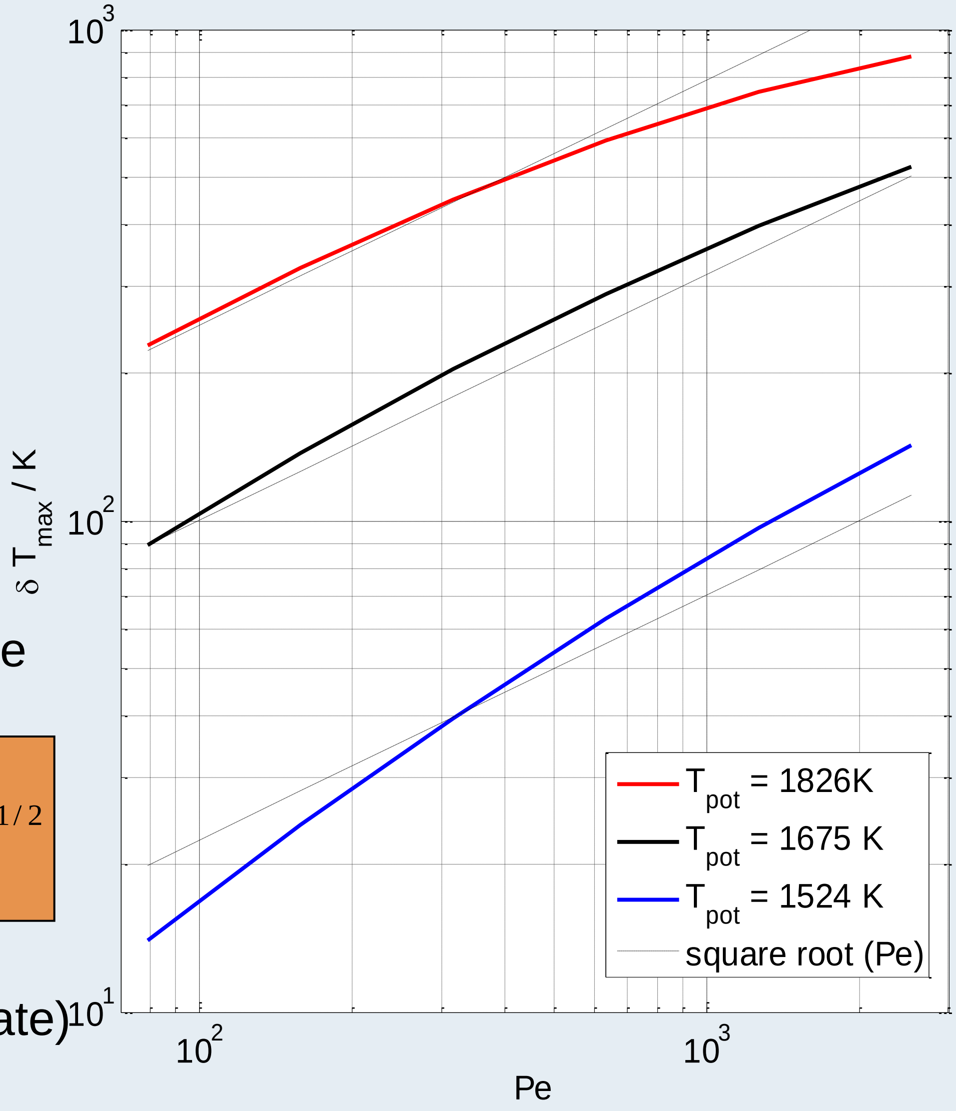
$\dot{e}_0$ - Strainrate of thinning
h_0 - Initial thickness
κ - Thermal diffusivity

- Potential temperature of mantle

$$\boxed{\delta T_{\max}' = \left(\frac{T_{pot} - T_{s0}}{c_1\, T_{Qtot}}\right)^2 \cdot Pe^{1/2}}$$

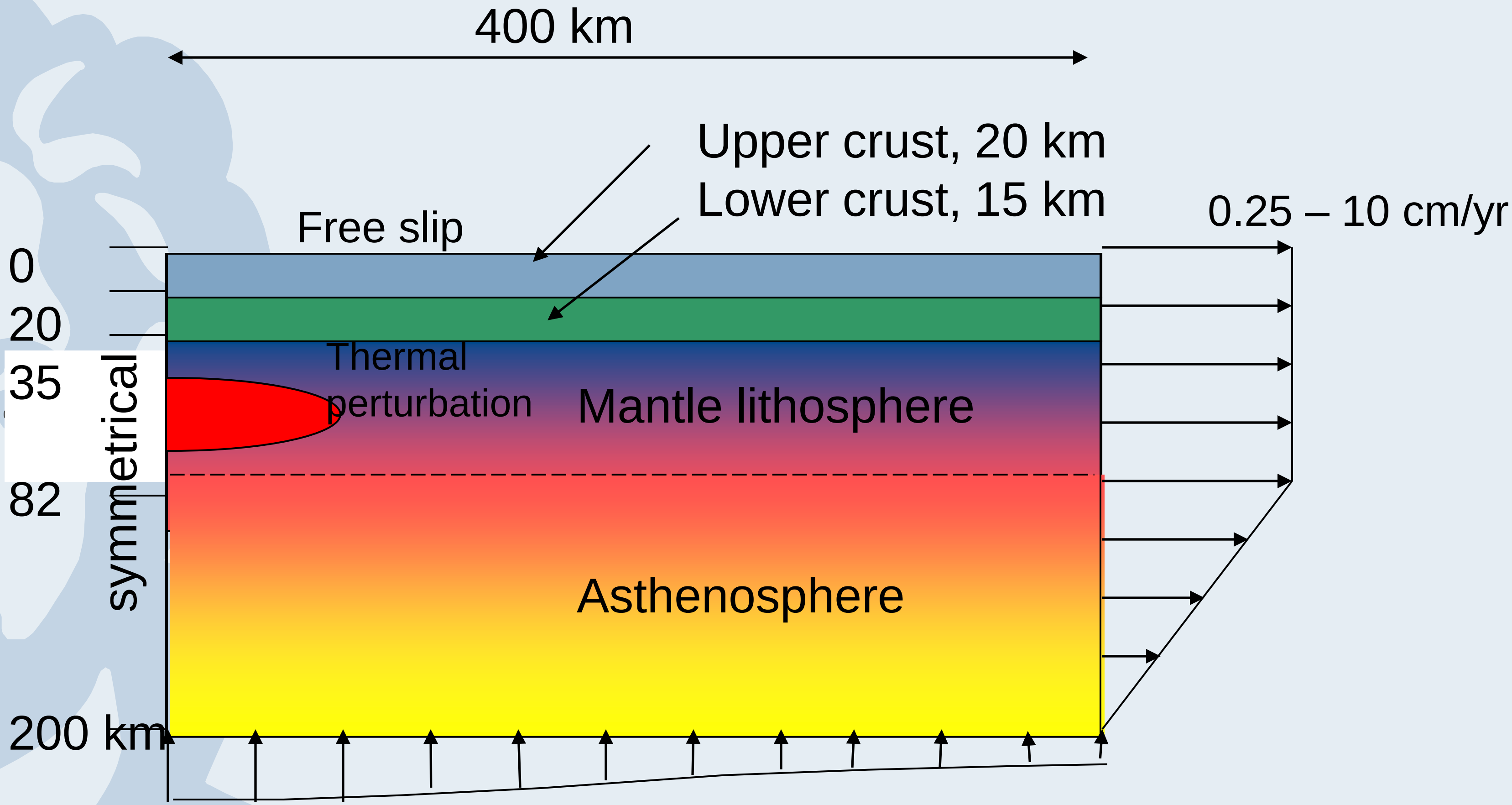
Scaling law, depends on:

- Square root of Pe (=thinning rate)
- Square of T_{pot} - $T_{solidus}$



Application to a dynamic rift model

Model setup


400 km
Upper crust, 20 km
Lower crust, 15 km
Free slip
0.25 – 10 cm/yr
0
20
35
82
200 km
symmetrical
Thermal
perturbation
Mantle lithosphere
Asthenosphere

Governing equations of two-phase flow melt - matrix

Mass

$$Melt: \frac{\partial(\rho_f \varphi)}{\partial t} + \vec{\nabla}(\rho_f \varphi \vec{v}_f) = \frac{DM}{Dt}$$

$$Matrix: \frac{\partial \rho_s (1-\varphi)}{\partial t} + \vec{\nabla}(\rho_s (1-\varphi)\vec{v}_s) = -\frac{DM}{Dt}$$

Momentum:

$$Melt: \vec{v}_f - \vec{v}_s = -\frac{k_\varphi}{\eta_f \varphi}(\vec{\nabla}P + \rho_f g \delta_{i3})$$

$$Matrix: -\rho g \delta_{i3} - \vec{\nabla}P + \frac{\partial \tau_{ij}}{\partial x_j} = 0$$

Rheological equ of state, (P-T-non-Newtonian), permeability relation

$$\tau_{ij} = \eta_s \left(\frac{\partial v_{si}}{\partial x_j} + \frac{\partial v_{sj}}{\partial x_i} \right) + \delta_{ij}(\eta_b - \frac{2}{3}\eta_s)\vec{\nabla}\vec{v}_s \qquad k_\varphi = \frac{a^2}{b}\varphi^n$$

Energy:

$$\rho c_p \left(\frac{\partial T}{\partial t} + \vec{v} \cdot \vec{\nabla} T + \frac{\alpha g}{c_p} v_z T \right) = \vec{\nabla} \cdot (k \vec{\nabla} T) + \rho H + \psi - L \left(\frac{\partial M}{\partial t} + \vec{v}_s \cdot \vec{\nabla} M \right) + Q_{intru}$$

Rheology

Creep laws:

- Mohr-Coulomb-Plasticity (Byerlee)
- P-T-dependent power law (from lab experiments);

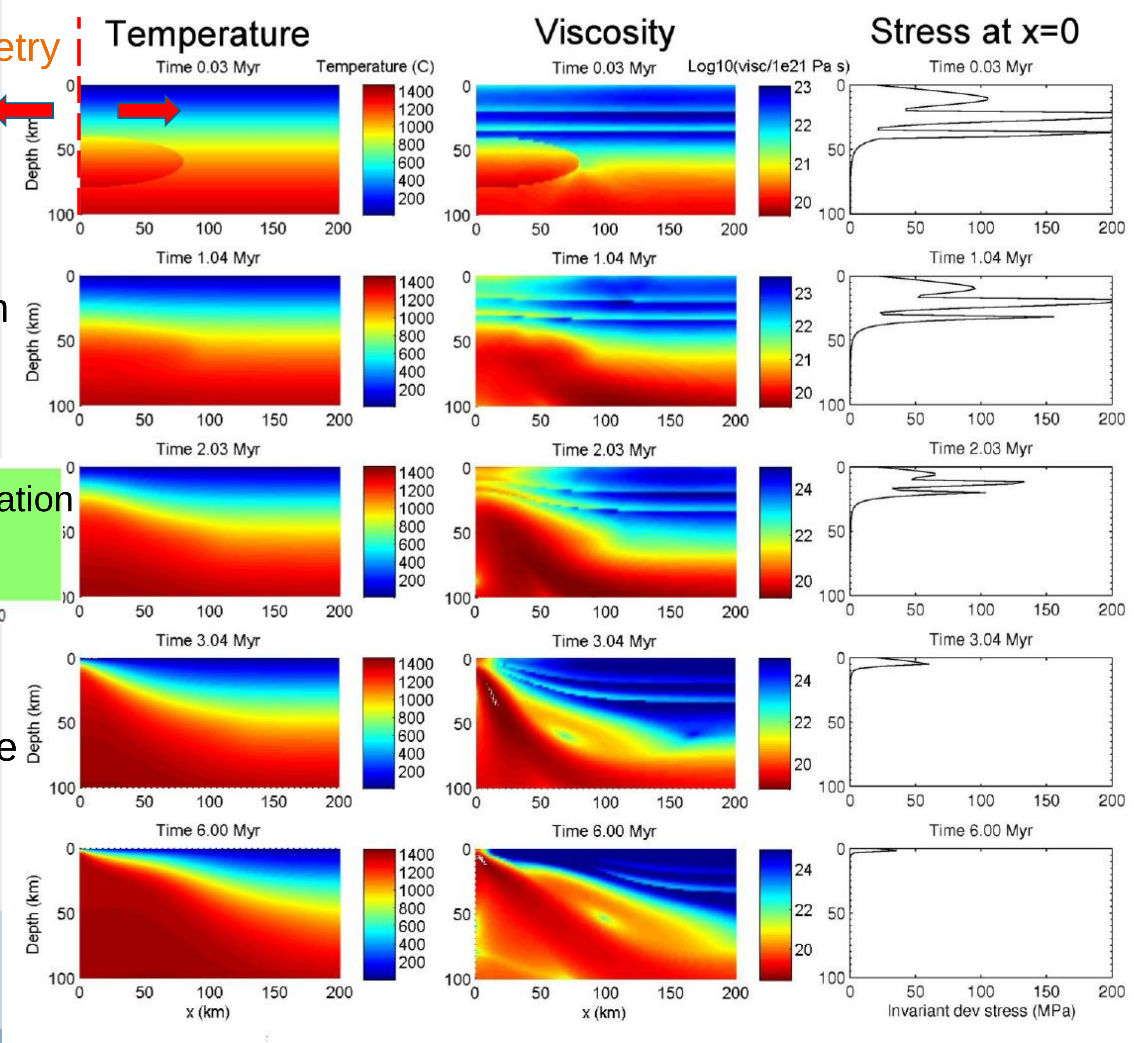
- Upper crust:	Westerly Granite, n = 3.5
- Lower crust:	Clinopyroxenite, n= 2.6
- Mantle (solid)	Aheim Dunite, n=3.6
- Mantle (partielly molten)	dto. weakened by exp(-a ϕ)

- no elasticity

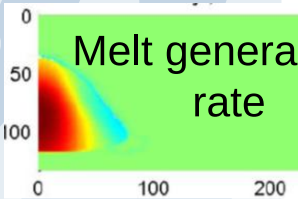
Numerical approach: 2D Finite Differences, markers

Dynamic rift model

Symmetry axis



- Hot asthenosphere at shallow depth
- Melt generation within upwelling asth.



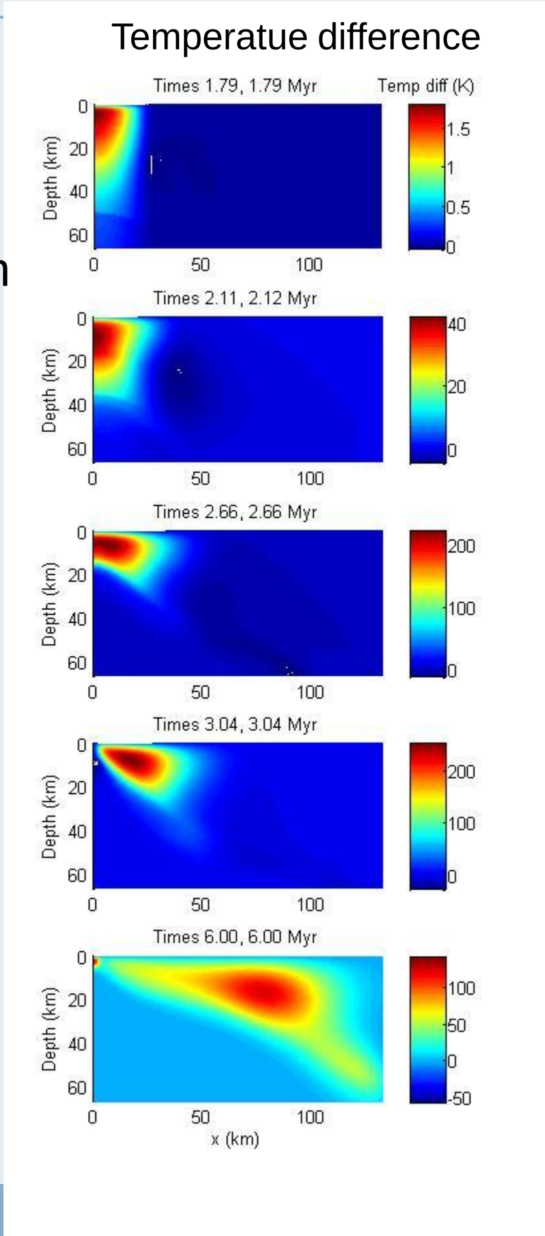
Melt generation rate

- Rheological stratification of the lithosphere
- 3 stress maxima
- Total lithospheric strength decreases

Effect of intrusional weakening

- Temperature increase by 200 - 250 K
- Compare with 1D prediction
  Pe = 521
  $T_{pot} – T_{solidus}$ = 250 K

→ δT_{max} = 209 K

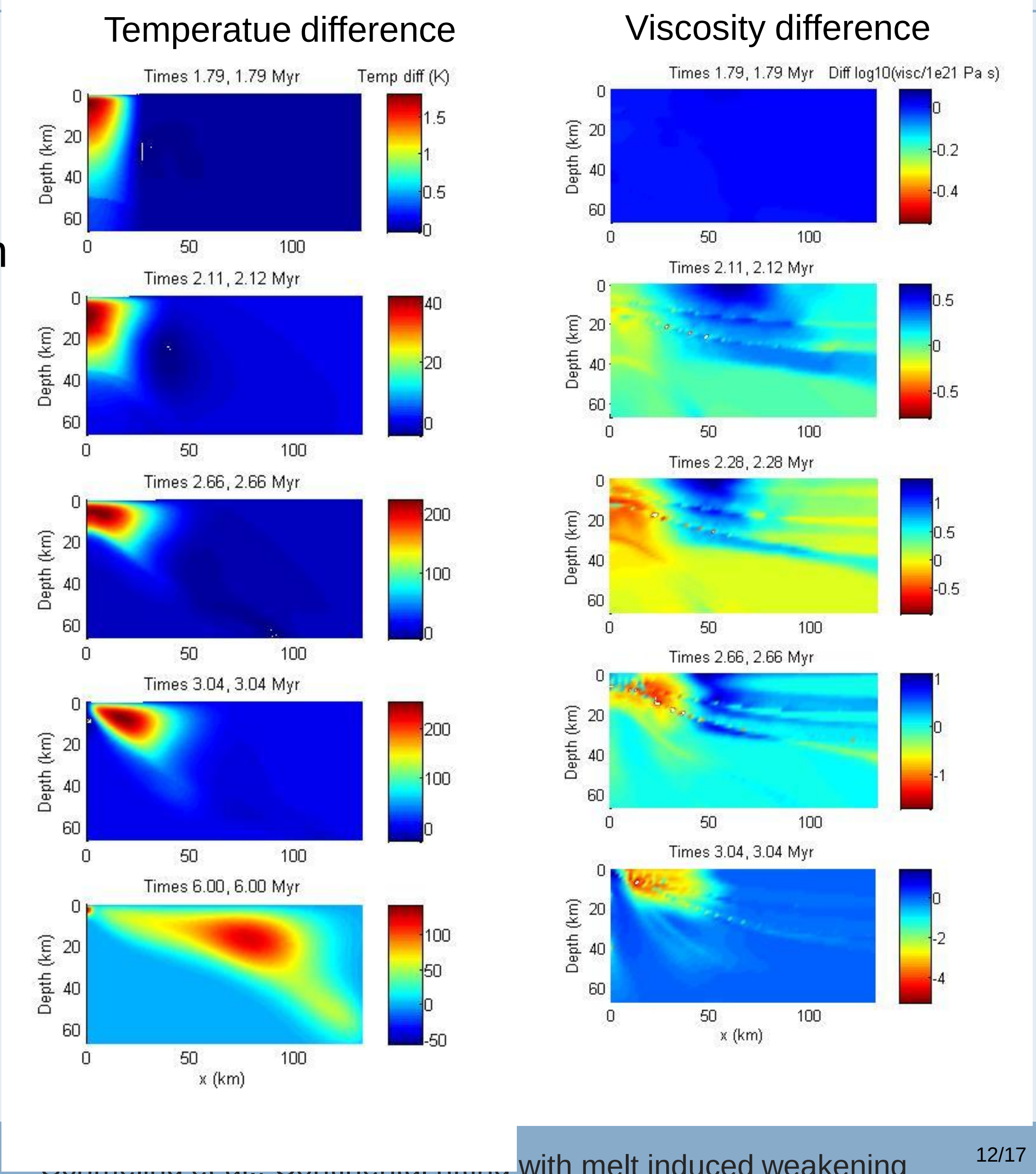


Effect of intrusional weakening

- Temperature increase by 200 - 250 K
- Compare with 1D prediction
  Pe = 521
  $T_{pot} - T_{solidus} = 250$ K

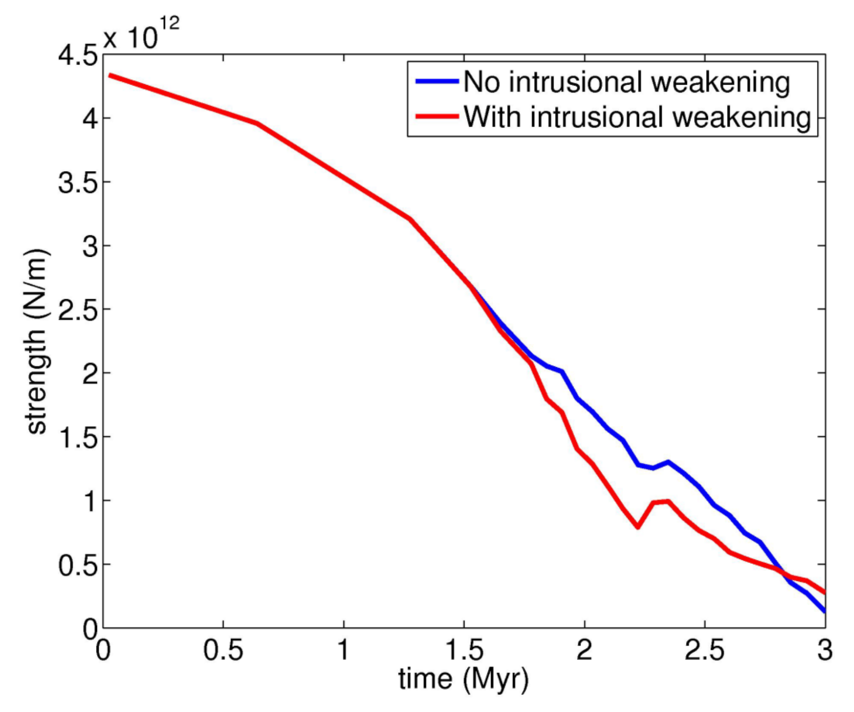
→ $\delta T_{max} = 209$ K

- Weakening: effective viscosity lower by up to one order of magnitude
- more effective melting (see below)

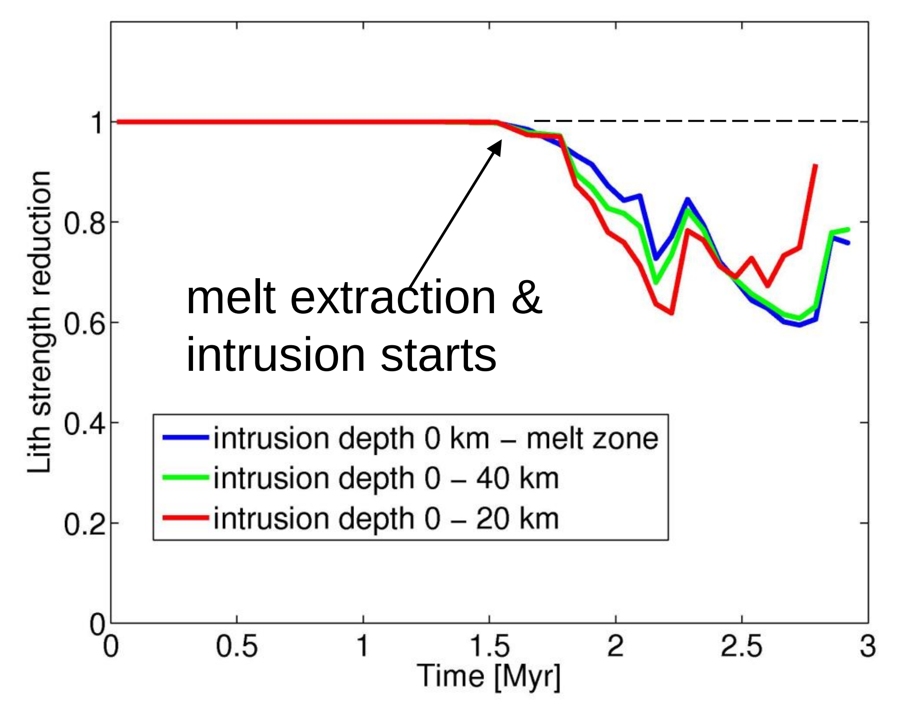


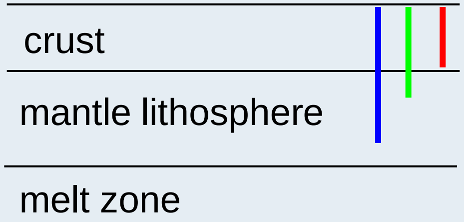
Effect of intrusional weakening

Reduction of lithospheric strength



Effect of emplacement depth

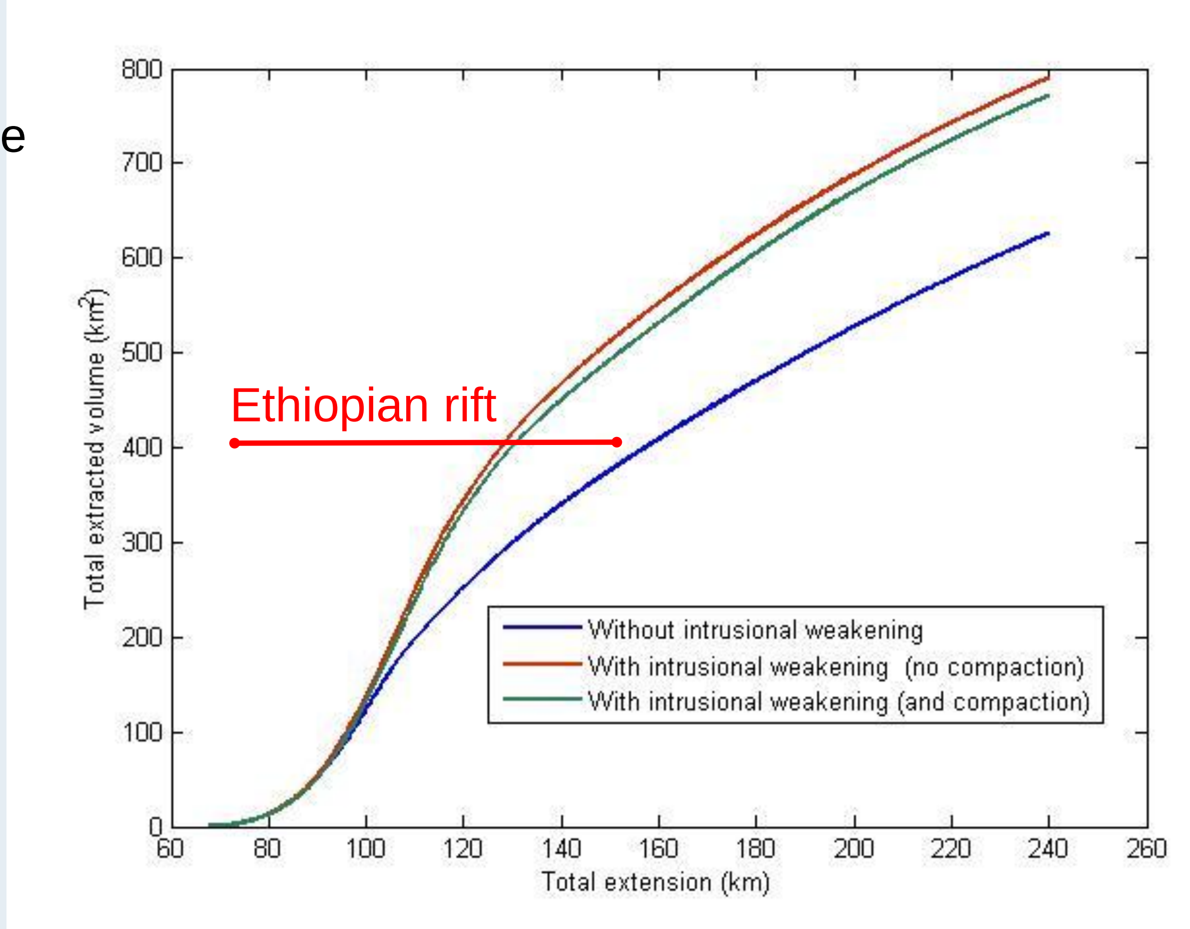




Effect of intrusional weakening

Feed back on meltproduction:

→ more effective rifting, higher melt production rate



Conclusions

Rifting with intrusional magma emplacement

1D kinematic thinning model:

- heating several 100 K
- Follow simple scaling law:
- Depends on square root of thinning and square of ($T_{pot} – T_{solidus}$)

2D dynamic rift model

- Rheological weakening by an order of magnitude or more
- strong decrease in lithospheric strength
- feed back on melt production
- May help explaining EARS magma volumes

- Compaction counteracts this effect, but only weakly