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Modelling of instabilities and bifurcation in Geomechanics, 6-8 Oct. 2016

Multiphysics couplings and strain localization in geomaterials

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WEAKENING MECHANISMS AND STRAIN LOCALIZATION

Deformation bands in geomechanics

- Deformation bands in the form of shear or compaction bands are observed are observed on a very large range of scales from sub-millimetric (grain size) to kilometric scale (geological structures).

- Strong heterogeneity of mechanical (e.g. strength) and physical properties (e.g. porosity, permeability) induced by the deformation bands.

- Major role of localized deformation bands
  - ✓ in the failure of engineering structures (e.g. foundations, oil wells instability..),
  - ✓ in the nucleation of earthquakes and landslides
  - ✓ in the flow of fluids (hydrocarbon exploration and production, deep waste storage repositories, CO2 sequestration, geothermal systems…)

Multiphysics weakening mechanisms

Softening behavior favors strain localization.

- Mechanical degradation of the rock properties (microcracking, grain crushing and grain size reduction…), (e.g. Das et al., 2011).

- Thermal pressurization of the pore fluid (e.g. Rice, 2006, Ghabezloo & Sulem, 2009)

- Chemical reactions such as dissolution/ precipitation, mineral transformation at high temperature (dehydration of minerals, decomposition of carbonates, …) (e.g. Castellanza & Nova, 2004, Hu & Hueckel, 2007, Sulem & Famin, 2009, Sin & Santamarina, 2010, Brantut & Sulem 2012).

Hydromechanical coupling and instability

Key mechanism of hydromechanical coupling in frictional geomaterials: dilation or contraction in due course of inelastic shear (e.g. dense/loose sands, tight/porous rocks)

- Inelastic volume changes in fluid-saturated geomaterials tend to cause a change in pore fluid pressure.

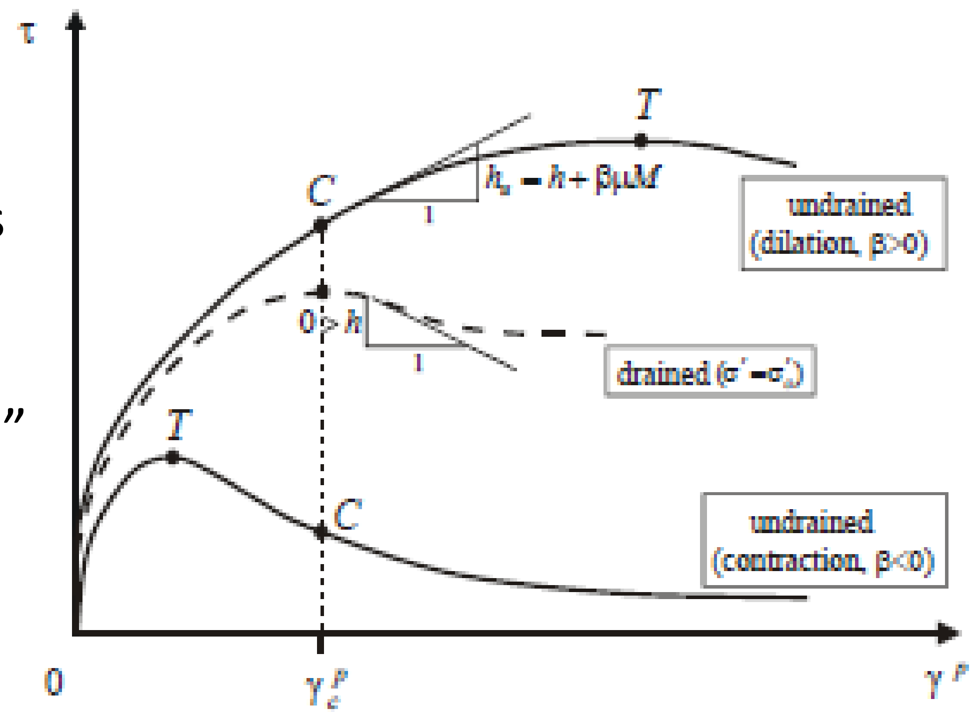
- Under drained boundary conditions and slow enough deformations, pore pressure remains constant as its alterations are equilibrated by pore fluid flow.

- Under undrained boundary conditions or if deformation occurs too rapidly for fluid flow to take place, pore pressure changes persist.

(e.g. Rice, 1975, Vardoulakis, 1986, 1995, Rudnicki & Garagash, 2000, Garagash, 2005, Benallal & Comi, 2003)

Hydromechanical coupling and instability

- Shear strength of geomaterials is affected by the effective compressive stress .
- Pore pressure drop in undrained shearing of dilatant material causes an increase of the undrained shear stress τ (dilatant hardening)
- Undrained shear stress “weakening” over the corresponding drained strength is observed in contracting geomaterials.
- Material strengthening, $h_u > h$, (or weakening), $h_u < h$, occurs for dilation (β> 0) (or contraction (β< 0).

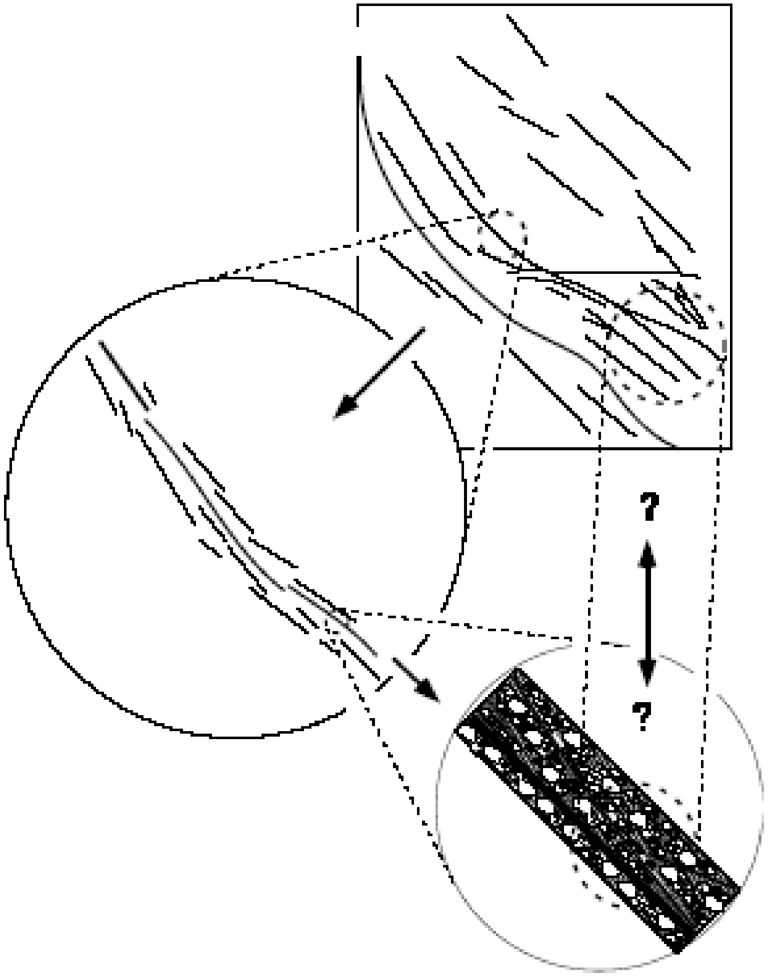


Thermo-Hydro-Chemo-Mechanical (THMC) coupling and instability

- Instability can be triggered by pore pressure changes: inelastic volume changes, thermal pressurization of the pore fluid, chemical reactions with release of a fluid phase (e.g. thermal decomposition of minerals) (e.g. Vardoulakis, 1996, Rice, 2006, Sulem et al., 2011, Rice et al. 2014).

- Instability can be triggered by softening of the solid skeleton: inelastic deformations, thermal weakening, chemical weakening (e.g. dissolution, dehydration reactions) (e.g. Rice, 1975, Brantut & Sulem, 2012 , Stefanou & Sulem, 2014, Sulem & Stefanou, 2016).

Fault structures at different scales



(Ben Zion & Sammis, 2003)

- <u>Regional scale</u> (1 à 100 km)

Network of oriented fractures of the earth crust (stress reorientation)

- <u>Local scale</u> (1 à 100m)

Individual faults in the network consist in an array of parallel deformation bands containing zones of intense fracturing (cataclasite) surrounded by damaged zones (breccia)

- <u>Small scale</u> (0.1 à 10 cm)

Shear band characterized by intense fracturation (ultra-cataclasite) and **strain localisation in very narrow slip zones (<1mm)**

Thickness of Principal Slip Zones
Examples from drilling in active faults

Fault system	Earthquake	Magnitude	Thickness of the PSZ	Reference
Nojima fault	Kobe, Japan (1995)	7.2	**1 mm**	Otsuki, 2003
Chelungpu fault	Chi Chi, Taiwan (1999)	7.6	**few mm**	Kuo et al., 2013
Longmenshan fault	Wenchuan, China (2008)	8	**1 cm**	Li et al. , 2013

Slip is localized in extremely thin zones

Energy partitioning during an earthquake

During an earthquake, the potential energy (mainly elastic strain energy and gravitational energy) stored in earth is released as:

- Radiated energy : Energy radiated by seismic waves

$\log_{10}E \sim 4.5 + 1.5\, M_w$ (E in joules, M_w is the magnitude of the earthquake)

For example for $M_w = 7$, $E = 10^{15}$ Joules, for $M_w = 9$, $E = 10^{18}$ Joules

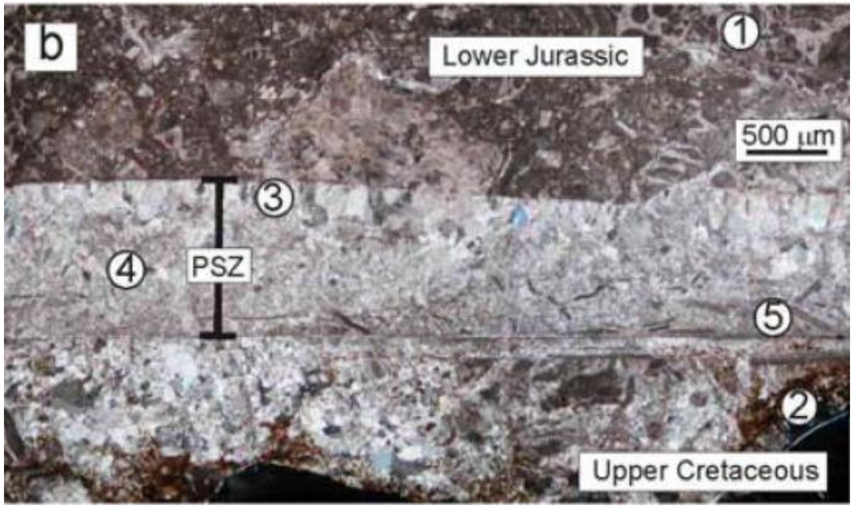
- Fracture energy: Energy associated with expanding the rupture area over the fault zone

- Thermal energy: Part of the frictional work (energy required to overcome fault friction) converted into heat

More than 90% of the mechanical work is dissipated into heat

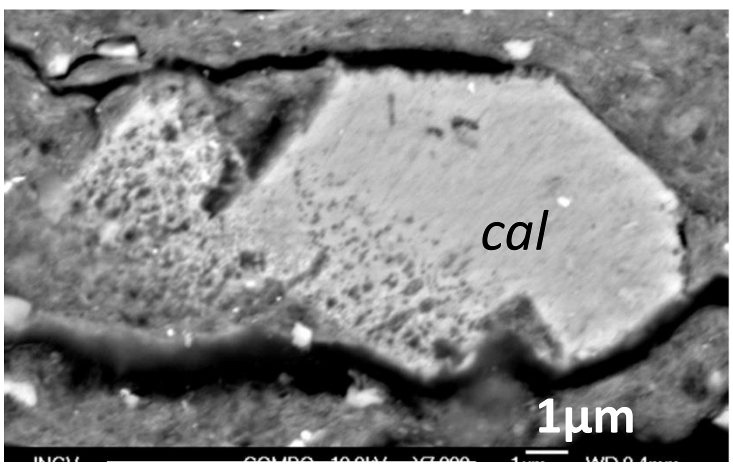
Thermally induced weakening mechanisms are of major importance

Observations of thermal decomposition of minerals in exhumed faults

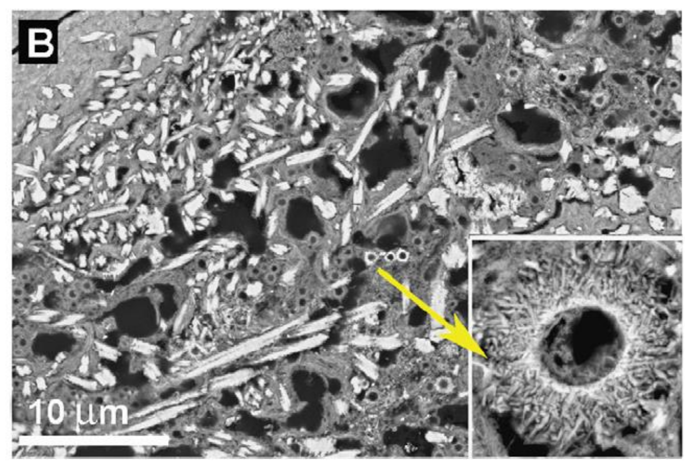


Spoleto thrust fault in Central Italy.
Principal slip zone (0.3 to 1mm)
5-10km of accumulated displacement

(from Collettini et al., 2012, *Geology*)



Calcite crystal showing decarbonation



Amorphous silicate phase: dehydration and amorphization of poorly cristalline clays

A key parameter: Width of the deformation band

Very narrow localized shear zone (typically ~ 100 µm) nested within the fault core where frictional heat is concentrated

Major role of the width of the slip zone:

- in the energy budget of the system: control of the feedback of the dissipative terms (e.g. frictional heating)
- in the rupture propagation mode (stronger weakening for thinner shear zones)

Evolution of the width of the slip zone in time:

Stronger weakening favors a decrease of the localized zone thickness, heat and fluid diffusion tend to broaden it.

SHEAR HEATING ANF PORE FLUID PRESSURISATION DURING RAPID SHEAR IN FLUID-SATURATED MATERIAL

Application to seismic slip

Effect of pore fluid pressure and temperature during seismic or gravitational slip

Thermal pressurization of pore fluids

- The permeability of the highly granulated fault gouge is very low.
- Fluids and heat are trapped inside the slip zone during an earthquake

Thermal pressurization of the fluid occurs because the thermal expansion coefficient of water is much greater than that of the rock particles.

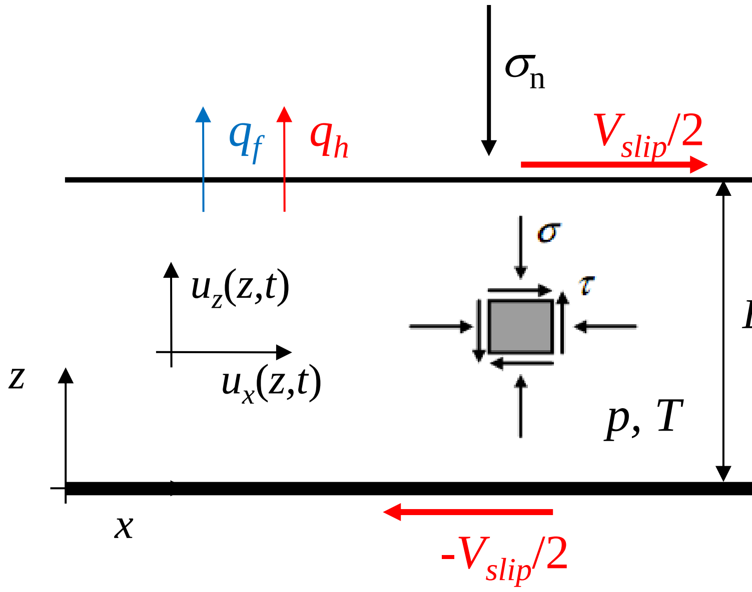
Shear heating is a mechanism of shear strength weakening because thermal pressurization of pore fluid reduces the effective stress

(Lachenbruch, 1980, Vardoulakis, 2002, Sulem et al. 2005, 2007, Rice 2006, Ghabezloo & Sulem, 2009, Rice et al. 2014).

Thermal pore fluid pressurization also occurs in large landslides (Habib, 1967, 1975, Vardoulakis, 2002, Veveakis et al, 2007)

Undrained adiabatic shearing of a saturated rock layer

Destabilizing effect of shear heating and pore fluid pressurization



$$\dot{\gamma}_0 = \frac{V_{slip}}{L}$$

Shear strain and volume strain

$$\gamma = \frac{\partial u_x}{\partial z} \qquad \varepsilon = \frac{\partial u_z}{\partial z}$$

Uniform state of stress in the layer

$$\frac{\partial \tau}{\partial z} = 0 \qquad \frac{\partial \sigma}{\partial z} = 0$$

It is assumed that the layer is at critical state (constant friction, no dilatancy)

$$\tau = \mu(\sigma - p)$$

Fluid mass balance

Fluid mass per unit volume of porous medium $m_f = \rho_f n$

n is the pore volume fraction (Lagrangian porosity)

ρ_f and is the density of the saturating fluid.

$$\frac{\partial m_f}{\partial t} = -\frac{\partial q_f}{\partial z}$$

q_f is the flux of fluid

$$\frac{\partial m_f}{\partial t} = n\frac{\partial \rho_f}{\partial t} + \rho_f \frac{\partial n}{\partial t}$$

$$q_f = -\frac{\rho_f}{\eta_f} k_f \frac{\partial p}{\partial z}$$

Darcy law for the fluid flow, with viscosity η_f through a material with permeability k_f

$$\frac{\partial \rho_f}{\partial t} = \rho_f \beta_f \frac{\partial p}{\partial t} - \rho_f \lambda_f \frac{\partial T}{\partial t}$$

$$\frac{\partial n}{\partial t} = n\beta_n \frac{\partial p}{\partial t} + n\lambda_n \frac{\partial T}{\partial t} + \frac{\partial n^p}{\partial t}$$

$\frac{\partial n^p}{\partial t}$: rate of plastic porosity change

$\beta_f = \frac{1}{\rho_f}\left(\frac{\partial \rho_f}{\partial P_p}\right)_T$: pore fluid compressibility

$\lambda_f = -\frac{1}{\rho_f}\left(\frac{\partial \rho_f}{T}\right)_{P_p}$: pore fluid thermal expansion coefficient

$\beta_n = \frac{1}{n}\left(\frac{\partial n}{\partial P_p}\right)_T$: pore volume compressibility $\beta_n = \frac{1}{n}\left(\beta_d - (1+n)\beta_s\right)$

$\lambda_n = \frac{1}{n}\left(\frac{\partial n}{\partial T}\right)_{P_p}$: pore volume thermal expansion coefficient

β_d : compressibility of the porous rock

β_s: compressibility of the solid phase

Pore fluid production and diffusion equation:

$$\frac{\partial p}{\partial t} = c_{hy} \frac{\partial^2 p}{\partial z^2} + \Lambda \frac{\partial T}{\partial t} - \frac{1}{\beta^*} \frac{\partial n^p}{\partial t}$$

$$\Lambda = \frac{\lambda_f - \lambda_n}{\beta_n + \beta_f}$$

is the coefficient of thermal pressurization

(typical values: 0.1 to 1 MPa/°C)

$\beta^* = n\left(\beta_n + \beta_f\right)$ is the storage coefficient.

For incompressible fluid and solid phase $\beta^* = 1/K$

$c_{hy} = k_f / (\beta \eta_f)$ is the hydraulic diffusivity

For plastically incompressible solid matrix:

$$\frac{\partial n^p}{\partial t} = \frac{\partial \varepsilon^p}{\partial t} = 0 \text{ for zero dilatancy}$$

Energy balance equation

$$\rho C \frac{\partial T}{\partial t} = E_F - \frac{\partial q_h}{\partial z}$$

E_F is the rate of frictional heat produced during slip

ρC is the specific heat per unit volume of the fault material

q_h is the heat flux and according to the Fourier law is proportional to the temperature gradient

$$q_h = -k_T \frac{\partial T}{\partial z}$$

k_T is the thermal conductivity of the saturated rock

It is assumed that all the plastic work is converted into heat

$$E_F = \tau \dot{\gamma}^p \approx \tau \dot{\gamma}$$

$$\frac{\partial T}{\partial t} = c_{th} \frac{\partial^2 T}{\partial z^2} + \frac{1}{\rho C} \tau \dot{\gamma}$$

$c_{th} = k_T / \rho C$ is the thermal diffusivity

Spatially uniform solution under undrained adiabatic conditions

The drainage and the heat flux are prohibited at the boundaries of the layer.

$$q_f = 0 \text{ and } q_h = 0$$

The normal stress σ_n acting on the sheared layer is constant.

$$\dot{\sigma} = 0$$

The undrained adiabatic limit is applicable as soon as the slip event is sufficiently rapid and the shear zone broad enough to effectively preclude heat or fluid transfer (e.g. earthquakes, landslides).

Summary of the governing equations

$$\text{mass balance: } \frac{\partial p}{\partial t} = \Lambda \frac{\partial T}{\partial t}$$

$$\text{energy balance: } \frac{\partial T}{\partial t} = \frac{1}{\rho C}\left(\sigma_n - p\right)\mu\dot{\gamma}_0$$

Spatially uniform solution under undrained adiabatic conditions

Solution:

$$p = p_0 + (\sigma_n - p_0)\left(1 - \exp\left(-\frac{\mu\Lambda}{\rho C}\dot{\gamma}_0 t\right)\right)$$

$$T = T_0 + \frac{(\sigma_n - p_0)}{\Lambda}\left(1 - \exp\left(-\frac{\mu\Lambda}{\rho C}\dot{\gamma}_0 t\right)\right)$$

In undrained adiabatic conditions, the pore-pressure increases towards its geostatic limit σ_n which corresponds to full fluidization exponentially with the slip displacement.

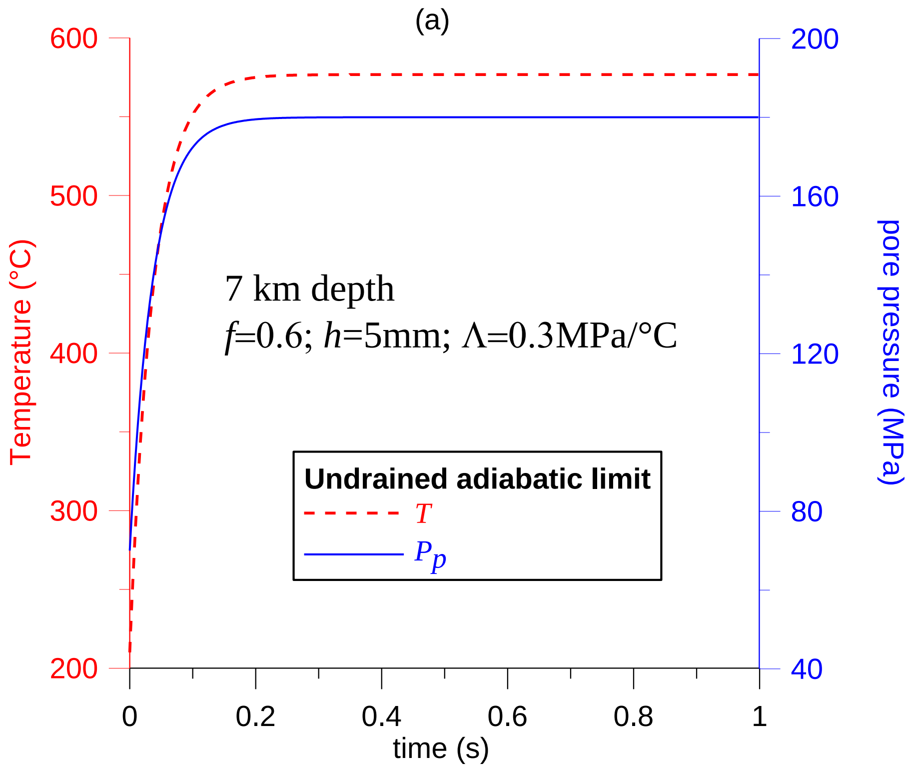
In due course of the shear heating and fluid pressurization process, the shear strength τ is reduced towards zero.

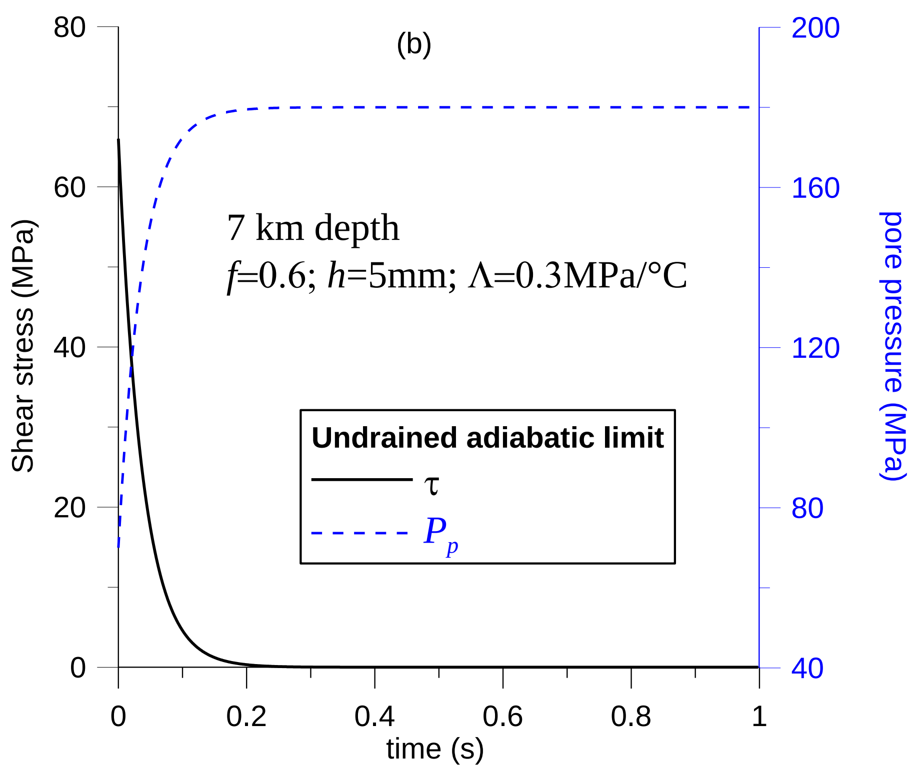
Example: Fault zone at 7 km depth

Initial conditions: $T_0 = 210°\text{C},\ p_0 = 70\text{MPa},\ \sigma_n = 180\text{MPa}$

Slip velocity: 1m/s; Shear band thickness: L=5mm

Maximum temperature: $T_{max} = T_0 + \dfrac{\sigma_n - p_0}{\Lambda} = 577°\text{C}$





Linear stability analysis of uniform shear

Rice et al., 2014

Rate-dependency of the friction coefficient is considered

$$\mu = \mu_0 + H \log \frac{\dot{\gamma}}{\dot{\gamma}_0}$$

Stability condition:

$$\lambda < \lambda_{cr}, \text{ with } \lambda_{cr} = 2\pi \sqrt{\frac{H \rho C \left(c_{th} + c_{hy} \right)}{\mu_0 \Lambda \left(\mu_0 + 2H \right) \dot{\gamma}_0}}$$

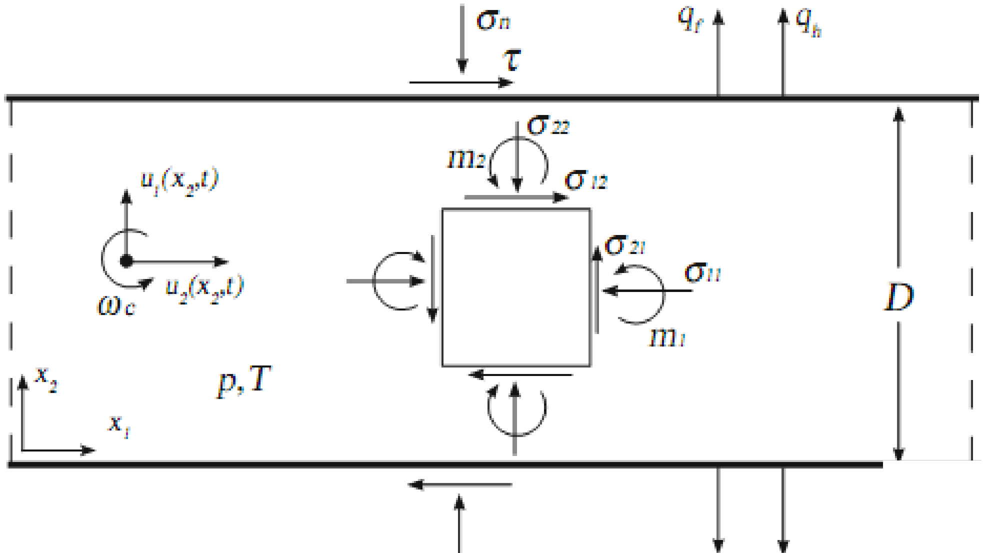
Competing processes: Fluid and thermal diffusion and rate-dependent frictional strengthening tend to expand the localized zone, while thermal pressurization tends to narrow it.

Only shear zones with a thickness $h < \lambda_{cr}/2$ will support stable homogeneous shear

Typical values at seismogenic depth: few tens of microns

The localized zone thickness may be comparable with the gouge grain size

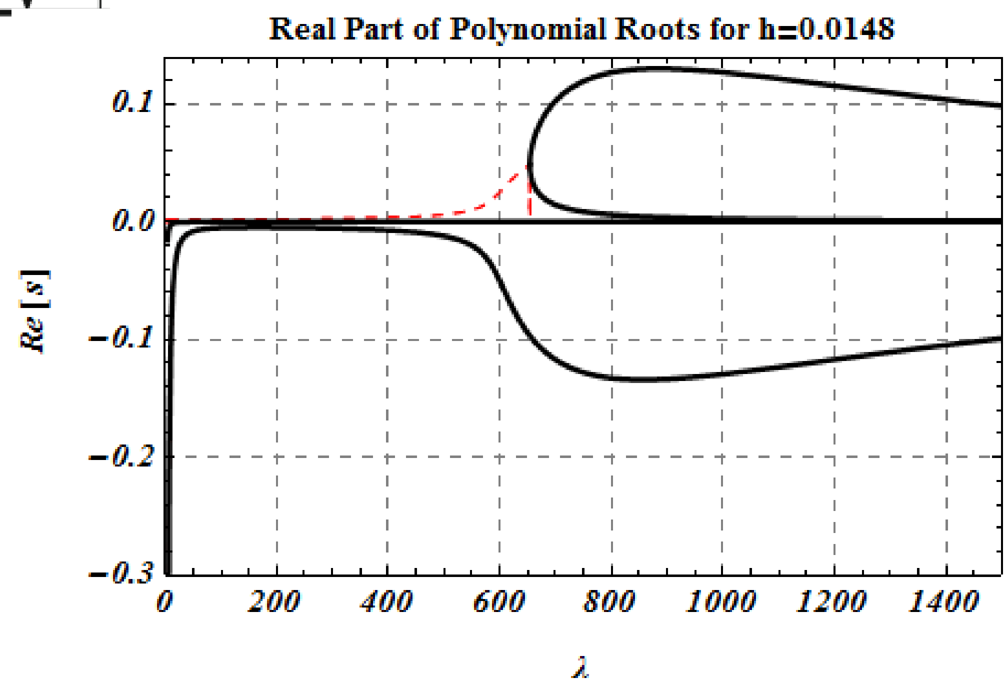
Stability analysis of undrained adiabatic shearing of a rock layer with Cosserat microstructure



Sulem, Stefanou, Veveakis(2011), *Granular Matter*

Strain hardening elasto-plasticity for 2D Cosserat continuum
Material length *R* related to the material grain size

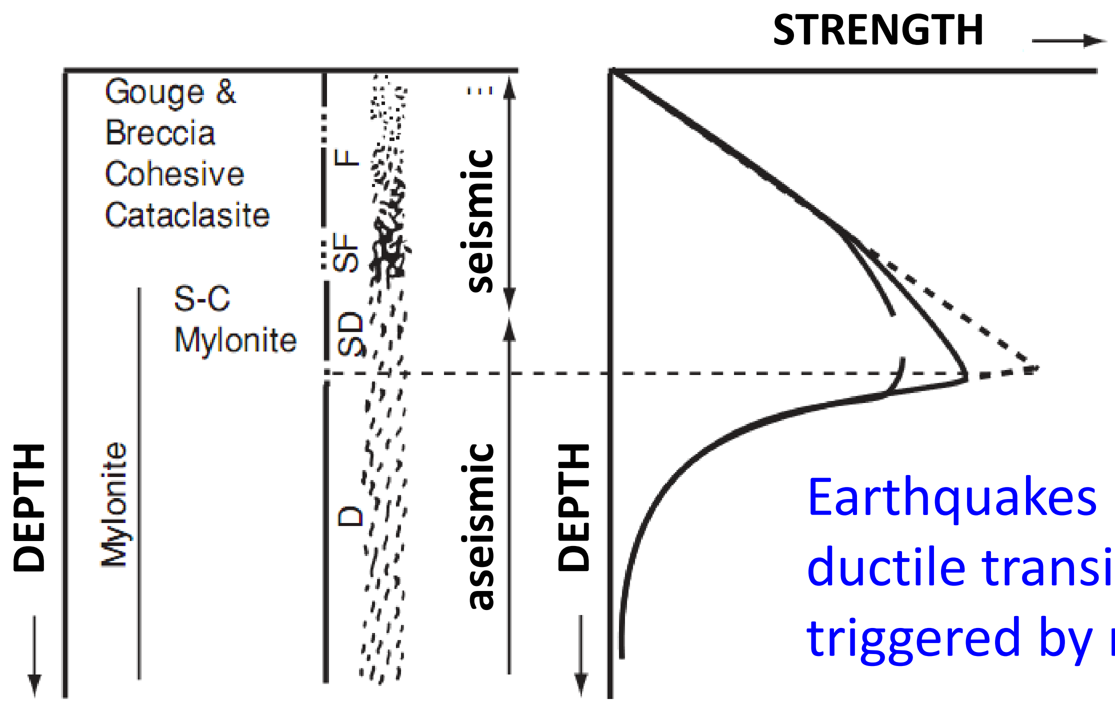
Dominant instability mode with a selected finite wave length and a finite growth coefficient



With *R*=1 to 10μm (grain size for highly finely granulated fault core) the obtained localized zone thickness is about 0.1 to 1 mm.

THCM COUPLINGS AND STABILITY OF FAULT ZONES

Deep earthquakes triggered by metamorphic reactions



Earthquakes below the brittle-ductile transition of the crust may be triggered by metamorphic reactions

Metamorphic dehydration reactions may produce weaker products
example: dehydration of lizardite (serpentinite)

$$\underbrace{5\,\mathrm{Mg_3Si_2O_5(OH)_4}}_{\text{lizardite}} \longrightarrow \underbrace{\mathrm{Mg_3Si_4O_{10}(OH)_2}}_{\text{talc}} + 6\,\underbrace{\mathrm{Mg_2SiO_4}}_{\text{olivine}} + 9\,\mathrm{H_2O}.$$

Ultra-fine grained olivine weaker than the serpentinite agregates

Chemically weakening and slip instability

Brantut & Sulem, (2012), *J. Appl. Mech.*

Reaction rate:

$$\frac{\partial \xi}{\partial t} = A(1-\xi)\exp\left(-\frac{E_a}{RT}\right)$$

Constitutive model:
(rate hardening/reaction weakening)

$$\tau = f(\dot{\gamma},\xi)\sigma', \quad f(\dot{\gamma},\xi) = f_0 + a\ln(\dot{\gamma}/\dot{\gamma}_0) - b\xi$$

Mass balance:

$$\frac{\partial p}{\partial t} = c_{hy}\frac{\partial^2 p}{\partial z^2} + \Lambda\frac{\partial T}{\partial t} + \frac{1}{\rho_f \beta^*}\frac{\partial m_d}{\partial t} - \frac{1}{\beta^*}\frac{\partial n_d}{\partial t}$$

Fluid diffusion | Thermal pressurization | Fluid production | Inelastic porosity change

Energy balance:

$$\frac{\partial T}{\partial t} = c_{th}\frac{\partial T^2}{\partial z^2} + \frac{\tau\dot{\gamma}}{\rho C} - m_0\frac{\Delta H}{\rho C}\frac{\partial \xi}{\partial t}$$

Heat diffusion | Frictional heat | Heat consumed in the chemical reaction

Effect of lizardite dehydration @ 30km depth along subduction zones

Table 1
Parameter values for lizardite dehydration at a depth of around 30 km.[44]

Quantity	Value
Friction coefficient, f_0	0.6
Rate strengthening parameter, a	0.002
Reaction weakening parameter b	0.5
Specific heat capacity, ρC	2.7 MPa °C^{-1}
Thermal dependency of the chemical kinetics, c_T	2.58×10^{-7} °C^{-1} s^{-1}
Depletion dependency of the chemical kinetics, c_μ	2.12×10^{-6} s^{-1}
Initial shear stress, τ_0	240 MPa
Nominal strain rate, $\dot{\gamma}_0$	10^{-6} s^{-1}
Thermal pressurization coefficient, Λ	0.5 MPa °C^{-1}
Thermal diffusivity, c_{th}	10^{-6} m^2 s^{-1}
Hydraulic diffusivity, c_{hy}	10^{-6} m^2 s^{-1}

Linear stability analysis

$$\dot{\gamma}_0 = 10^{-6} \text{ s}$$

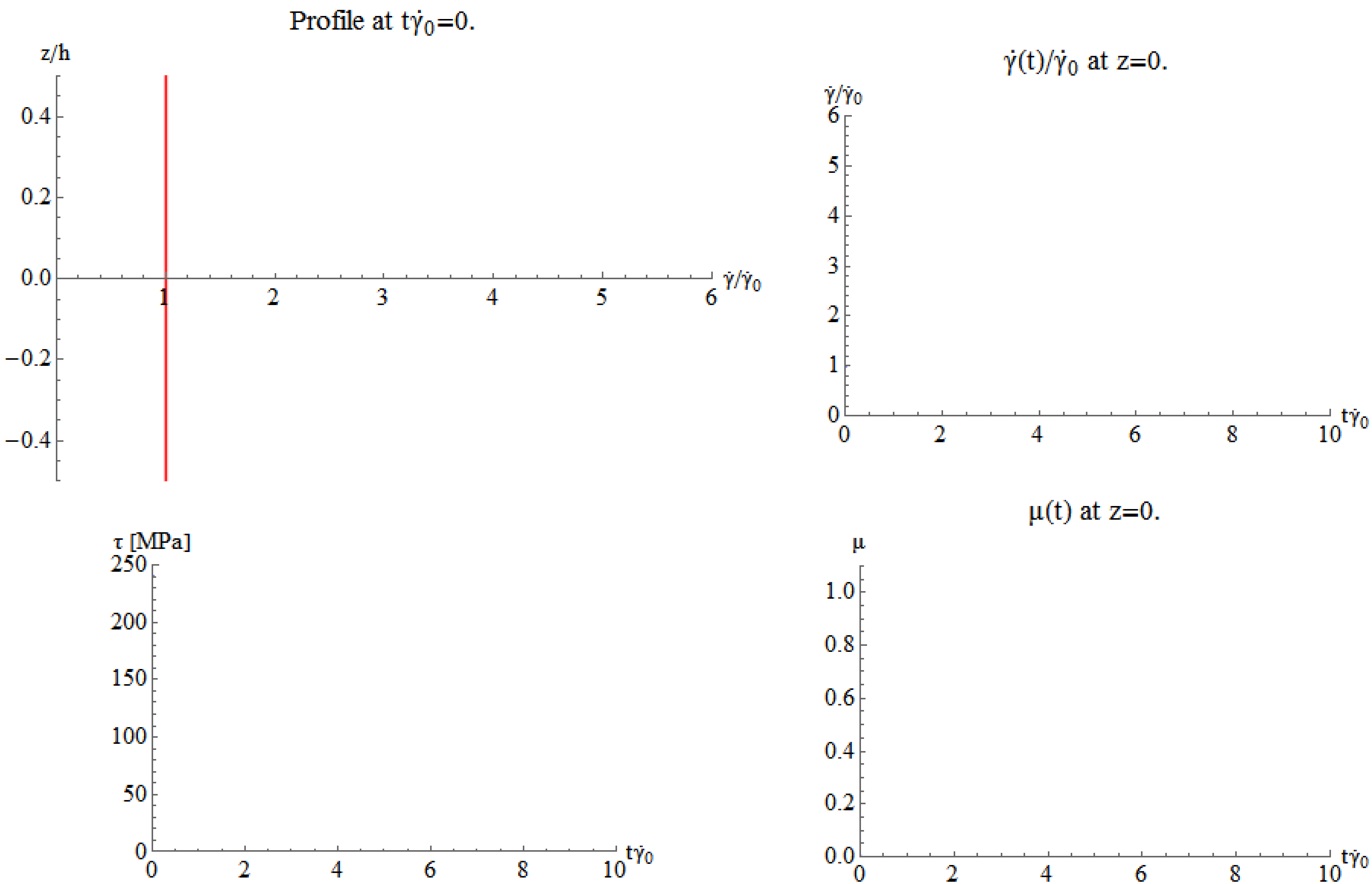
$$D = 5 \text{ m}$$

$$\lambda_{cr}^{ch} = 0.12 \text{ m}$$

$$\lambda_{cr}^{ch} = 2\pi \sqrt{\frac{a c_{th}}{\gamma_0} \frac{\rho C}{b \tau_0} \frac{c_\mu}{c_T}}$$

Short lived slip instability (depletion of the reactant)
Nucleation of transcient slip events, 'slow' earthquakes

$$\dot{\gamma}_0 = 10^{-6}\mathrm{s}$$

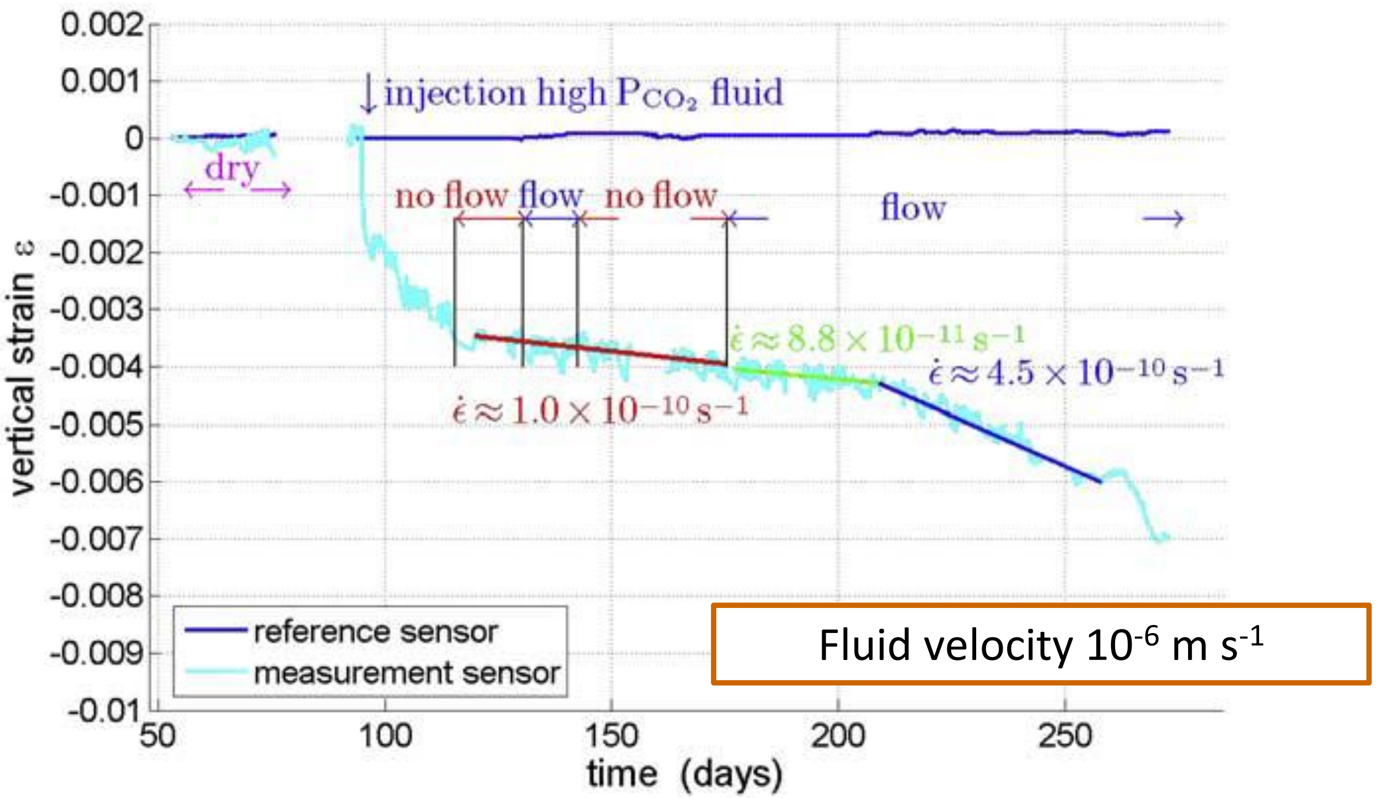


CHEMICAL DISSOLUTION AND COMPACTION BANDS: AN EXAMPLE OF MULTI-SCALE ANALYSIS

Strong coupling between chemical weakening and dissolution kinetics

Stefanou & Sulem, 2014, *J. Geoph. Res*.

Creep due to CO_2 injection in Lavoux limestone



Y. Le Guen, F. Renard, R. Hellmann, E. Brosse, M. Collombet, D. Tisserand, and J.-P. Gratier, "Enhanced deformation of limestone and sandstone in the presence of high P_{CO2} fluids," *Journal of Geophysical Research*, 2007, 112.

R. H. Brzesowsky, S. J. T. Hangx, N. Brantut, and C. J. Spiers, "Compaction creep of sands due to time-dependent grain failure: Effects of chemical environment, applied stress and grain size," *Journal of Geophysical Research*, 2014, 119.

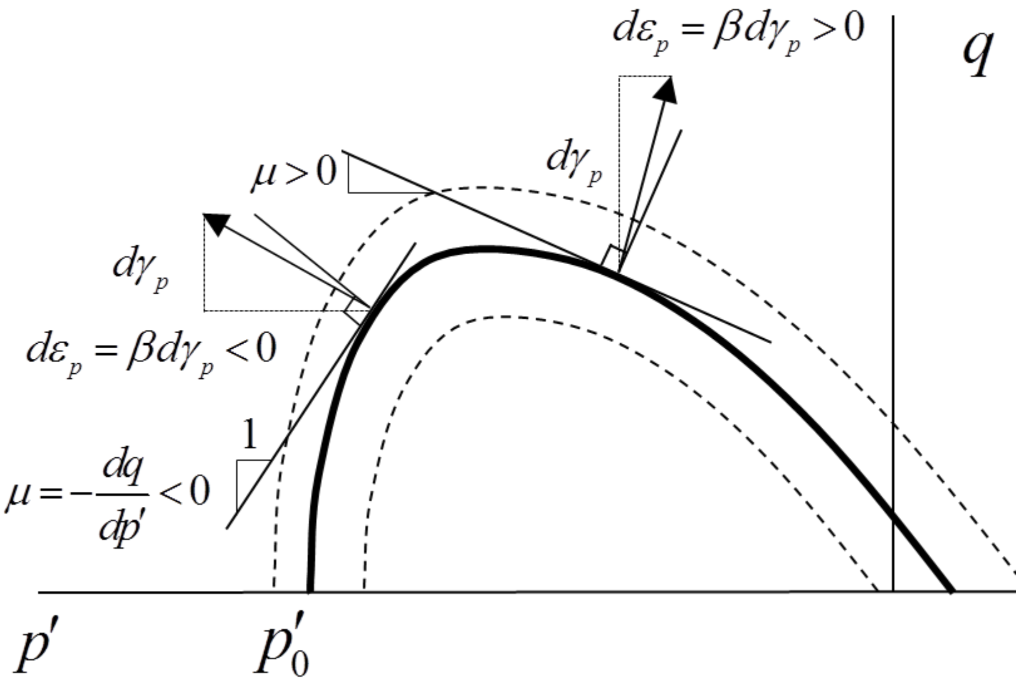
Is the deformation homogeneous?

Pure compaction bands?

What is the influence of a reactive fluid flow on

deformation band formation?

Conditions for localization due to mechanical softening



Criterion for compaction bands:

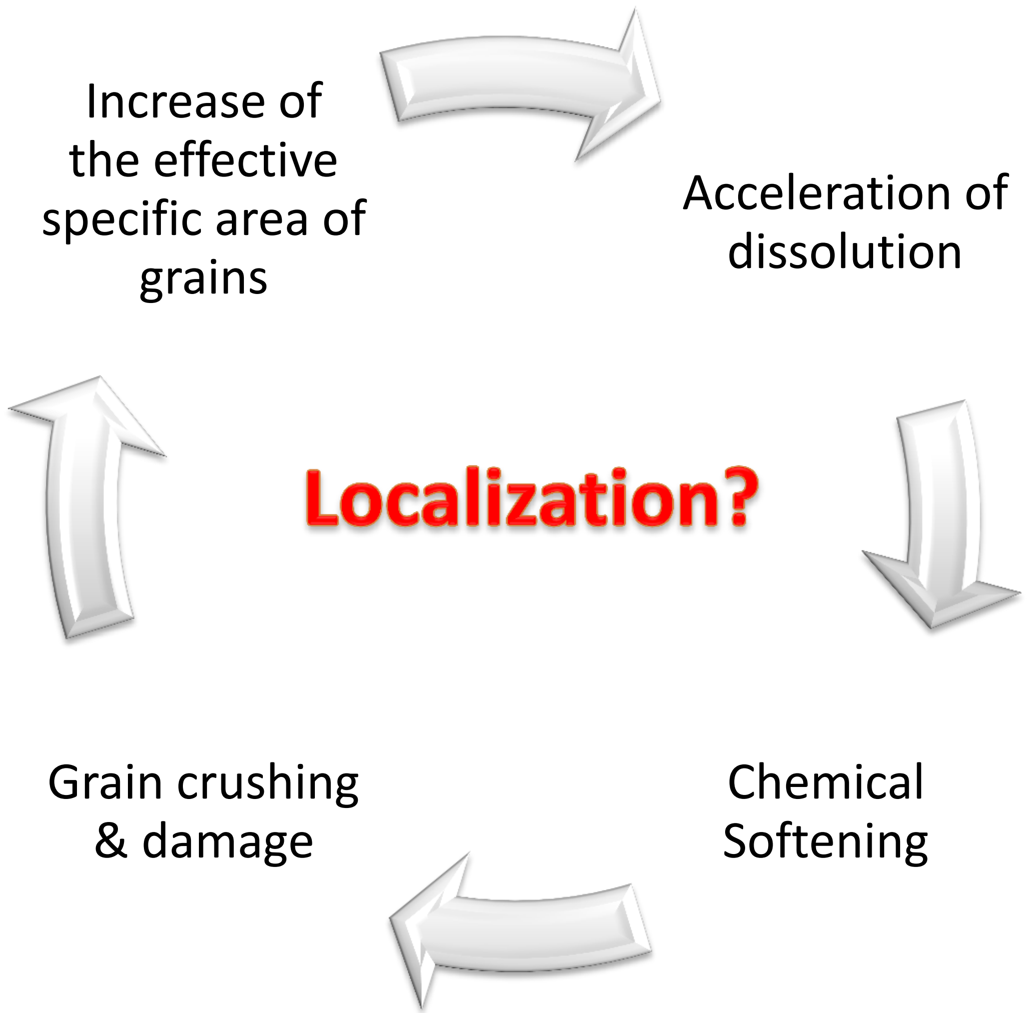
$$\beta + \mu \leq -\sqrt{3}$$

Perfect associate plasticity:

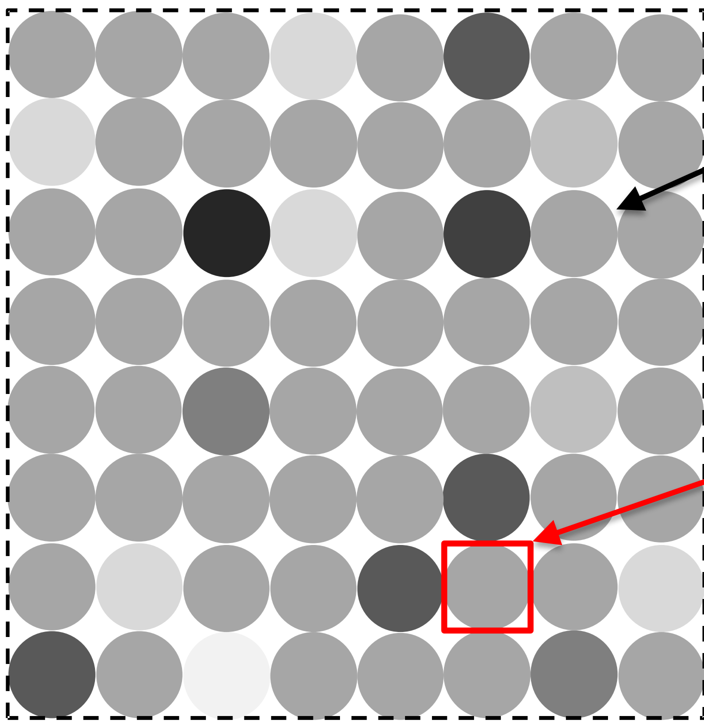
$$\beta = \mu = -\frac{\sqrt{3}}{2}$$

Issen & Rudnicki (2000)
Rudinicki & Rice (1975)

Conceptual model & chemical softening



Distinction of scales



Macro-scale / Elementary volume (REV)

- Constitutive behavior
- Momentum balance
- Mass balance

Micro-scale / Single Grain

- Reaction kinetics of dissolution
- Grain crushing, solid skeleton damage, microcracking …

Reaction kinetics (micro-scale)

$$\underset{(3)}{\text{solid}} + \underset{(1)}{\text{solvent}} \rightleftharpoons \underset{(2)}{\text{solution}}$$

E.g. dissolution of quartz: $SiO_2\text{(solid)}+2H_2O\text{(liquid)} \rightleftharpoons H_4SiO_4\text{(aqueous solution)}$

or diss. of carbonates: $CaCO_3\text{(solid)}+H_2CO_3\text{(aqueous solution)} \rightleftharpoons Ca\left(HCO_3\right)_2\text{(aqueous solution)}$

$$\frac{\partial w_2}{\partial t} = k^* \frac{S}{e}\left(1 - \frac{w_2}{w_2^{eq}}\right)$$

w_2 is the mass fraction of the dissolution product in the fluid

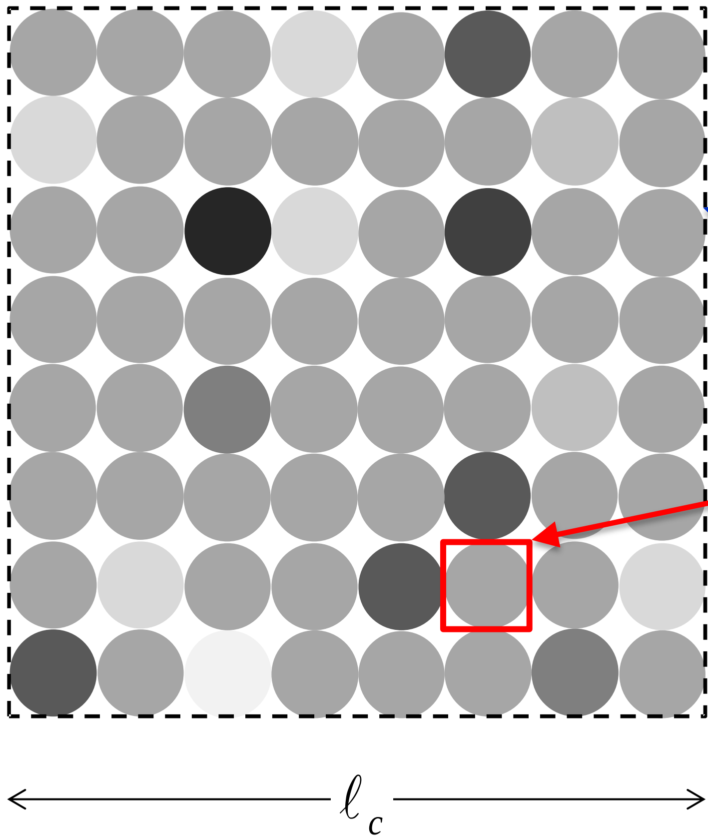
k^* is a reaction rate coefficient

e is the void ratio

$S \propto \frac{1}{D}$ is the specific area of a single grain of diameter *D*

Upscaling of the dissolution process

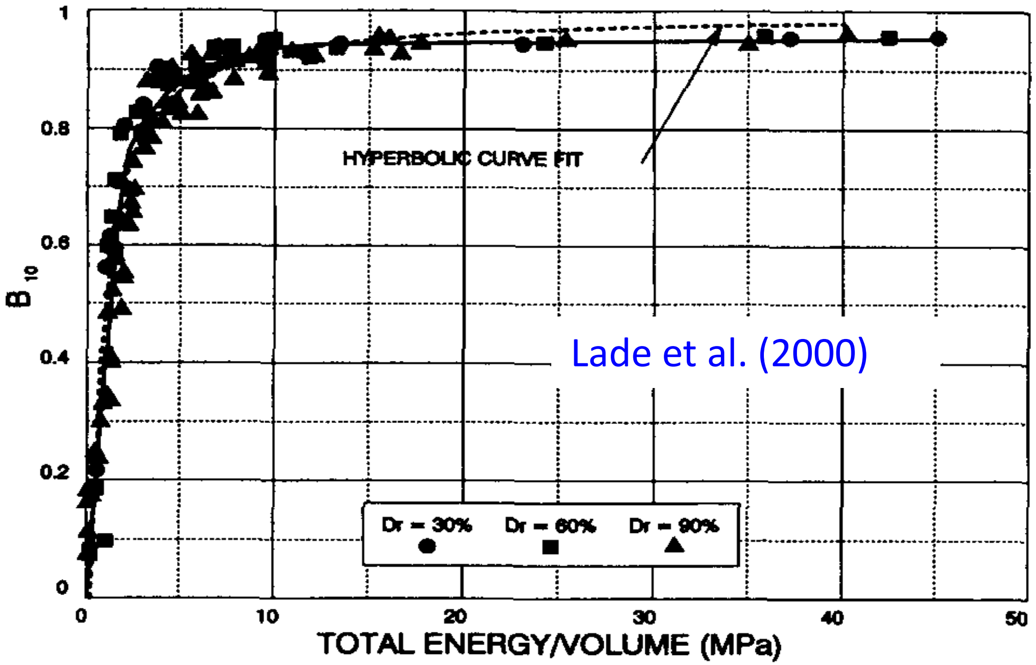
micro > macro



$$w_2^M = \frac{1}{V_T}\int_{V_T} w_2 dV \approx w_2 + \ell_c^2 \frac{\partial^2 w_2}{\partial z^2}$$

ℓ_c : Characteristic length

Evolution of the effective grain size
Grain crushing (micro-scale)

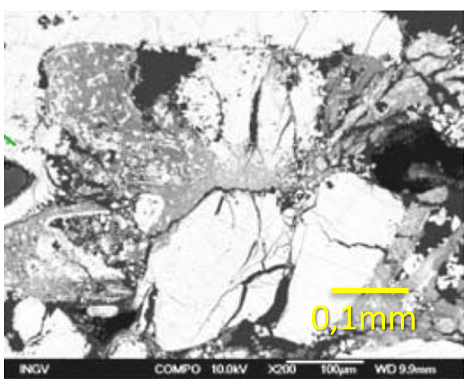


Lade et al. (2000)

$$D = D_0\left(\frac{a}{a+E_T}\right)$$

or

$$S = S_0\left(1+\frac{E_T}{a}\right)$$



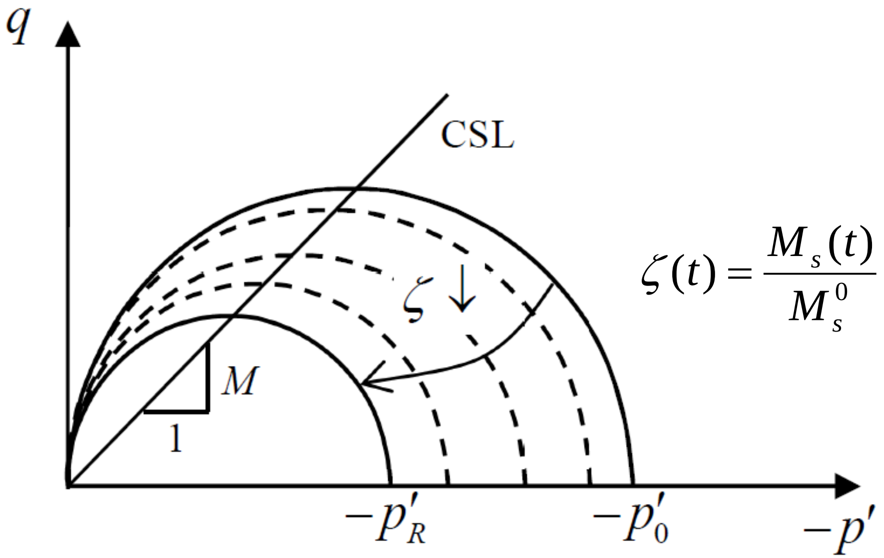
Baud et al. (2009)

a is a material constant which expresses the influence of grain crushing

E_T is the total energy density given to the system

Grain breakage: Einav (2007), JMPS

Constitutive behavior (macro-scale)



Modified Cam-Clay plasticity model as an example

Plasticity criterion

$$f \equiv q^2 + M^2 p'(p' - p'_c) = 0$$

$$p'_c \equiv p'_R - \left(p'_R - p'_0\right)\zeta^{\kappa}$$

Evolution of chemical softening

$$\frac{\partial \zeta}{\partial t} = -\frac{\mu_3}{\mu_2}\frac{\rho_f}{\rho_s} e \zeta \frac{\partial w_2^M}{\partial t}$$

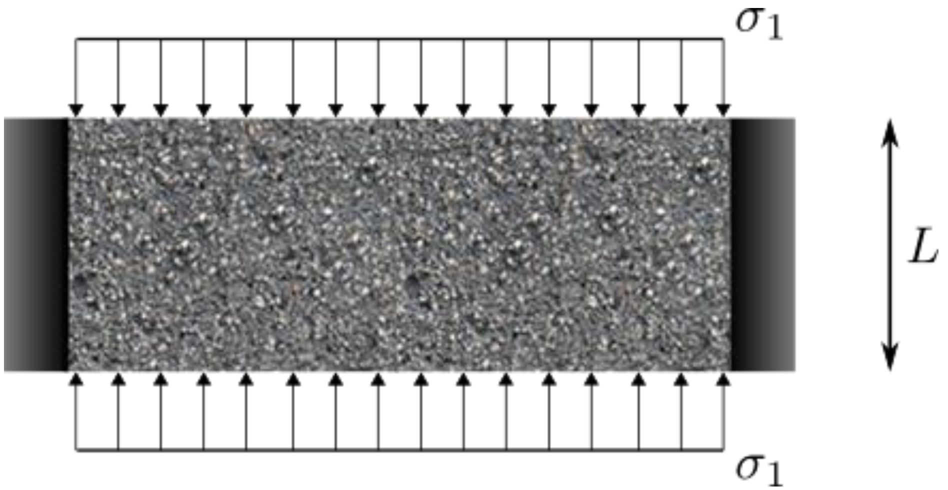
$$\frac{\partial w_2^M}{\partial t} = \frac{\partial w_2}{\partial t} + \ell_c^2 \frac{\partial}{\partial t}\frac{\partial^2 w_2}{\partial z^2}$$

Non-local constitutive law

Bo & Hueckel (2007) Comp. Geot.

Equilibrium (macro-scale)

$$\frac{d\sigma_1}{dz} \approx 0$$



Oedometric conditions

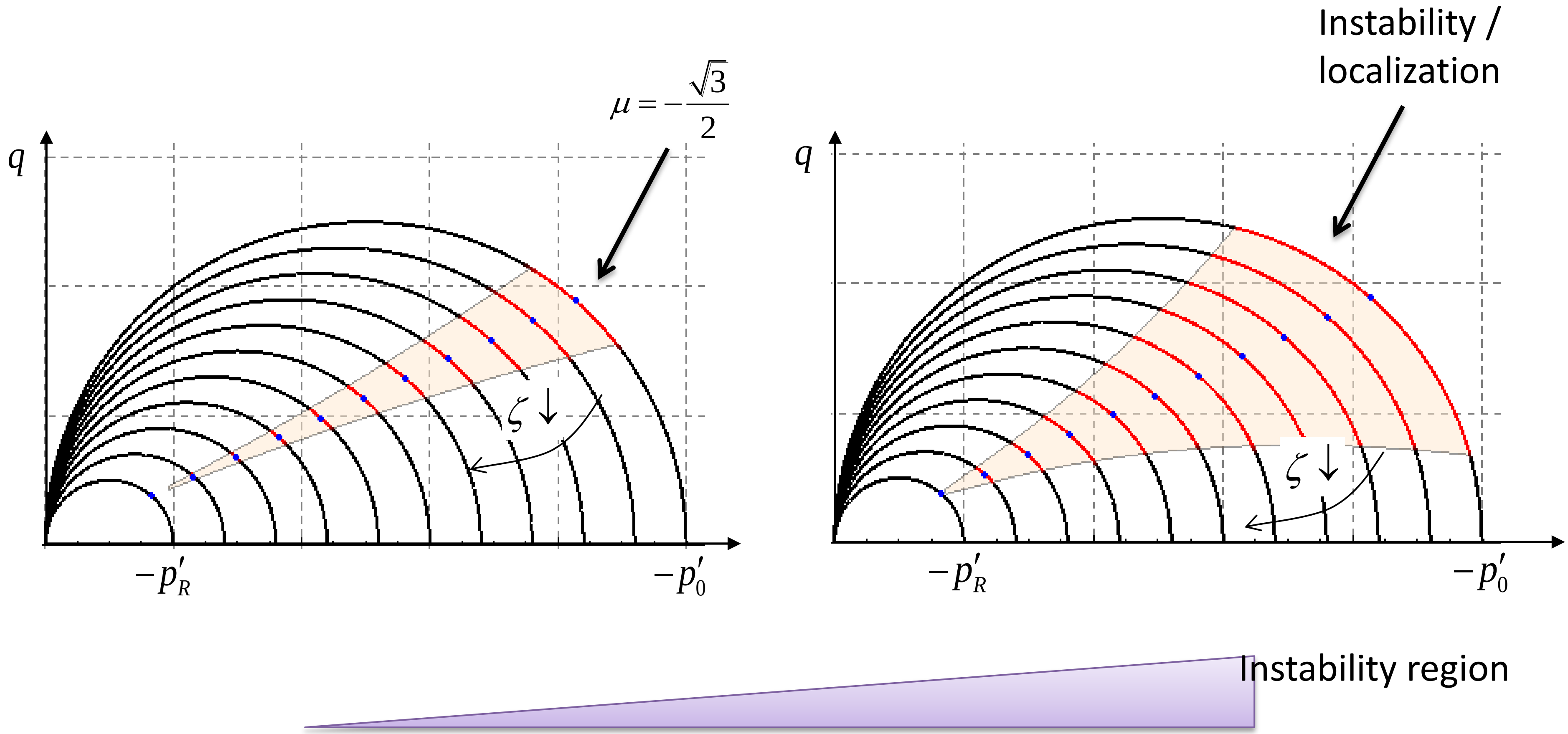
$$p'(z) - \frac{2}{\sqrt{3}} q(z) = \sigma_1 = const$$

Mass balance (macro-scale)

$$\frac{\partial p_f}{\partial t} = c_{hy} \nabla_{\mathbf{x}}^2 p_f - \frac{1}{n\beta_f}\frac{\partial \varepsilon}{\partial t} - \cancel{c_{p,ch}\frac{\partial w_2}{\partial t}}$$

c_{hy}	is the hydraulic diffusivity
n	is the porosity
β_f	is the fluid compressibility
$c_{p,ch}$	is the chemical pressurization coefficient
p_f	is pressure of the fluid
ε	is the volumetric strain

Bifurcation analysis -> Conditions for localization



More crushable solid skeleton ($a^{-1} \Uparrow$)

Compaction banding in a reservoir

Carbonate grainstone

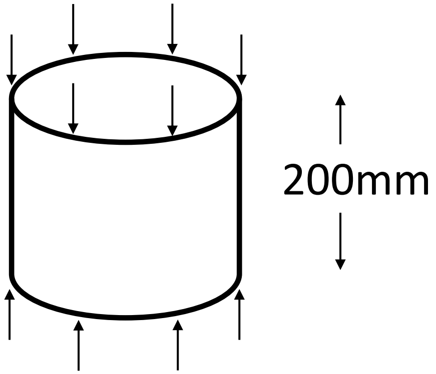
Initial stress state at 1,8km (oedometric)

$$\sigma_V \simeq 45\text{MPa}$$

$$p_f \simeq 18\text{MPa}$$

σ_V=const.

open flow



modeling window

(oedometric conditions)

Elastic constants

$$K = 5\text{GPa}$$

$$G = 5\text{GPa}$$

Cam clay yield surface

$$p'_R = 30\% \, p'_0$$

$$p'_0 = 35\text{MPa}$$

$$M = 0,9$$

Physical properties

$$c_{hy} = 10^{-3}\,\text{m}^2\,\text{s}^{-1}$$

$$D_0^{50} = 0,2\text{mm}$$

$$n = 25\%$$

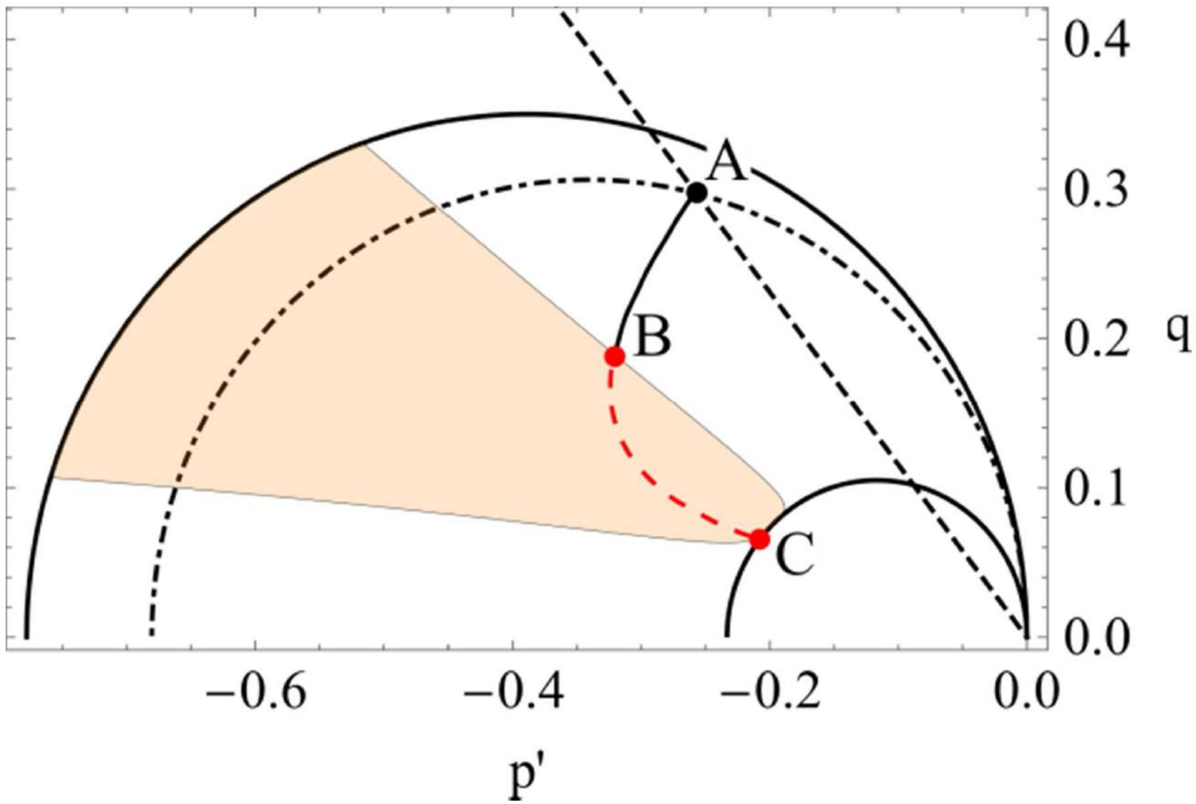
Chemical parameters

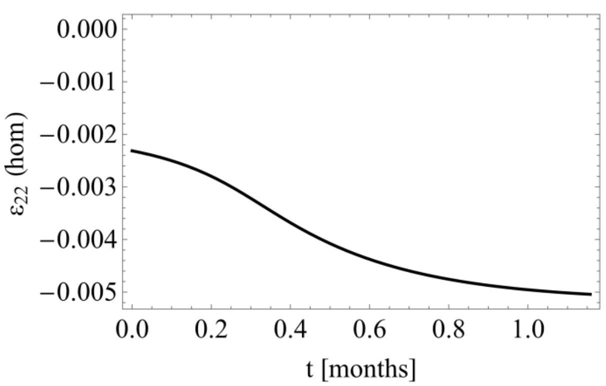
$$k^* = 1,6\,10^{-10}\,\text{m}\,\text{s}^{-1}$$

$$\kappa = 2$$

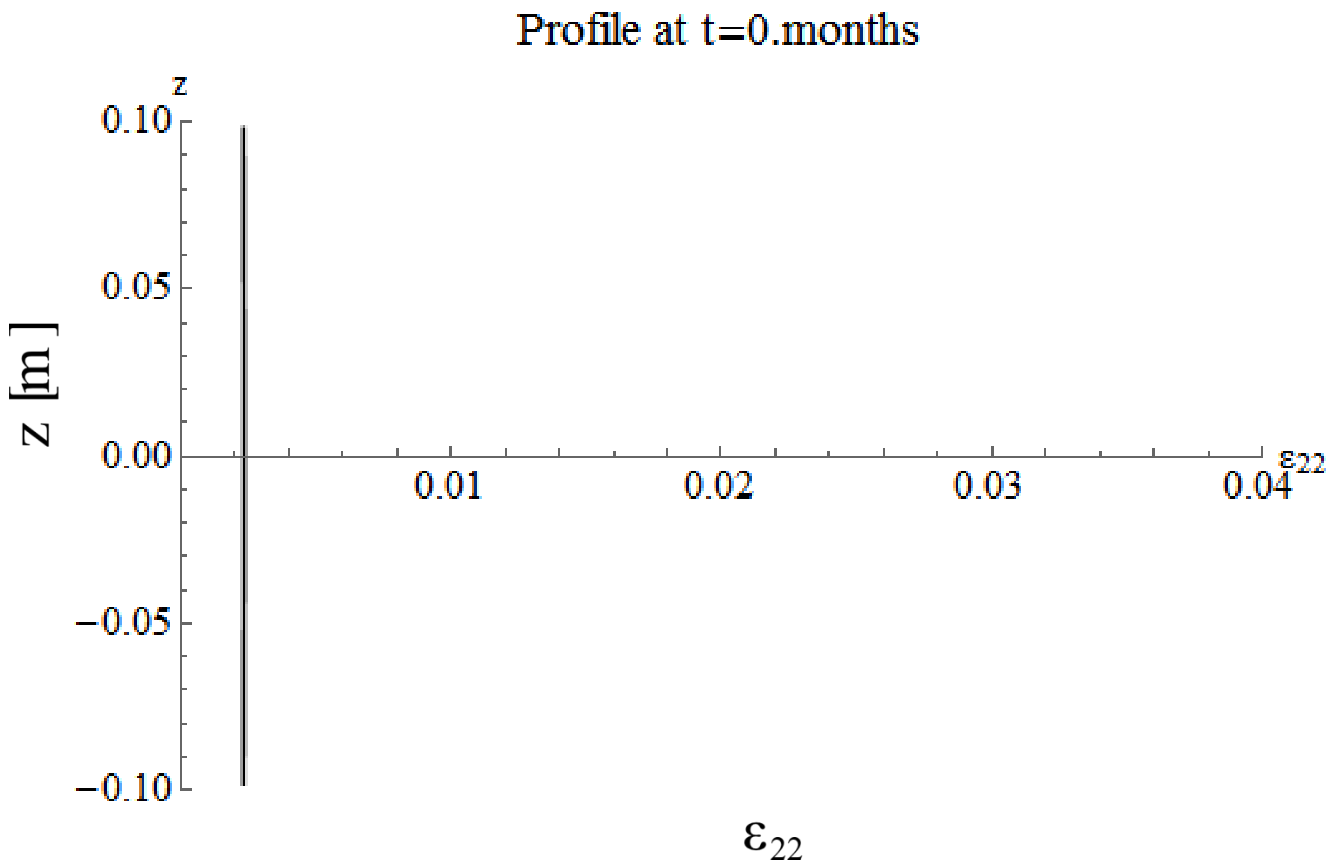
Grain crushing parameter: $a = 1\,\text{MPa}$

Homogeneous deformation under open flow conditions



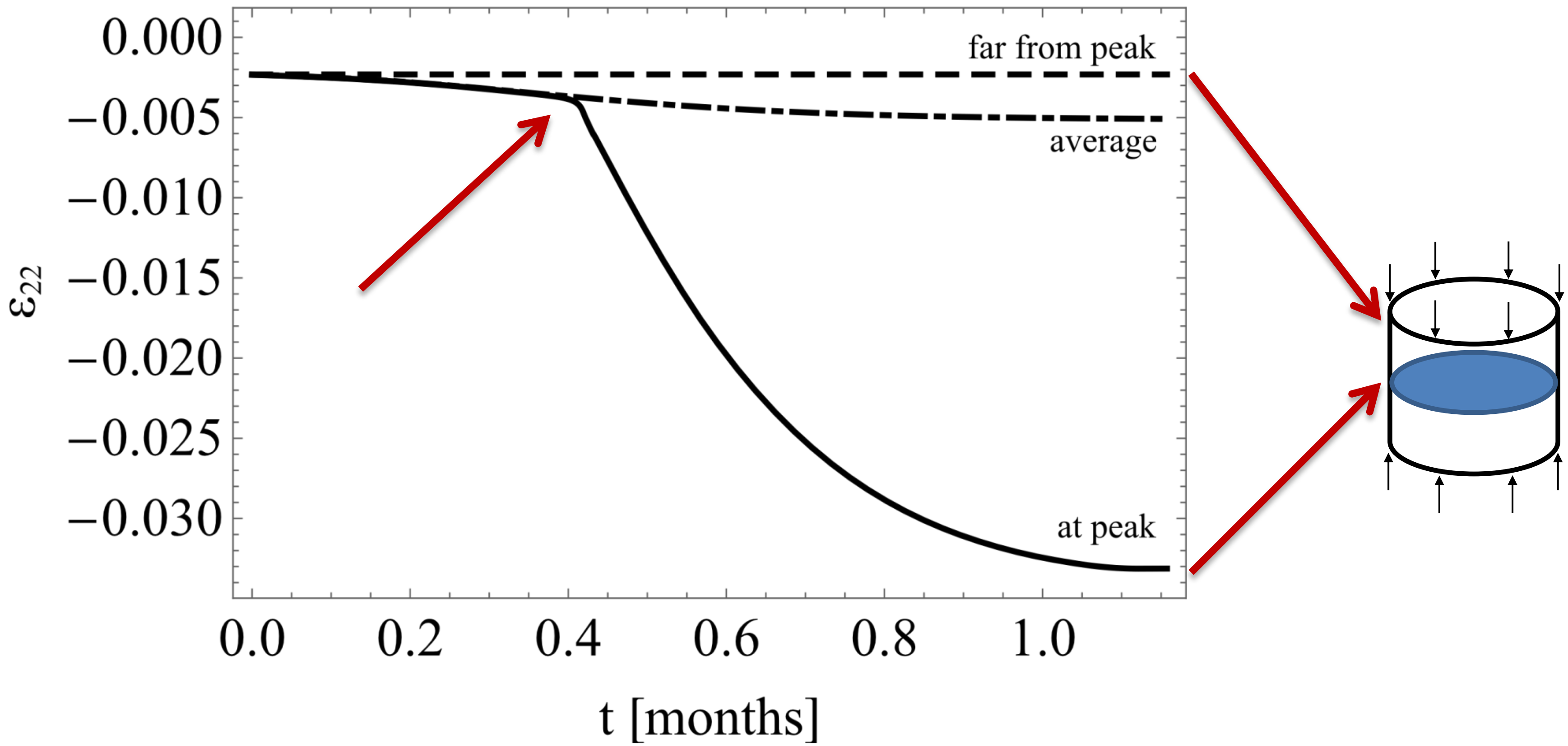


Localization – compaction banding



Sulem, J., & Stefanou I., 2016, Thermal and chemical effects in shear and compaction bands, Geomechanics for Energy and the Environment

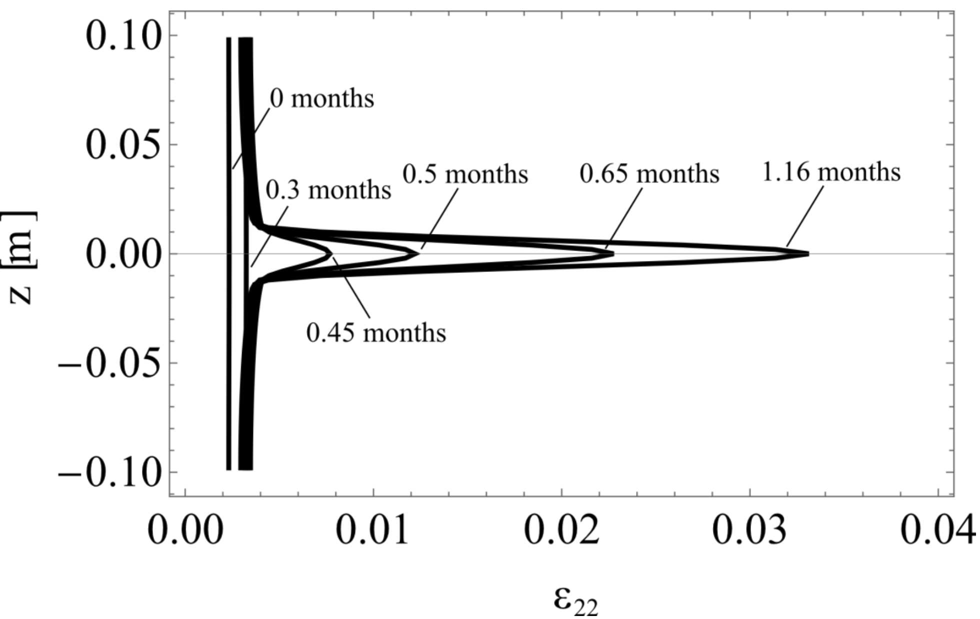
Localization – compaction banding



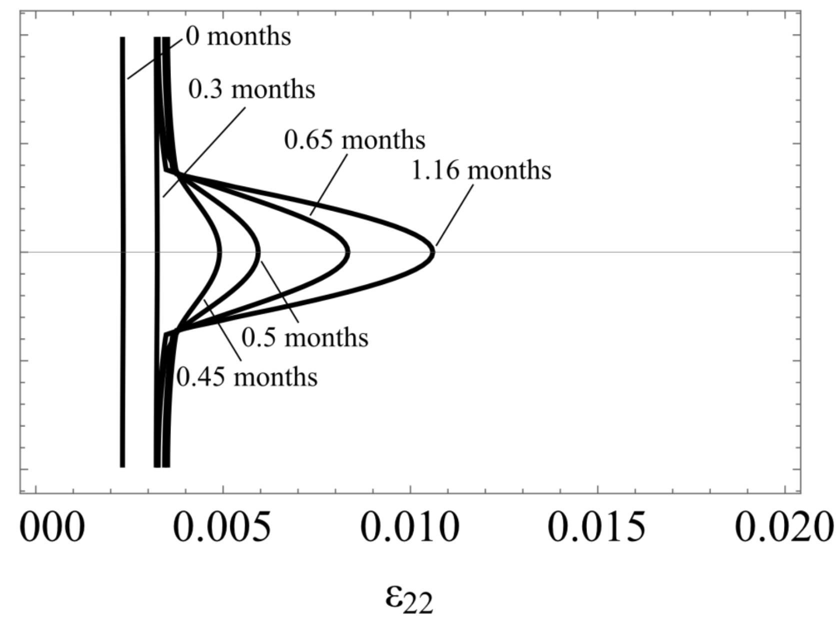
5-10 times more compaction than the average inside the CB

Effect of chemical heterogeneity

ℓ_c =4mm (20 grains)



ℓ_c =16mm (80 grains)



CONCLUSIONS

- ✓ Strain localization is favored by strength softening. Softening can be due to mechanical/thermal degradation (softening of the yield surface) and/or softening due to THMC couplings (thermal/chemical pressurization of the pore fluid)
- ✓ Finite thickness of the localized zone can be obtain for rate dependent constitutive law, higher order continua or non local constitutive models
- ✓ Heat and fluid diffusion tend to stabilize the system and broaden the localized zone whereas fluid pressurization tends to destabilize it.
- ✓ Stronger weakening leads to narrower localized zone.
- ✓ The actual thickness of the localized zone plays a major role in the energy budget of the system: control of the feedback of the dissipative terms (e.g. frictional heating)
- ✓ Chemo-mechanical couplings can induce instabilities of various forms (shear localization, compaction banding…)