Geothermics

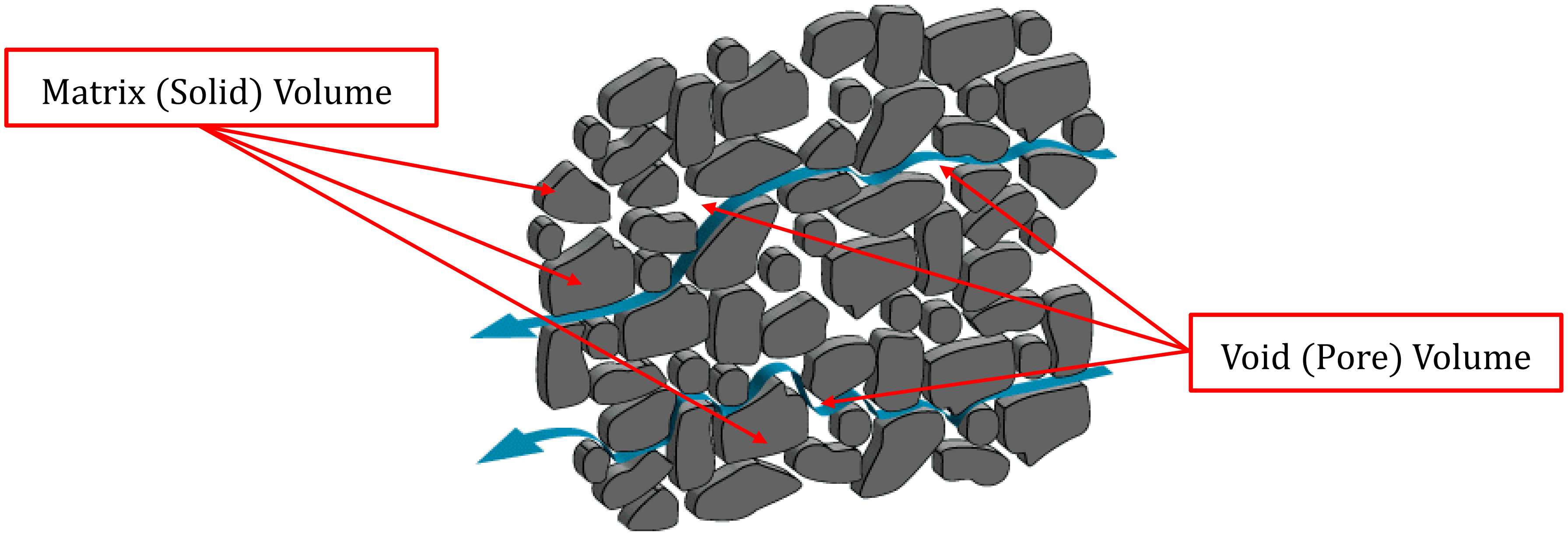
Course Outline:

1. Thermal conditions of the early Earth and present-day Earth’s structure
2. Thermal parameters of the rocks
3. Thermal structure of the lithospheric continental areas (steady state)
4. Thermal structure of the lithospheric oceanic areas
5. Thermal structure of the lithosphere for transient conditions in various tectonic settings
6. Heat balance of the Earth
7. Thermal structure of the sedimentary basins
8. Thermal maturity of sediments
9. Mantle convection and hot spots
10. Magmatic processes and volcanoes
11. **Heat transfer in hydrogeological settings**

Porous Media

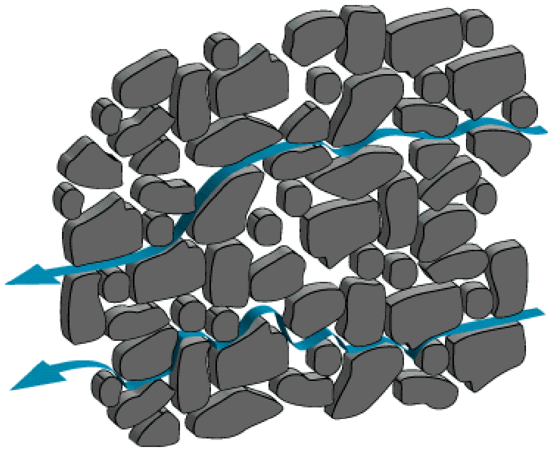
Porous Media is a portion of space occupied by heterogeneous matter (Solid Matrix, Gas and/or Liquid and Void Space)

NB: a minimum number of voids must be interconnected, in order to allow for the fluid phases to move !



Hornberger et al. (1998)

Fluid Flow in sedimentary basins



Hornberger et al. (1998)

Pore space (φ) is mostly filled with water

- Meteoric water: groundwater originated from rainwater
- Connate water: Seawater buried with sediments
- Juvenile water: hydrothermal water by igneous activity
- Formation water: produced during oil or geothermal production

Porosity

- Consists of rock and pore space (void space)

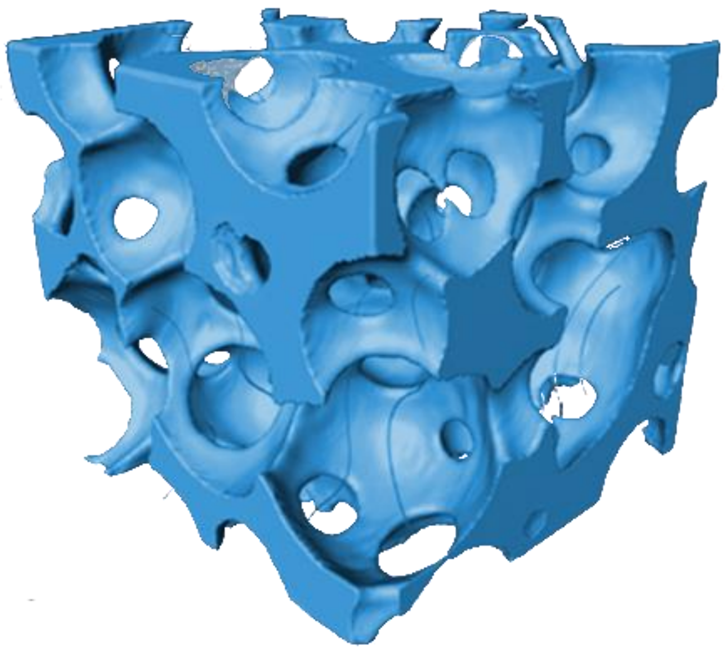
Total porosity

$$\frac{V_{void}}{V_{total}}$$

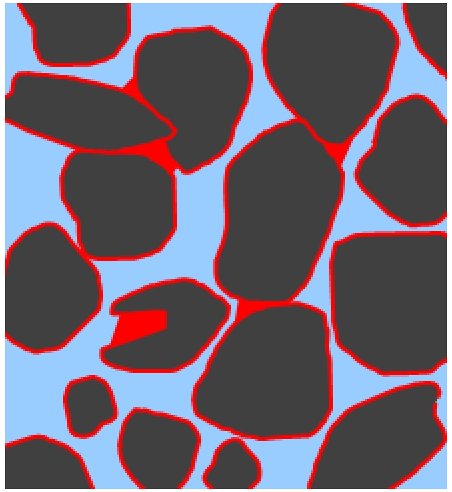
$$\varphi + c = 1$$

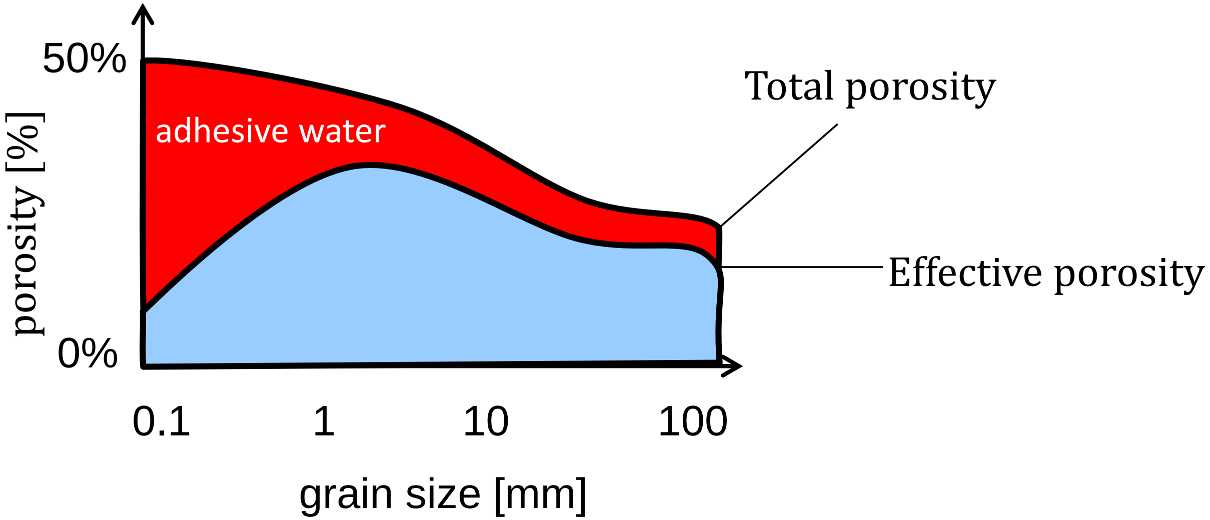
Matrix Volume fraction

$$\frac{V_{pore}}{V_{total}}$$



Effective Porosity



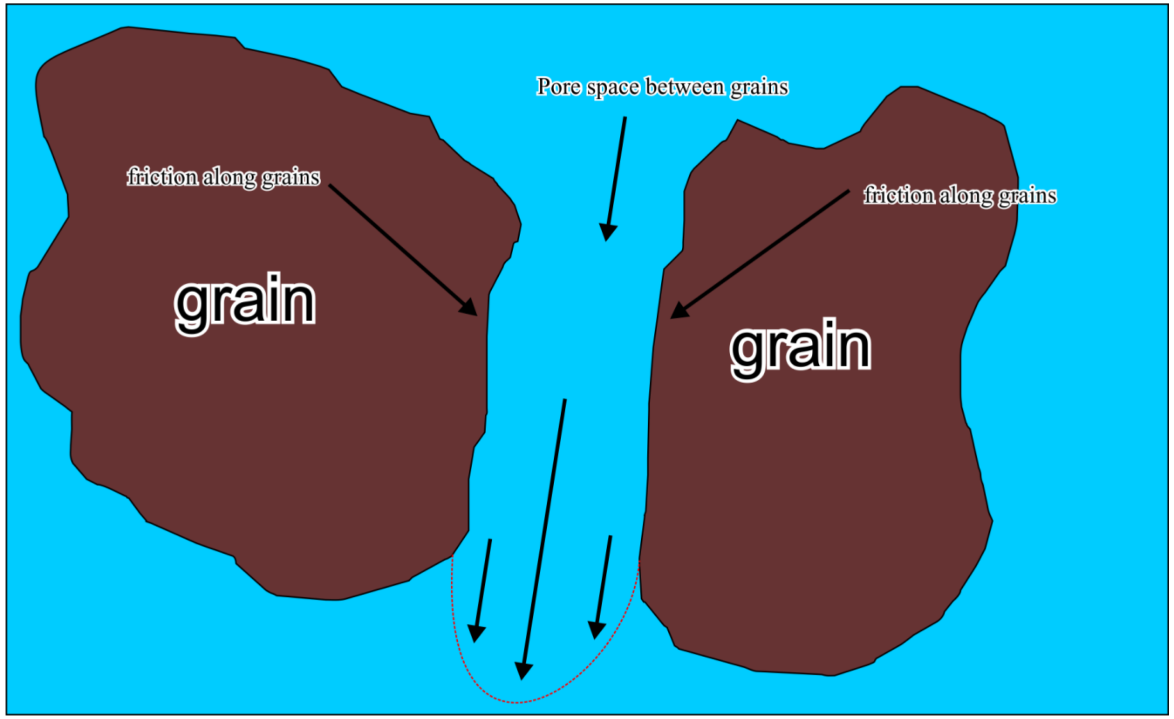


Effective porosity

$$\varphi_{eff} = \varphi_{tot} - \varphi_{adh}$$

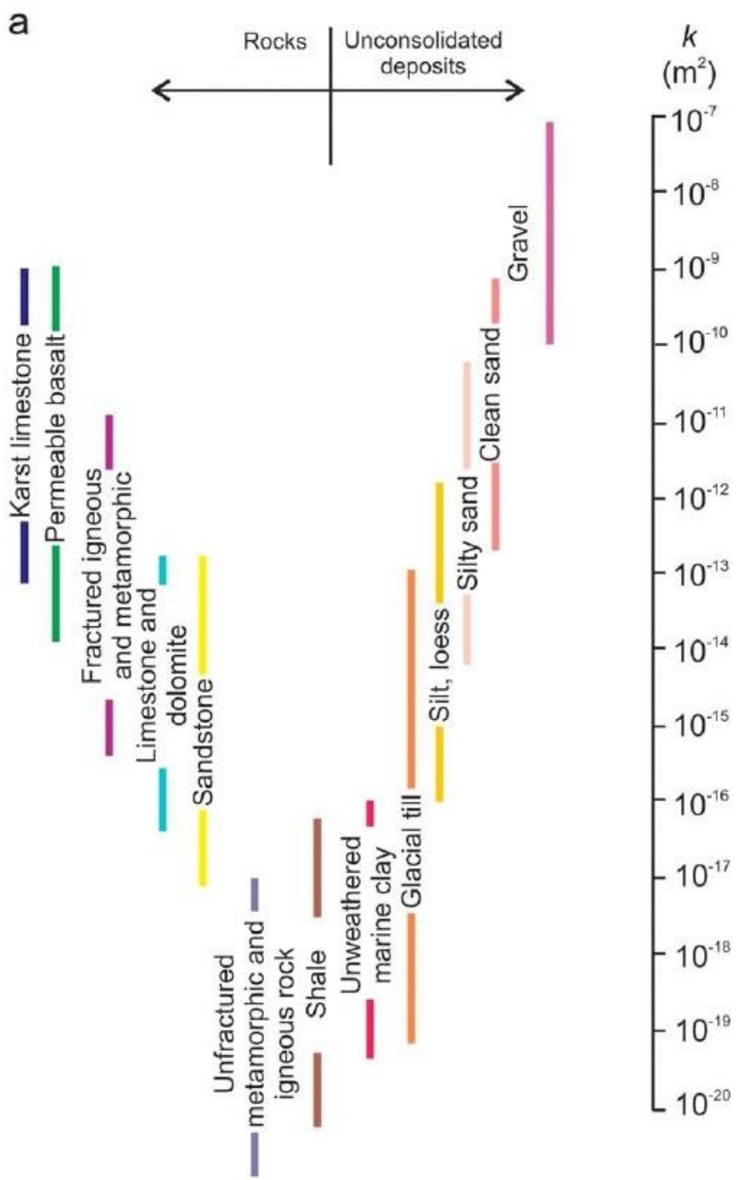
Permeability

Friction Loss along the walls of the grains

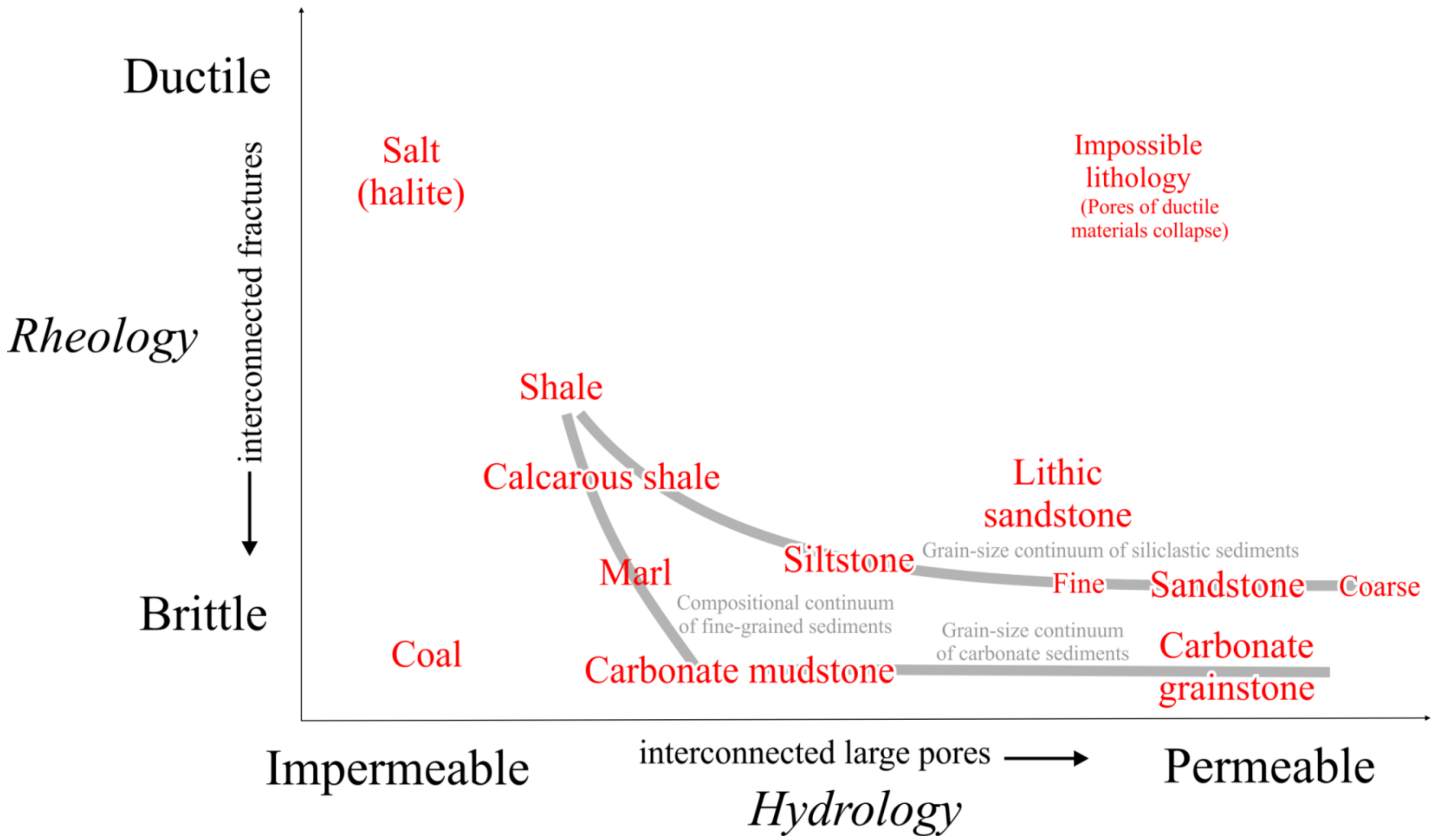


Smaller pores → more frictional resistance → lower permeability

Permeability and Rheology



Gleeson et al. 2011, JGR, 38



Permeability

K=hydraulic conductivity

$$K = \frac{k}{\mu}(\rho g), \text{SI } [\mathrm{m}s^{-1}] \text{ darcy} = \ 10^{-12}\ \mathrm{m}s^{-1}$$

$$\left.\mathrm{k} = \text{intrinsic permeability, SI } [\mathrm{m}^2]\right\} \text{Porous medium property}$$

$$\left.\begin{array}{l}\mu = \text{dynamic viscosity, SI [Pas]} \\ \rho = \text{fluid density, SI } [\mathrm{kg}m^{-3}]\end{array}\right\} \text{Fluid properties}$$

$$g = \text{gravitational constant, SI } [ms^{-2}]$$

- A combined property of the medium and the fluid
- Ease with which fluid moves though the medium

Fluid Flow in sedimentary basins

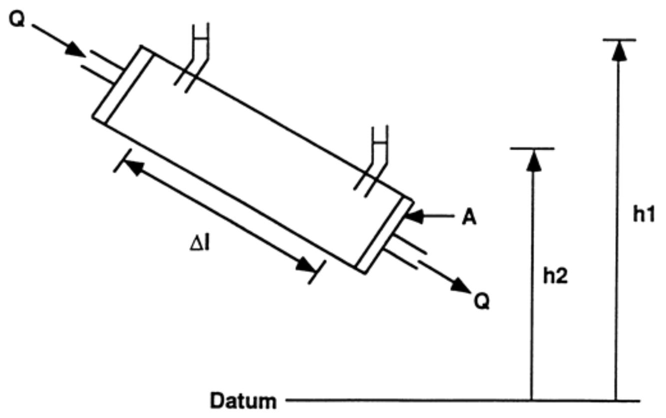
Because …

Groundwater flow in sedimentary basins follows Darcy's law

$$v_f = -K\,\nabla h$$

- Flux of water through a permeable formation (v_f)is proportional to the head loss (∇h = distance between top and bottom of the soil column)

- The constant of proportionality is called hydraulic conductivity (K, SI $[ms^{-1}]$)

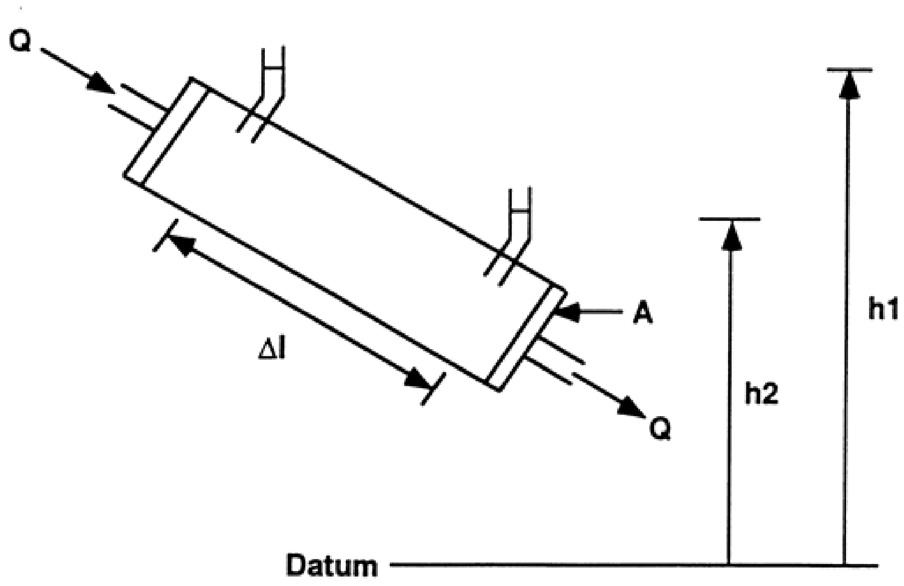


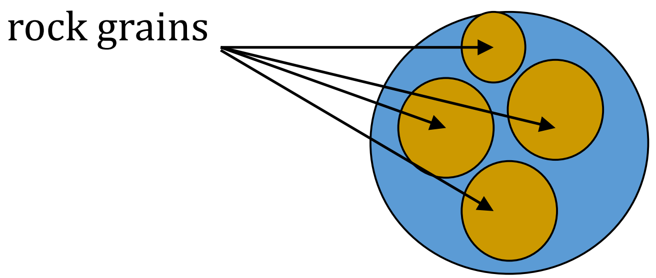
Fluid Flow in sedimentary basins

Darcy velocity

$$v_f = -K\,\nabla h$$

- Macroscopic concept (easily measured)
- Flow rate per unit of cross sectional Area ($v_f = \frac{Q}{A}$)
- Different from the real microscopic velocities
- Microscopic velocities are real, they are impossible to measure





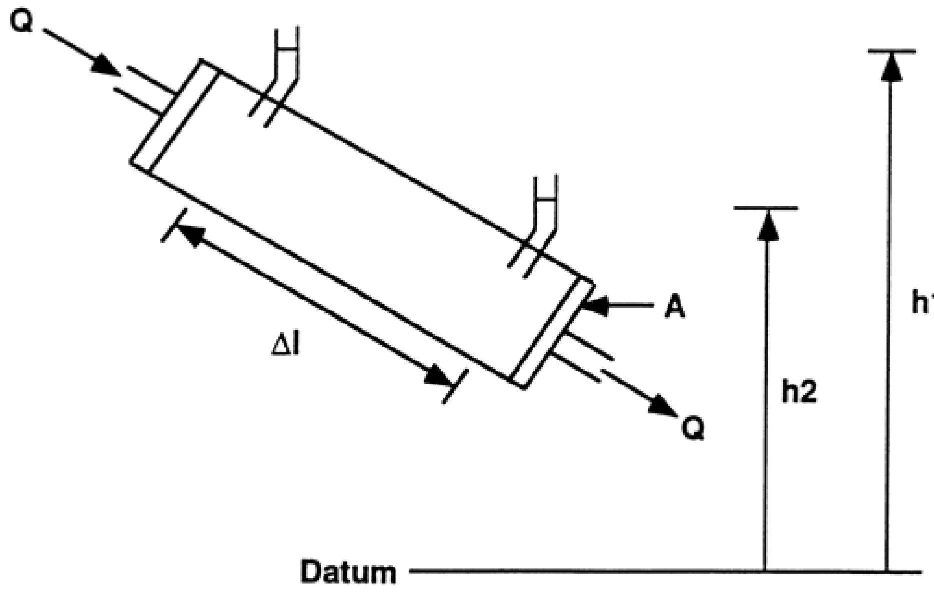
Seepage velocity [v_s]

Fluid Flow in sedimentary basins

Seepage velocity [v_s]

$$Q = A \cdot v_f = A_v \cdot v_s$$

Q = flow rate
A = total cross sectional area of the material
A_v = cross sectional area of the voids
v_f = Darcy velocity
v_s = seepage velocity



$$v_s = v_f \left(\frac{A}{A_v}\right)$$

$$v_s = v_f \left(\frac{A \cdot \Delta l}{A_v \cdot \Delta l}\right) = v_f \left(\frac{V}{V_v}\right)$$

$$v_s = v_f \cdot \varphi$$

Principle of Bernoulli

Bernoulli's principle: for an ideal fluid flow, an increase in the speed of a fluid occurs simultaneously with a decrease in pressure or a decrease in the fluid's potential energy

$$\frac{\sigma v^2}{2} + \sigma g z + P = const. \qquad \frac{\sigma v^2}{2} = P_D$$

v=fluid flow velocity at a point, ***g***=gravity acceleration **z** =elevation of the point above a reference plane, ***P*** =static pressure at the point, ***σ***=density of the fluid, P_D = dynamic pressure, and ***σgz*** = pressure formed by the elevation of the point above a reference plane

$$h = z + \frac{P}{\sigma g}$$

h=piezometric (or hydraulic) head

$$P_T = P + P_D$$

P_T=total pressure

$$h_T = h + v^2/2g$$

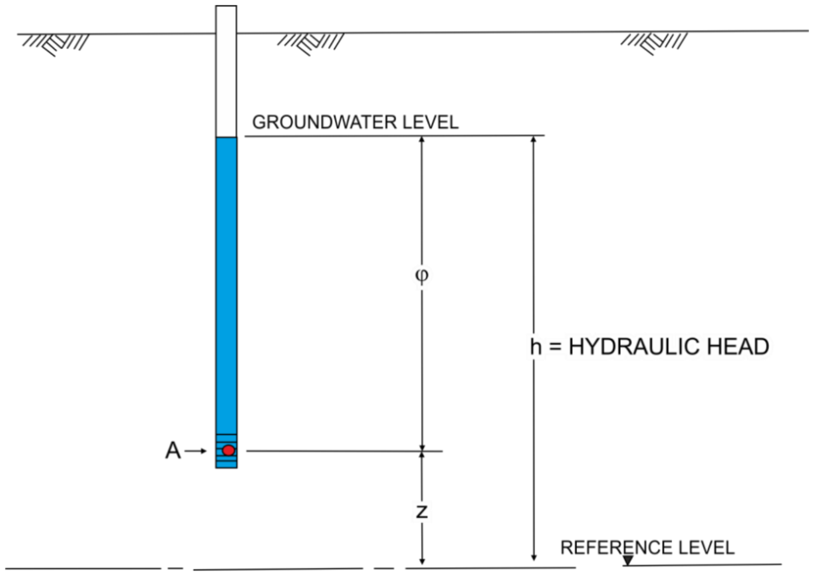
h_T=dynamic head

$$\frac{P}{\sigma g} = \text{ pressure head}$$

$$v^2/2g = \text{velocity head}$$

I=Hydraulic Gradient

$$I = \frac{dh}{dl} = \frac{\Delta h}{\Delta l}.$$



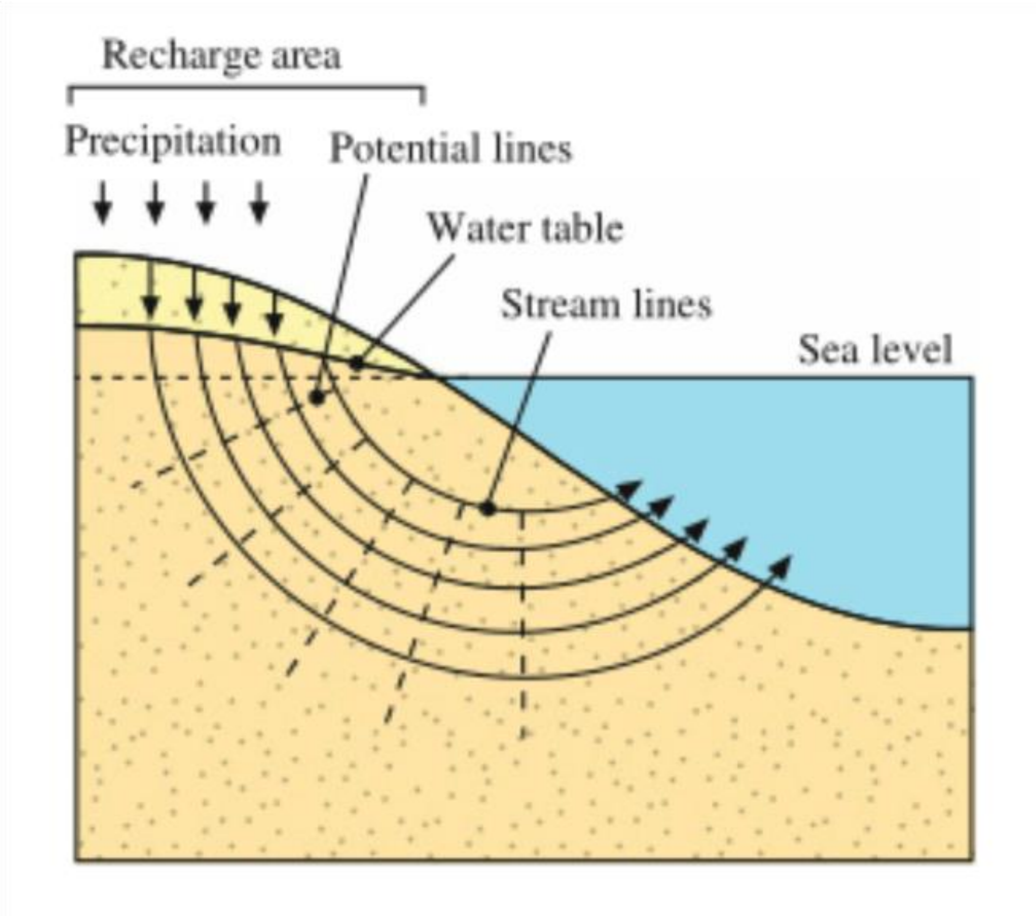
Δh = h_2-h_1 = difference between hydraulic heads measured at two different points *Δl* = distance between the points at which hydraulic heads were measured

Water in the underground

- An aquifer (saturated zone) is an underground layer of water-bearing permeable rock.
- A typical hydrological system on land is composed of an unsaturated zone, an unconfined aquifer (between the water table and the first confining bed or aquitard), and confining aquifers separated by aquitards.
- Water has an extremely high heat capacity (C_{pw} = 4186 $Jkg^{-1}K^{-1}$) compared to rocks (C_{pRocks} ~ 1000 $Jkg^{-1}K^{-1}$) and the constant movement of groundwater favours a fast heat transfer.
- Groundwater flow involves three major processes: groundwater recharge, groundwater flow through permeable aquifers, and groundwater discharge.
- Water in underground storage is usually divided in the part that drains under the influence of gravity (**specific yield**), and the part that is retained as a film on rock surfaces and in very small pores (**specific retention**).
- Specific yield shows how much water is actually available, and specific retention shows how much water remains in the rock after it is drained by gravity.

Fluid Flow in sedimentary basins

Porewater flow is oriented perpendicular to isopotential lines (steepest hydraulic gradients)

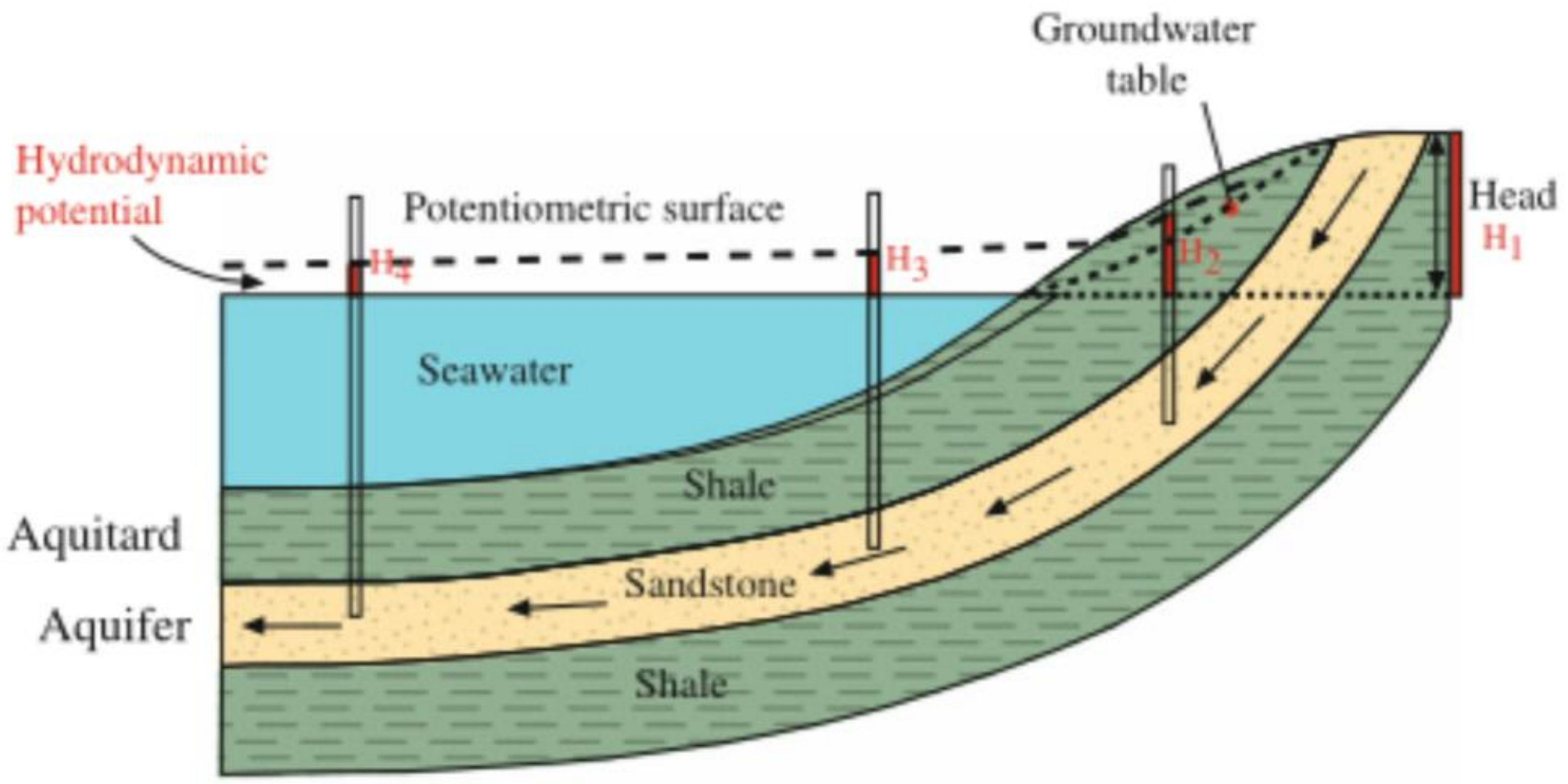


Only for homogeneous sediments

Bjørlykke (1994)

Fluid Flow in sedimentary basins

But sediments are heterogeneous, groundwater flow is highly compartmentalized …



Bjørlykke (1994)

Fluid Flow in sedimentary basins

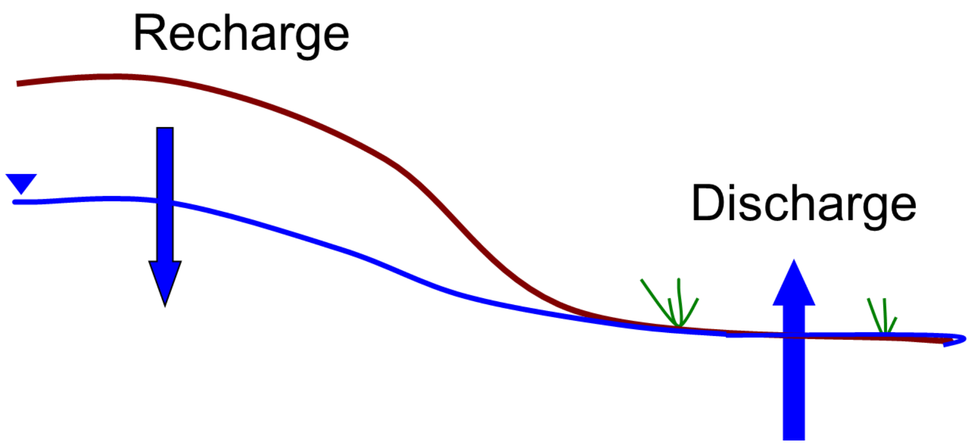
Darcy's law: $v_f = -K\,\nabla h$

- provides an accurate description of the flow of groundwater in almost all hydrogeologic environments

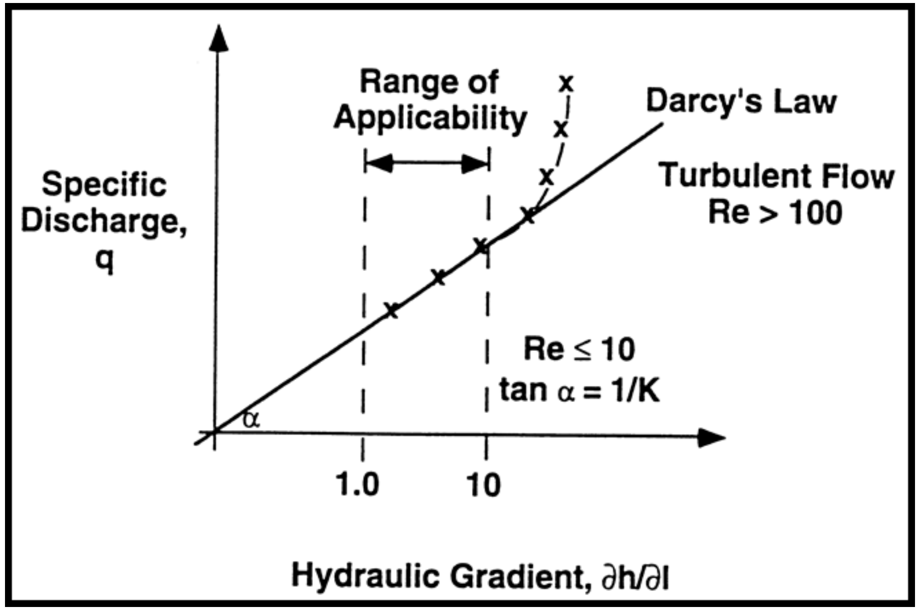
- allows an estimate of:
  - velocity (flow rate) moving within the aquifer
  - average time of travel from the head of the aquifer to a point located downstream

Equations of groundwater flow

Darcy's law $(v_f = -K\,\nabla h)$ + Conservation of fluid mass (Inflow = Outflow)



Fluid Flow in sedimentary basins



- For Re > 10 or where the flow is turbulent (pumped wells)
- Water flows through extremely fine-grained materials (colloidal clay)

Goundwater recharge/discharge

- Groundwater recharge is a hydrologic process where water infiltrates from the surface into underground horizons to replenish aquifers with water.
- Recharge of the saturated zone occurs by percolation of water from the land surface through the unsaturated zone and occurs both naturally (by natural precipitation, rivers and lakes) and artificially (by human activities, such as borehole injection, artificial ponds).
- Groundwater recharge areas are at local heights and discharge areas are at local depressions, since these processes are driven by gravity.
- The rate of groundwater recharge that occurs in any particular area depends on different factors, such as the climate, topography, superficial geology, the kind of aquifer of the area.

Total groundwater recharge

$$R_T = \sum_i R_i.$$

The recharge depends mainly on precipitation and thus is intermittent, while the discharge is a continual process.

Goundwater recharge/discharge

- Different methods are used to estimate the recharge rate: lysimeter method, soil water budget models, the water table fluctuation method, numerical modeling of the unsaturated zone, the zero flux plane method, the Darcy method, the tritium profiling method, the chloride profiling method, etc.
- An empirical relationship for recharge as a function of annual precipitation is:

$$R = 2.0(\mathrm{Pr} - 15)^{0.4}$$ (Kumar and Seethapathi 2002)

R = net recharge due to precipitation during the year (inch), ***Pr*** = annual precipitation (inch)

- If chloride is present in an unsaturated zone as a result of atmospheric deposition, and in the absence of other sources or sinks of chloride ions in the unsaturated zone, the recharge rate (*R*):

$$R = \frac{C_{ic}}{C_Z}\mathrm{Pr}$$

C_Z = mean chloride ion concentration in soil water (in mg/l), ***Pr*** =precipitation (in mm/year), C_{ic} = chloride ion concentration in rainfall (in mg/l).

Trasmissivity

Transmissivity is the rate at which water is transmitted through a unit width of an aquifer under a unit hydraulic gradient. The transmissivity (*Tr*) of an aquifer is equal to the hydraulic conductivity of the aquifer multiplied by the saturated thickness (b) of the aquifer

$$Tr = \kappa b \qquad Tr = \frac{Q}{W}\left(\frac{dl}{dh}\right)$$

Q = amount of water moving through the width (*W*) of the aquifer

b=satured thickness of the acquifer

κ=hydraulic conductivity

$$\kappa = K\sigma\frac{g}{\mu}$$

K =permeability (in m²), κ =hydraulic conductivity (m/s), μ =dynamic viscosity (Pa s), σ = density of the fluid (kg/m³) and *g* =gravity acceleration (in m/s²).

Discharge

$$q_T = \sum_i q_i$$

Groundwater discharge (*q*) is the flow rate of groundwater through an aquifer: $q = \kappa A \frac{dh}{dl}$

κ = hydraulic conductivity of the aquifer A =area the groundwater is flowing through

Water reservoir

Amount of water stored:

$$\Delta S = \sum_i R_i - \sum_i q_i$$

Only the part of aquifer porous water that will drain under the influence of gravity (usable volume capacity V_u):

$$V_U = K_{OP} V_0$$

V_0 = total volume of aquifer and K_{OP} = effective porosity of rock comprising the aquifer (e.g., 23 % for an unconfined acquifer)

- **Underground water can also be contained in reservoirs confined in fields at much greater depths. This water usually does not move, but can absorb significant amounts of heat due to the very high heat capacity of water.**

To estimate the extra-heat energy absorbed by water, if the initial volume of water is V_0 and the initial P and T within the reservoir are P_1 and T_1:

$$P_1 = P_0 + \frac{\alpha_W}{\beta_W}(T_1 - T_0) - \frac{1}{\beta_W}\frac{\Delta V}{V_0}$$

α_w= coefficient of the thermal expansion β_w= volume and compressibility of water, P_0 and T_0 = normal pressure and temperature of water in the reservoir, ΔV = change in the initial volume V_0.

Water and oil reservoir

If some water V_1 is pumped out of the reservoir at the pressure and temperature within the reservoir, the rest of water and the pressure within the reservoir will drop to P_2, since the water will take the initial volume of water within the reservoir V_0:

$$P_2 - P_1 = -\frac{1}{\beta_W}\frac{V_1}{V_0} \qquad V_0 = -\frac{1}{\beta_W}\frac{V_1}{(P_2 - P_1)}$$

(change of volume = V_1, change of temperature is negligible)

If the reservoir is filled with water and oil (P_1 and T_1= initial pressure and temperature):

$$V_0 = V_{W0} + V_{N0}$$

V_{W0} = initial volumes of water V_{N0} = initial volumes

If some water/oil V_1 is pumped out of the reservoir:

Change of volume

$$V_1 = \Delta V_{W1} + \Delta V_{N1} \qquad \Delta V_{N1} = V_1 - \Delta V_{W1}$$

Change of pressure

$$P_2 = P_1 + \frac{\alpha_W}{\beta_W}(T_2 - T_1) - \frac{1}{\beta_W}\frac{\Delta V_{W1}}{V_{W0}} \qquad P_2 = P_1 + \frac{\alpha_N}{\beta_N}(T_2 - T_1) - \frac{1}{\beta_N}\frac{V_1 - \Delta V_{W1}}{V_{N0}}$$

Heat in groundwater flow

Groundwater flow can be described by Darcy's law, in which the velocity of water *u* is connected with its pressure in the form.

$$u = -\frac{\kappa_p}{\eta_w} \nabla (p + \rho_w g z) = -\frac{\kappa_p \rho_w g}{\eta_w} \nabla H_h = -\lambda_c \nabla H_h$$

$H_h = z + p/(\rho_w g)$

u = average velocity per unit area, k_p = permeability, η_w = water dynamic viscosity and ρ_w = water density, p= pressure, H_h is the hydraulic head, z is the elevation above a standard datum, g = gravity acceleration, gH_h = hydraulic potential, and $\lambda_c = k_p \rho_w g/\eta_w$ = hydraulic conductivity

The heat transferred from rock to water depends on *u*, the volumetric heat capacity of water ($\rho_w c_w$) and the thermal gradient in the direction of *u*.

$$\rho c \frac{\partial T}{\partial t} = k \nabla^2 T - \rho_w c_w (\mathrm{u} \cdot \nabla T)$$

ρ_w = water density, c_w = specific heat of water, c = specific heat, ρ = density, k = thermal conductivity of the water–rock matrix, t =time

Heat in groundwater flow
vertical flow

When the flow in an aquifer occurs due to an externally applied pressure gradient, the heat is transferred through a mechanism known as advection. When the flow occurs due to buoyancy effect, the mechanism of heat transfer is the convection.

$$\frac{d^2T}{dz^2} = \frac{\rho_w c_w \kappa_p}{k \eta_w} \left[\frac{d}{dz}(p + \rho_w g z) + g z \frac{d\rho_w}{dz} \right] \frac{dT}{dz}$$

α_w=water thermal expansion coefficient

$$\alpha_w = (d\rho_w/dT)/\rho_w$$

For thermal steady state conditions the conductive heat transfer, heat advection, and thermal convection in a homogenous permeable layer is expressed by:

$$\frac{d^2T}{dz^2} = \frac{\rho_w^2 c_w g \kappa_p}{k \eta_w} \left(\frac{dH_h}{dz} + \alpha_w z \frac{dT}{dz} \right) \frac{dT}{dz}$$

Vertical flow for conductive and advective heat transfer for steady-state thermal conditions and for a uniform, isotropic, homogeneous and satured aquifer:

$$\frac{d^2T}{dz^2} - \frac{c_w \rho_w u_z}{k} \frac{dT}{dz} = 0$$

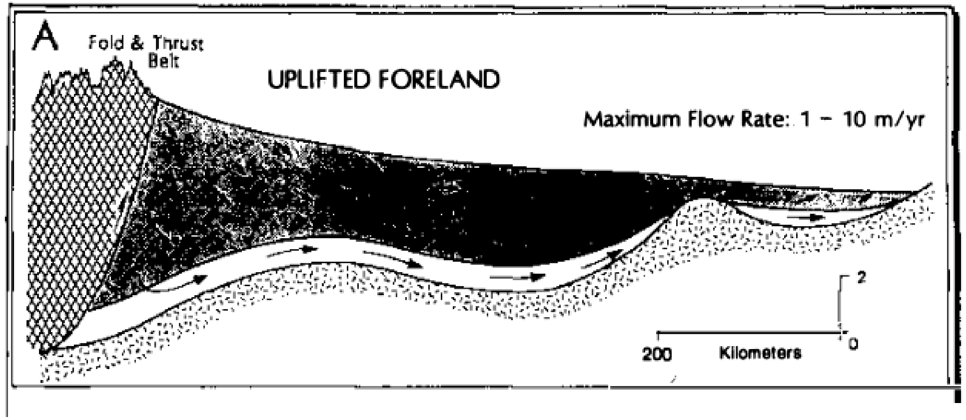
$$T = T_1 + (T_2 - T_1) \frac{\exp(\beta_z z/h) - 1}{\exp(\beta_z) - 1}$$

u_z=Darcy velocity in the vertical direction

h=vertical distance within the aquifer

$$\beta_z = c_w \rho_w u_z h/k$$ (dimensionless parameter)

Fluid Flow in sedimentary basins

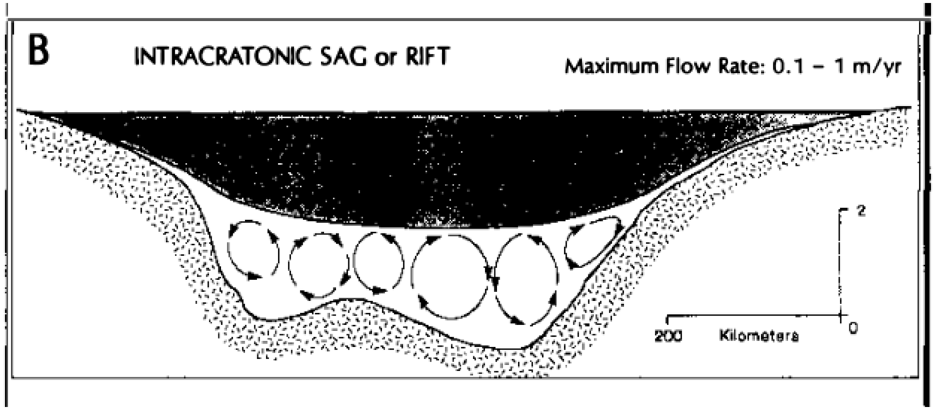


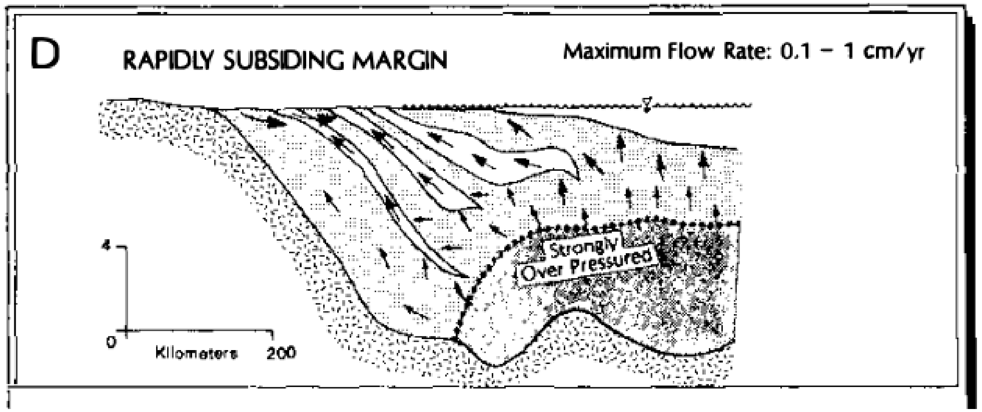
Topographic-driven

Groundwater table drives meteoric water inside a basin

Density-driven (convection)

Subsurface water flow by density gradients in fluid density due to differences in temperature and salinity.



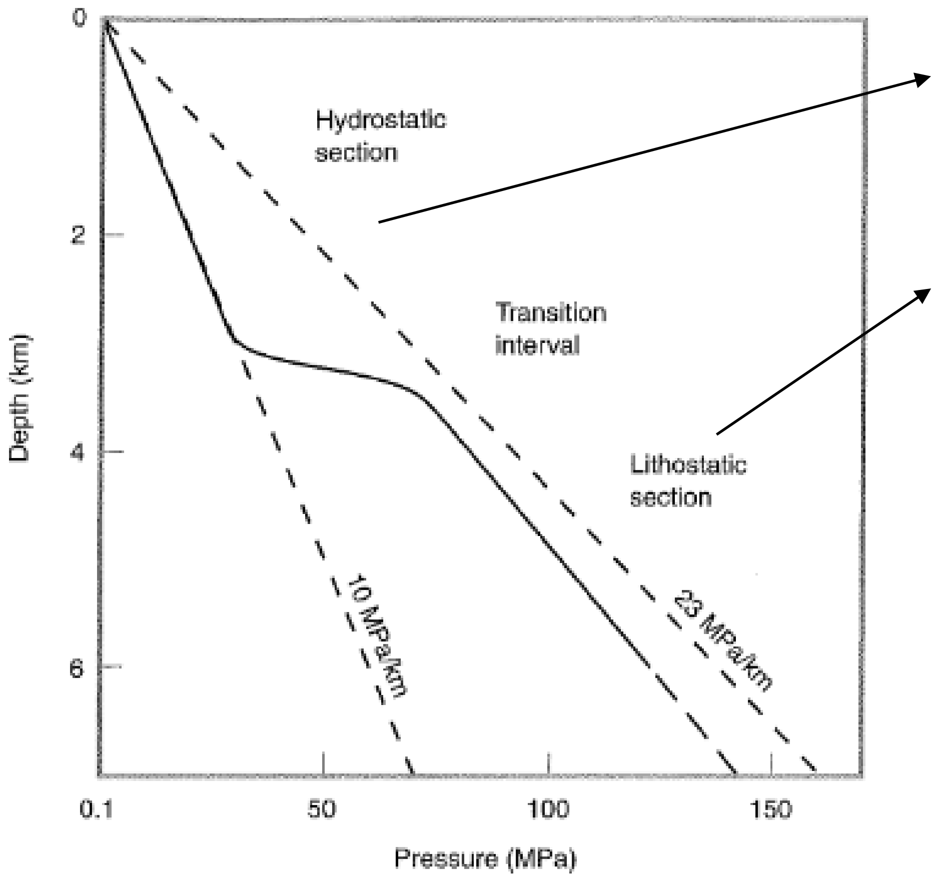


Compaction-driven

Upward pushing of connate water driven by the effective stress and chemical compaction

Compaction-driven groundwater flow

Flow induced by ANOMALOUS PORE PRESSURE



HYDROSTATIC pore pressure
weight of the overlying fluid

LITHOSTATIC pore pressure
weight of the entire burden
(fluid + matrix)

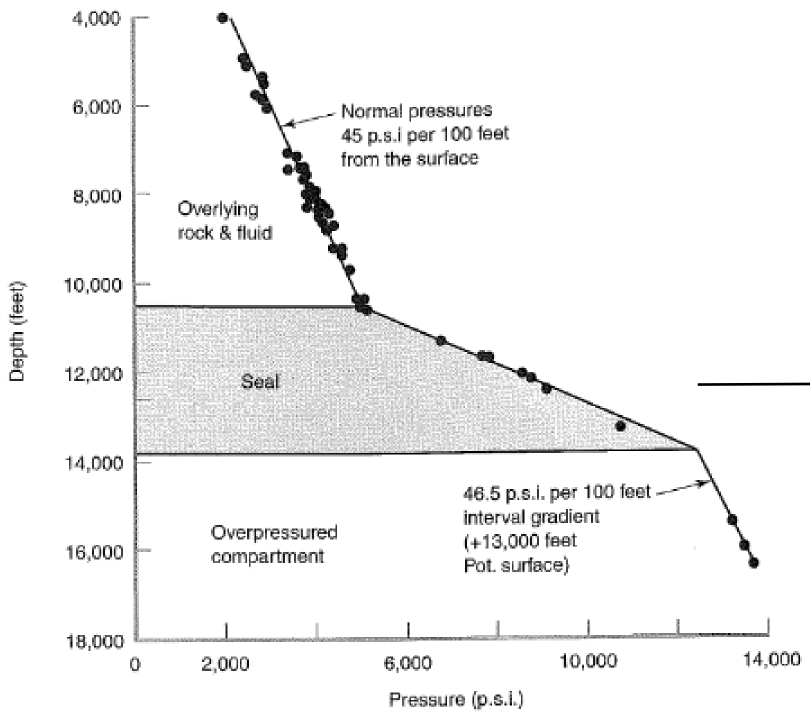
OVERPRESSURE (geopressure) higher than hydrostatic

UNDERPRESSURE lower than hydrostatic

How anomalous pressure are created and maintained ?

Abnormal Pore Pressure

STATIC SCHOOL

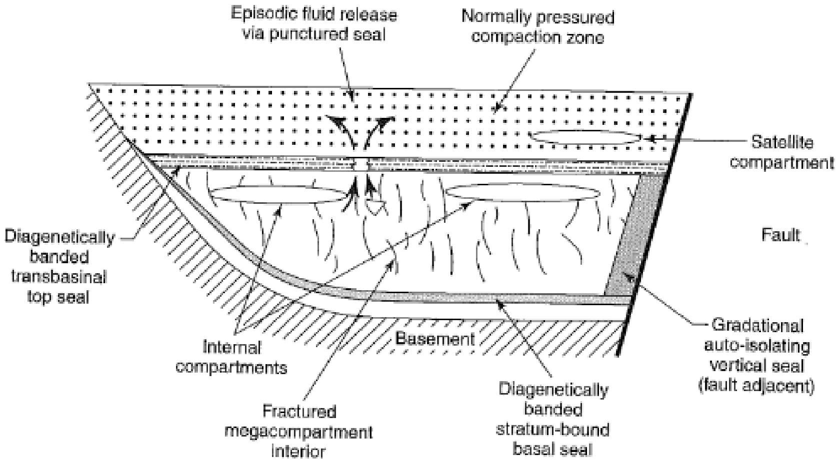


Anomalous pore pressures are maintained by SEALS

" ... a zone of rocks capable of hydraulic sealing, that is, preventing the flow of oil, gas and water. The term does not refer to capillary seals ... the term refer to seals that prevent essentially all pore fluid movement over substantial intervals of geologic time. "

PRESSURE COMPARTMENTS

" A three dimensional hydraulically isolated volume of the Earth's crust that has a fluid pressure different from the ambient surrounding."



Iceland Deep Drilling Project (IDDP)

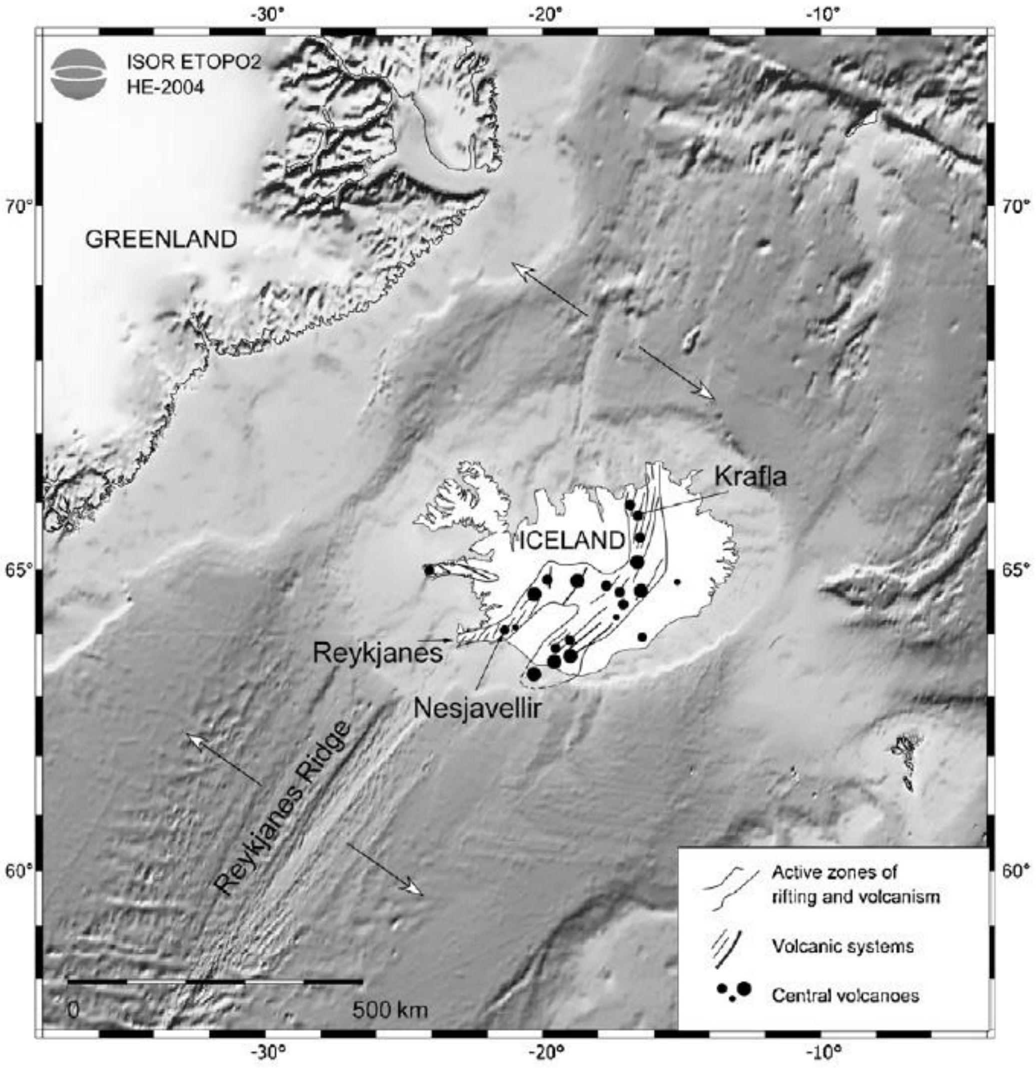
IDDP is a long term program by an industry-government consortium established in 2000, which aims at investigating unconventional, very high-temperature, geothermal systems, to improve the economics of geothermal resources, minimize the environmental impact of harnessing geothermal reservoirs, evaluate the volume of deep accessible geothermal resource.

IDDP aims to investigate the power potential and economics of the *T-P* regime of supercritical fluids:

- Supercritical water has much higher enthalpy and lower viscosity than a two phase mixture of steam and water at subcritical *T* and *P*.
- An aqueous hydrothermal fluid at supercritical conditions with a *T* of 400°C and a *P* of 25MPa has more than five times the power-producing potential of liquid water at a *T* of 225°C (it has a higher thermodynamic efficiency).
- Supercritical conditions can lead to extremely high rates of mass and energy transport and play a major role in high *T* water/rock reaction and the transport of dissolved metals.
- Hydrous fluid systems at supercritical pressures can only be reached at great depths in natural hydrothermal systems than in volcanic complexes (for pure water the critical *P* and *T* are reached at 22.1 MPa and 374°C).

Iceland Deep Drilling Project (IDDP)

Why Iceland?



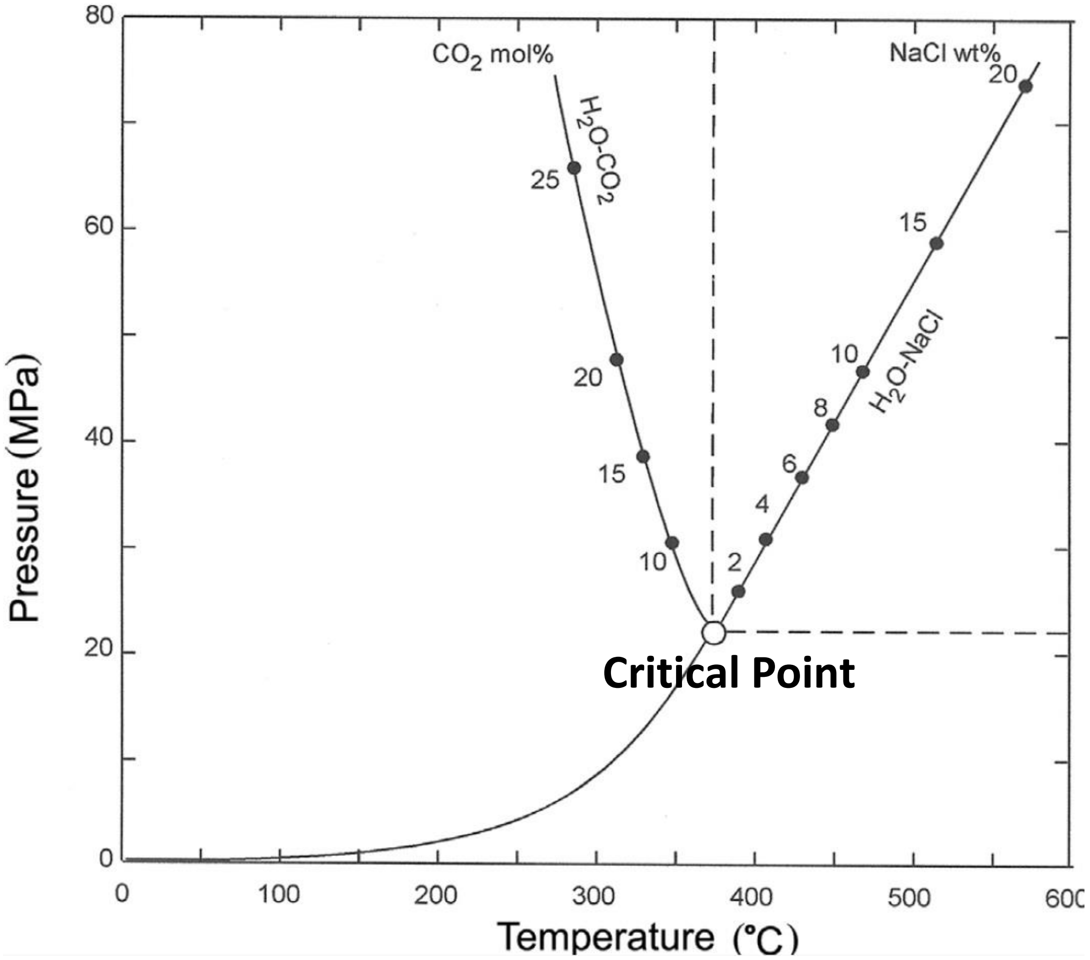
Friðleifsson et al., 2014, Geothermics, 49

- **Iceland leads the world in geothermal development on a per capita basis: Direct geothermal use heats about 90% of its buildings and about 30% of its electrical production is geothermal.**

- **Greater abundance of hydrothermal systems in Iceland relative to their abundance on "typical" mid-ocean ridges are likely because of (1) the high heat flow associated with frequent volcanicity (30 volcanoes erupted in post-glacial time with about 2400 eruptions in the last 11 kyr), related to a hot spot under Iceland, (2) more frequent seismicity, and (3) higher permeability than that of typical oceanic crust.**

Iceland Deep Drilling Project (IDDP)

Boiling point curve and critical point curves for water

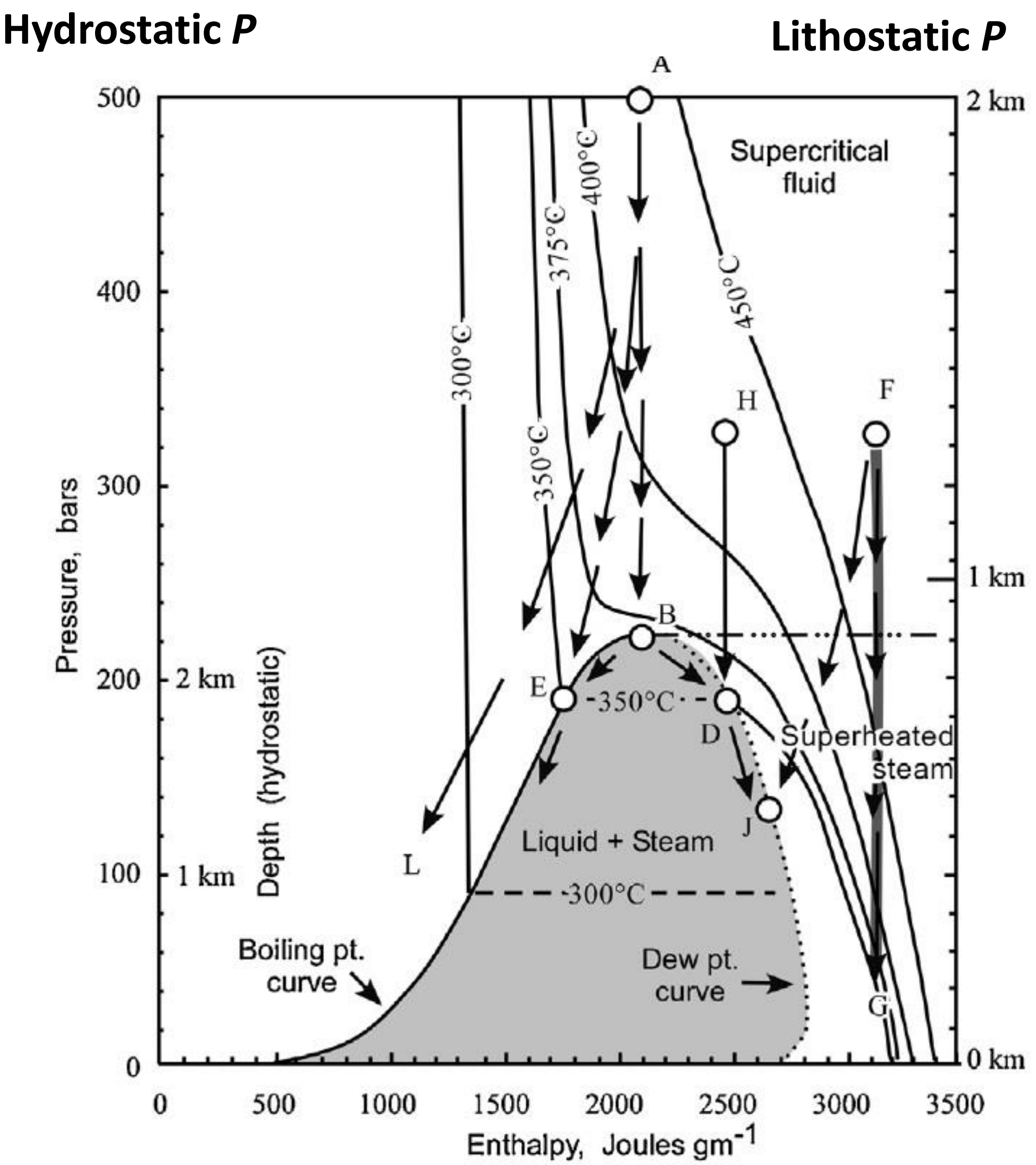


Elders et al., 2014, Geothermics, 49

- The critical *P* in a hot enough reservoir containing pure water would be reached at about 2.3 km depth (22.1 MPa and 374°C), and for fluids with seawater salinity at 3 km depth (29.8 MPa and 407°C).
- Dissolved salt increases the *T* and *P* of the critical point whereas dissolved gas reduces the *T* and elevates the *P* of the critical point.
- Black smokers on mid-ocean rifts occur at depths shallower than the critical pressure of seawater and thus can expel very hot hydrous fluids directly into the ocean without boiling occurring (subcritical conditions).

Iceland Deep Drilling Project (IDDP)

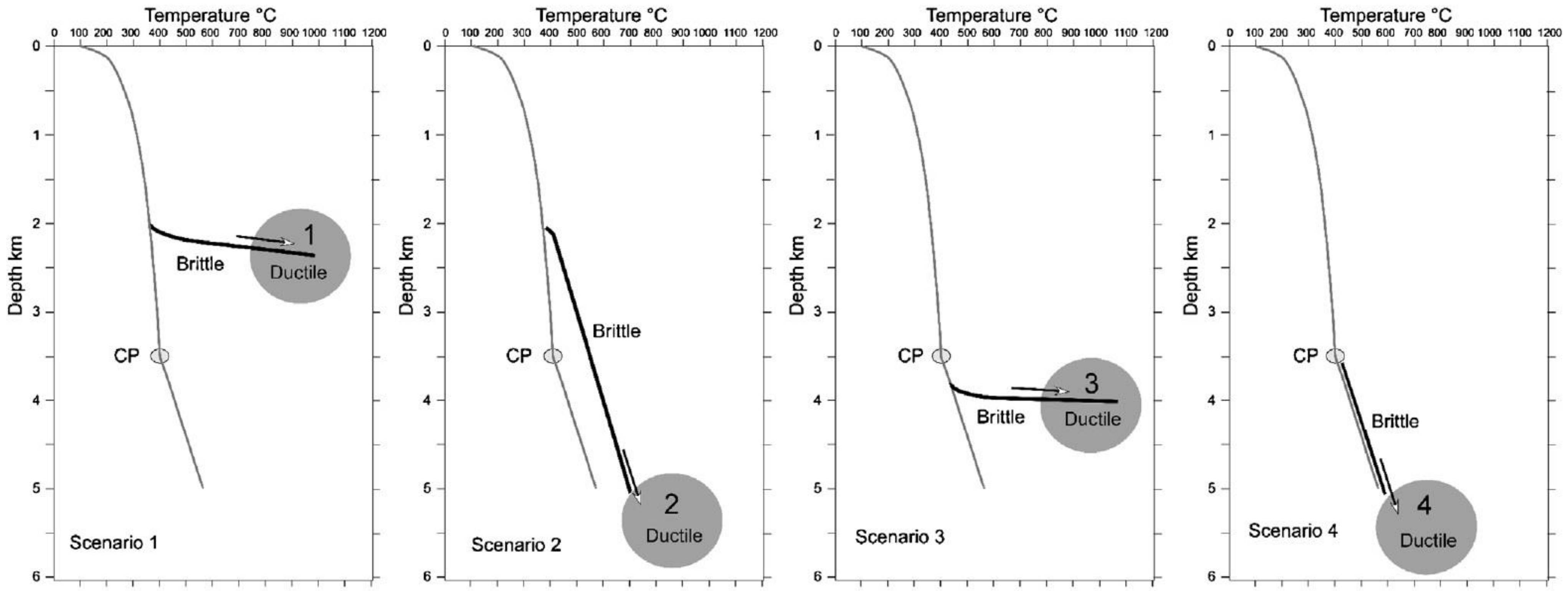
Pressure-enthalpy diagram (pure water, selected isotherms)



- If a hydrothermal fluid (at A) flows upward and decompresses and cools adiabatically it would reach the critical point (at B), and with further decompression separate into two phases, water and steam (E and D).
- The pathway H–D represents supercritical fluid that separates into steam and water at D and E, a situation representative of a vapor-dominated geothermal reservoir.
- IDDP aims to produce supercritical fluids to the surface in such a way that it transitions directly to superheated steam along a path like F-G (in subcritical pressure), resulting in a much greater power output than from a typical geothermal well.

Friðleifsson et al., 2014, Geothermics, 49

Iceland Deep Drilling Project (IDDP)



Friðleifsson et al., 2014, Geothermics, 49

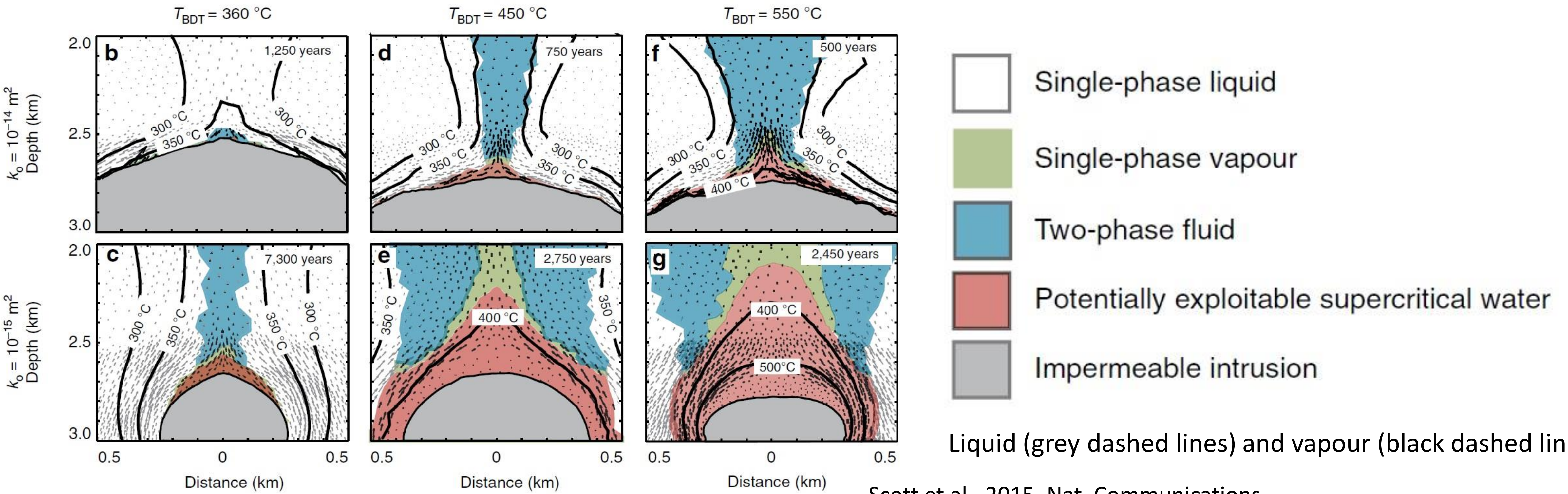
1. *T* path if we drill into magma at ~2 km depth, **2.** *T* path if we drill downwards along a contact aureole of a large magma intrusion, **3.** *T* path if we drill into magma at 4 km depth, **4.** *T* path if we drill beyond the critical point within the amphibolite facies rocks at supercritical *T* (> 400°C).

- Molten or recently crystallized, shallow intrusion responsible for superheated conditions are fairly widespread at Krafla.
- Hydrous phases may exist in the crust at depths where the average *T* exceeds 400°C. Expected *T* at all IDDP drill fields of Iceland, range from 550°C to 650 °C at 5 km.
- Geothermal wells in Iceland today typically reach a depth up to 3.0 km and produce steam up to 340°C, at a rate sufficient to generate about4–10MW of electricity.

Supercritical water resources and geology

- Water enthalpy in conventional high enthalpy geothermal systems depends strongly on the rock permeability. The permeability of volcanic rocks and crystalline basement hosting geothermal systems is in the range of 10^{-14} ('high' permeability) to $10^{-15} m^2$ ('intermediate' permeability).
- High host rock permeability allows rapid fluid advection near the intrusion, resulting in a higher rate of heat transfer from the intrusion to geothermal fluid, albeit with moderate fluid temperatures and enthalpies.
- Intermediate permeability reduces the overall rate of heat transfer but leads to higher water temperatures and enthalpies.
- Below a permeability of $10^{-16} m^2$, heat transport changes from being advection- to conduction-dominated (uneconomic rates of fluid production).
- The key control on the formation of supercritical resources is the brittle–ductile transition temperature T_{BDT}. Extensive supercritical water resources can develop if T_{BDT} is > 450 °C. In case the T_{BDT} is < 450 °C only minor supercritical resources develop because the threshold permeability is encountered at temperatures slightly higher than the critical temperature of water.

Supercritical water resources and geology



Liquid (grey dashed lines) and vapour (black dashed lines)

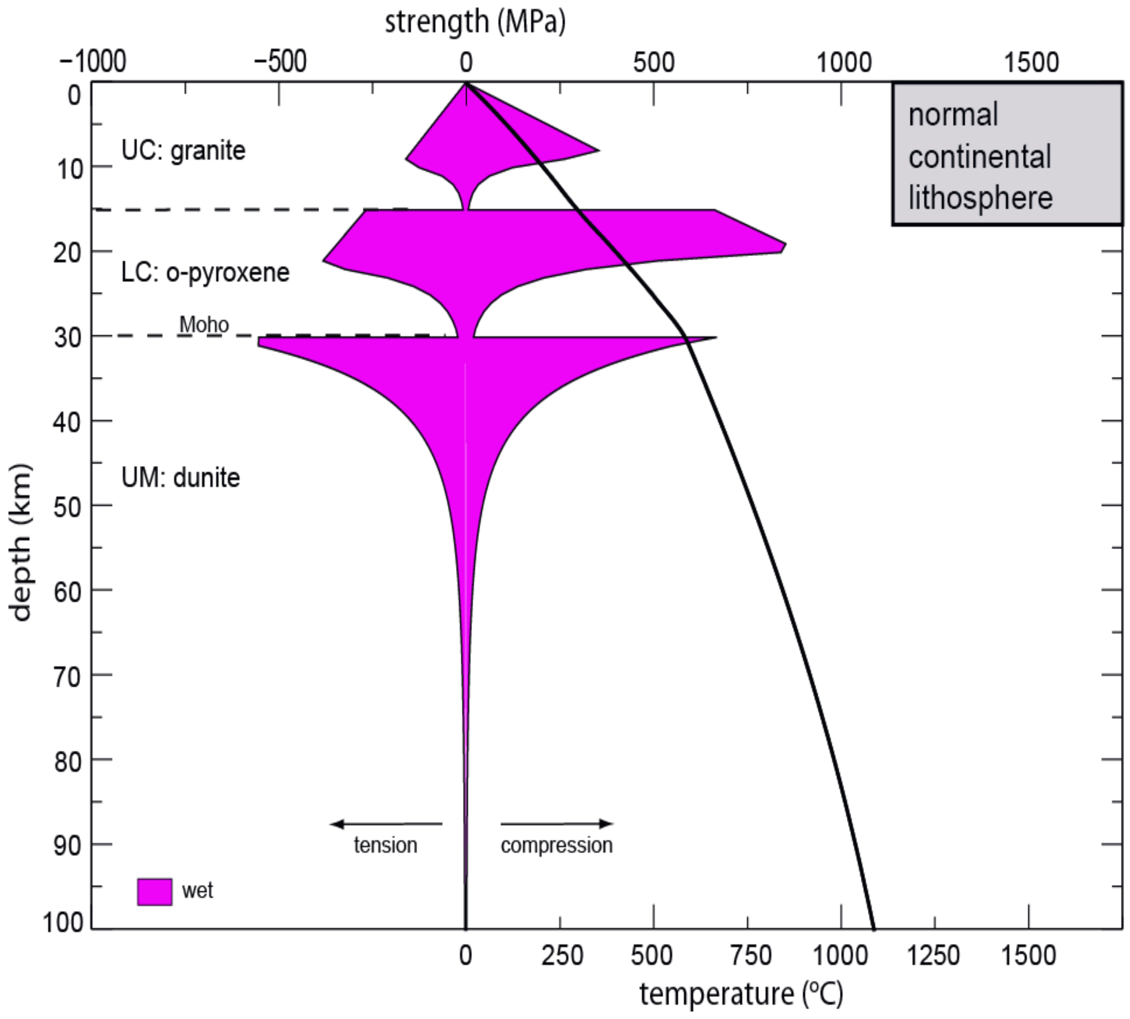
Scott et al., 2015, Nat. Communications

- In high-permeability host rocks, the rate of convective water circulation surpasses the ability of the intrusion to heat most circulating water to supercritical temperatures, and supercritical water flow is confined to a thin (~10 m) boundary layer on the perimeter of the intrusion.
- In contrast, supercritical resources in intermediate permeability systems are hotter. The water circulation rate near the intrusion is lower compared with high-permeability systems, so the conductive heat input across the BDT is sufficient to heat up a larger fraction of the circulating water to supercritical temperatures.
- The transition from supercritical to boiling conditions occurs over a small *P* range in high-permeability systems and more gradually in intermediate permeability systems.

Brittle-Ductile Transition (BDT) and rheology

Brittle-Ductile Transition occurs at different T for different lithology (rheology)

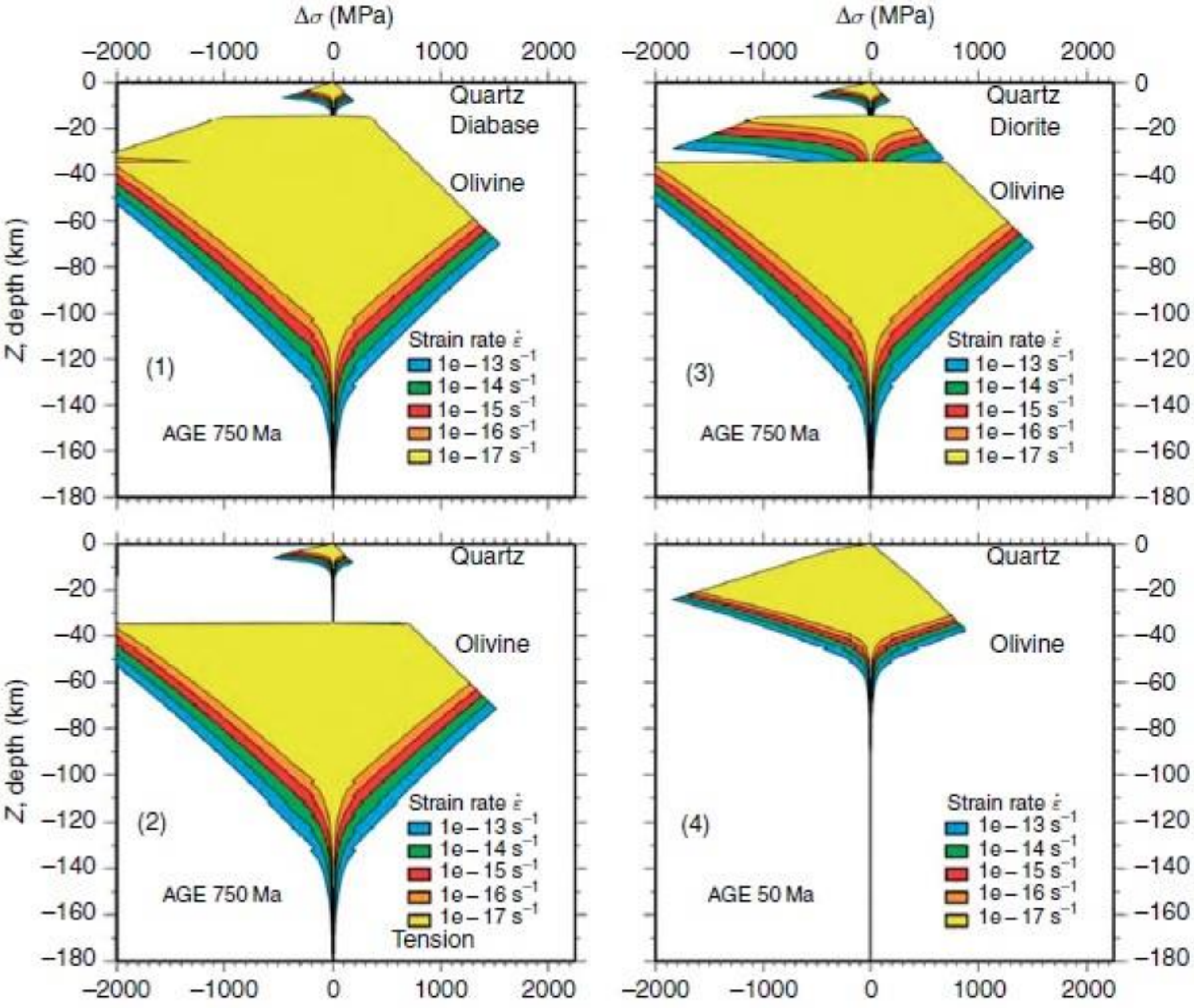
Continental lithosphere



- High-temperature hydro-thermal systems on land having an upper *T* limit of~400°C would imply that: (1) permeability effectively ceases at that temperature due to the BDT; (2) permeability is limited by self-sealing due to hydrothermal alteration at higher temperatures; or (3) *T* are controlled by transitions from subcritical to supercritical conditions.
- The *T* of the BDT depends on the silica content of the rock: this transition occurs about 380–400°C in rhyolites or granites rocks and at 500–600°C in basalts or gabbro (for strain rates ~10^{-15} s^{-1}).
- In Iceland, seismic events occur still at depths of ~8 km beneath the high-temperature geothermal systems in Iceland, where *T* is estimated to be > 700°C: fractures can form and persist for some period of time within rocks that should be deforming plastically in a longer time frame.
- High-enthalpy, high-pressure, supercritical fluid exists in the deeper reservoir of Iceland below 3 km. Such fluids dramatically increase the potential for rock fracturing by stress-corrosion micro-cracking.

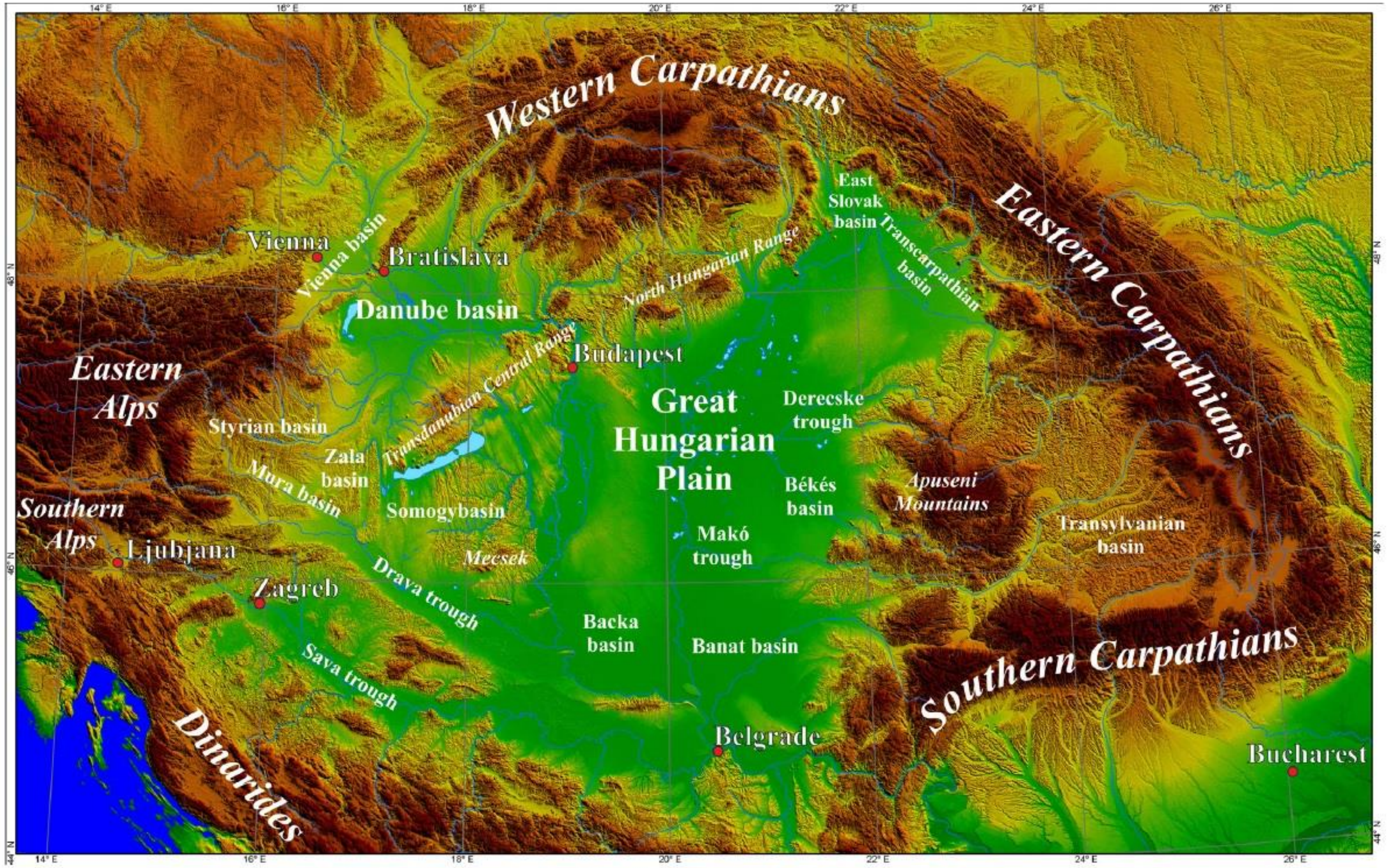
Brittle-Ductile Transition (BDT) and strain rates

YSEs strain rates dependent



- If the strain rates increases (by applying far field forces), the depth of the BDT increases.
- The strain rate dependence of the BDT suggests that supercritical waters may not be restricted to basaltic systems but may pertain also to silicic rocks if tectonic deformation rates are high enough.
- In extensional conditions the BDT extends a larger depth than in compressional conditions.

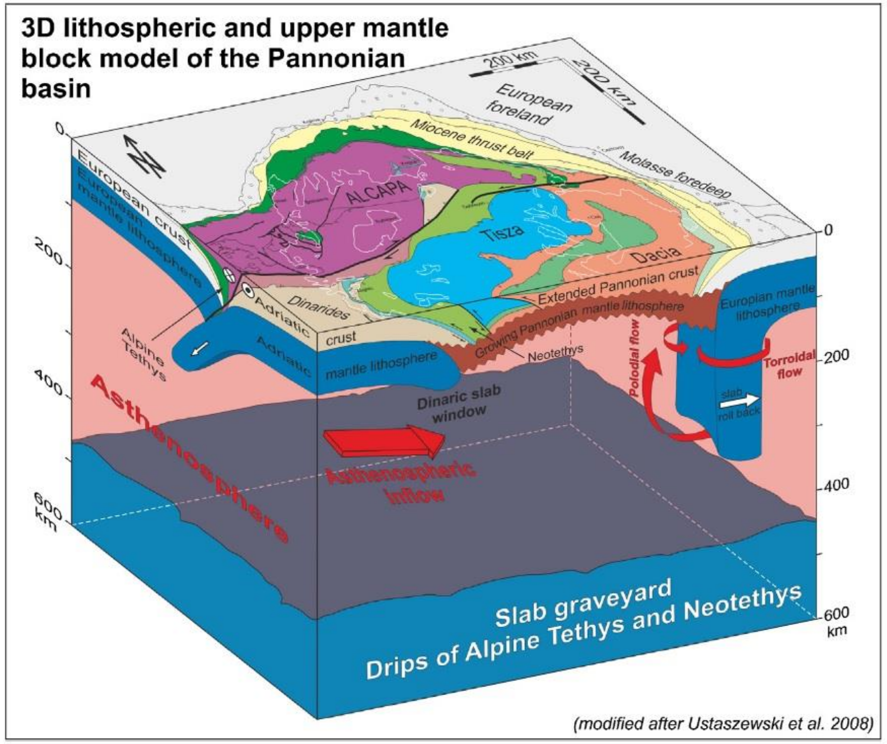
Pannonian Basin

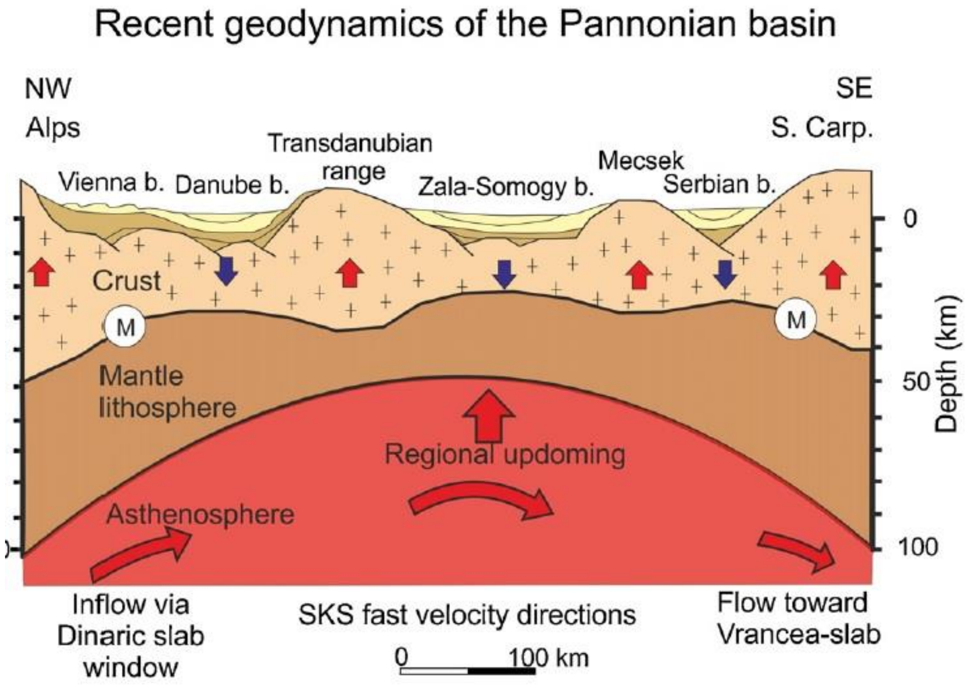


Horváth et al., 2015, Geothermics, 53

The Pannonian basin is a backarc basin, whose formation, started at the beginning of the Miocene, was accompanied by intensive calc-alkaline magmatism. The basin developed from extensional disintegration of orogenic terranes and subsequent events of basin inversion. These deformations resulted in variable basin, characterized by deep half grabens and relative basement highs.

Geodynamic evolution of the Pannonian Basin



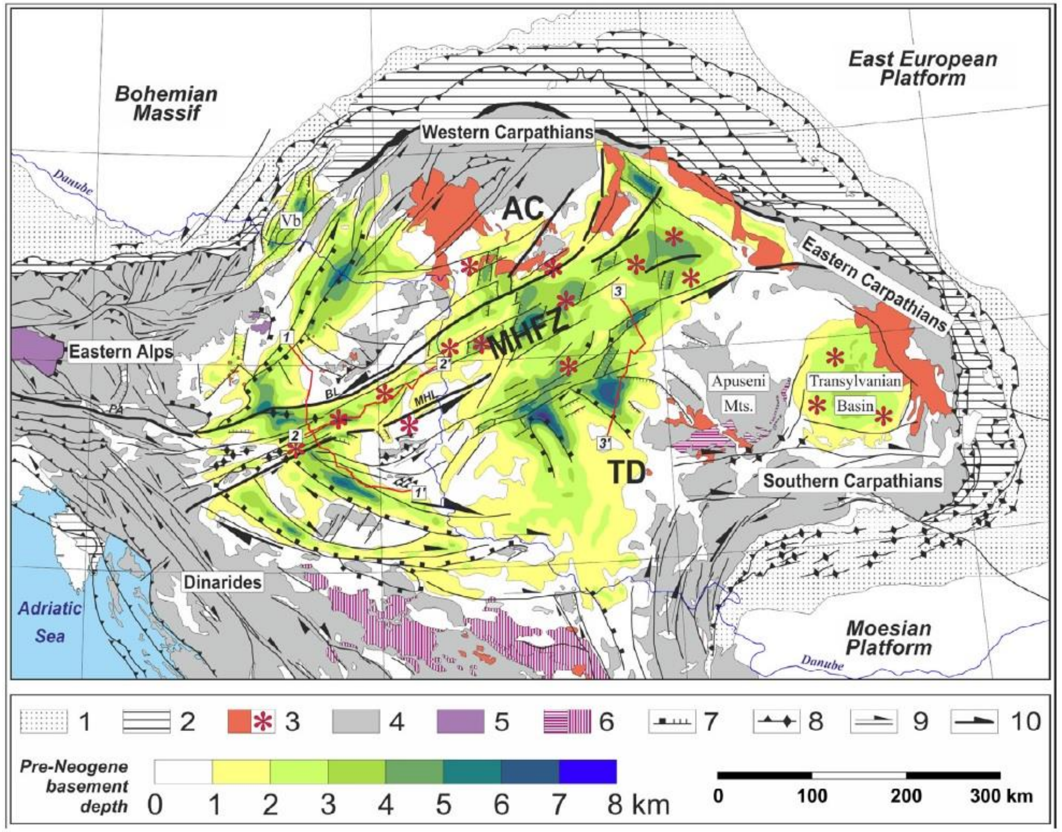


Horváth et al., 2015, Geothermics, 53

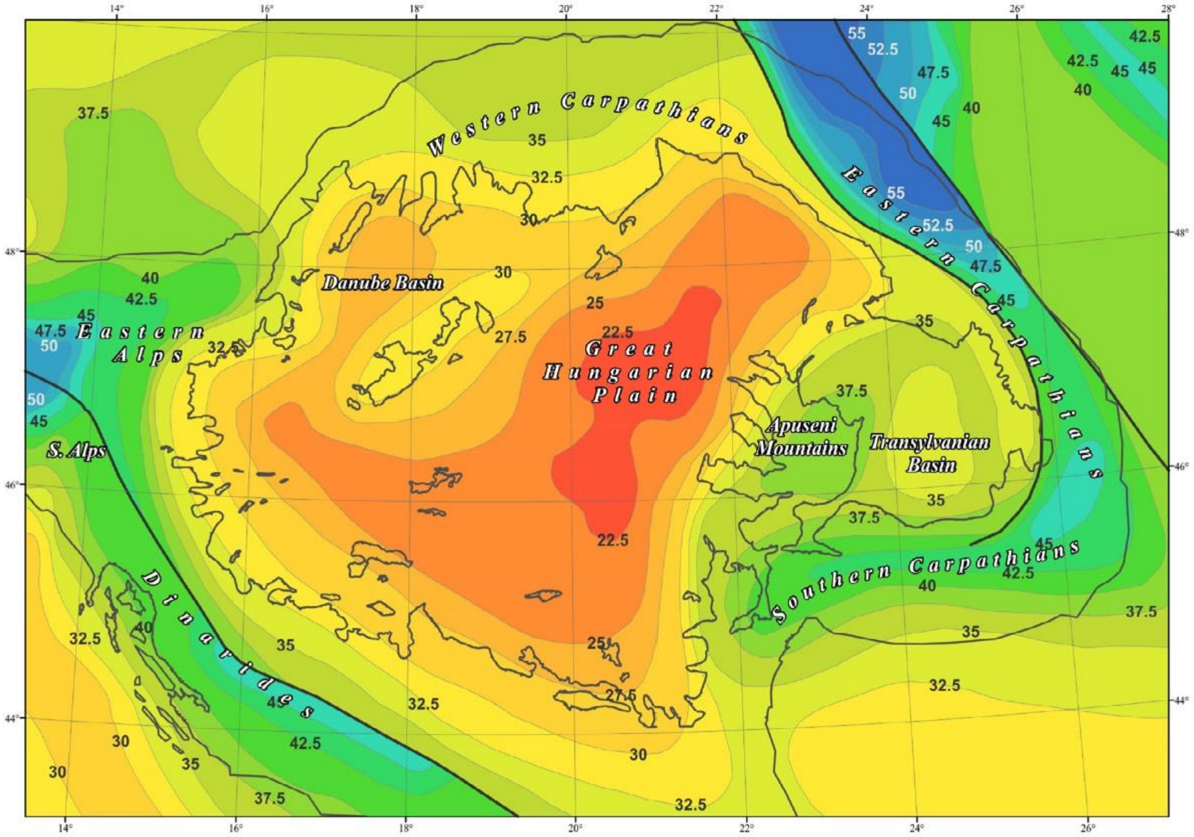
- Moderate crustal extension was accompanied by large attenuation of the mantle lithosphere during the syn-rift phase, which lead to felsic magmas formation between 21 and 11 Myr.
- A first phase of passive rifting due to extensional stresses generated by slab rollback was followed by an active mantle lithosphere thinning as a result of buoyancy induced asthenospheric uprise beneath the rift.
- The Pannonian basin is likely started to form with a simple shear phase, followed by a pure shear lithospheric deformation. In the Early and Middle Miocene during the retreat of the Carpathian slab, there was crustal extension and subsidence in the Pannonian region. From the Late Miocene, the whole lithosphere of the Pannonian basin extended in a uniform way.
- Currently, the attenuated crust is under compression, which generates differential vertical movements. Lithosphere is thin and asthenospheric flow system sustains it in an elevated position relative to its isostatic equilibrium position.

Pannonian Basin

Depth of the Basement



Crustal Thickness

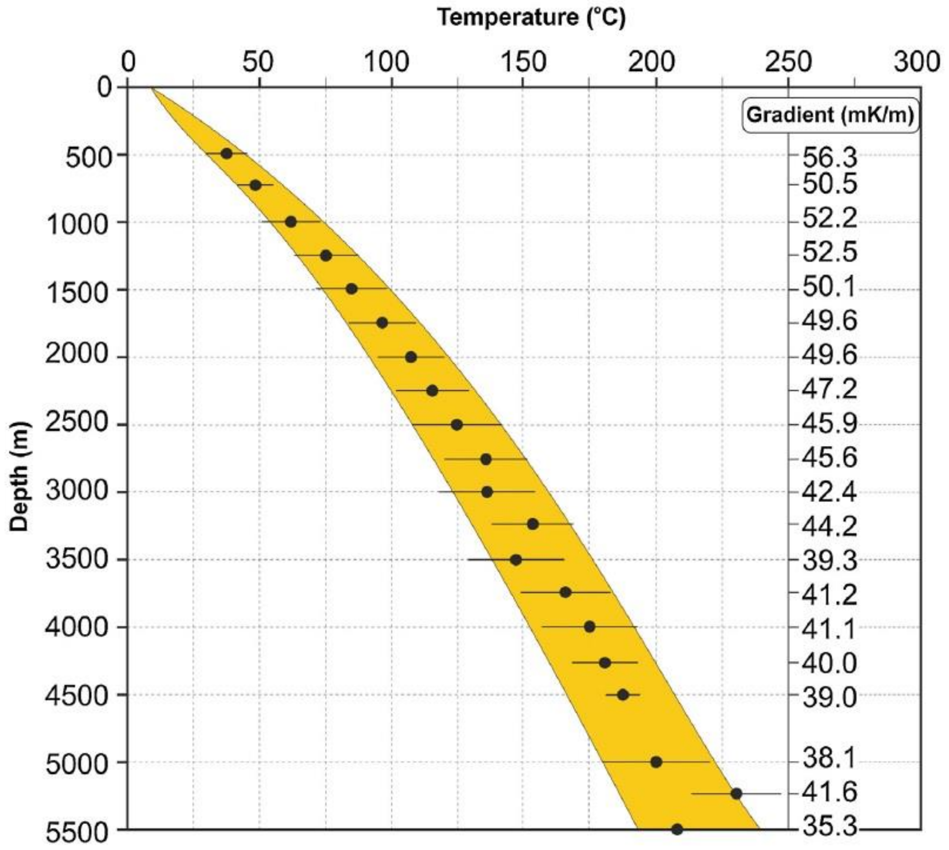


Horváth et al., 2015, Geothermics, 53

Lithospheric thickness in the Pannonian Basin is ~50-60 km, while the depth of Moho varies in the range of 32–22 km, mirroring the first order pattern of the basement subsidence.

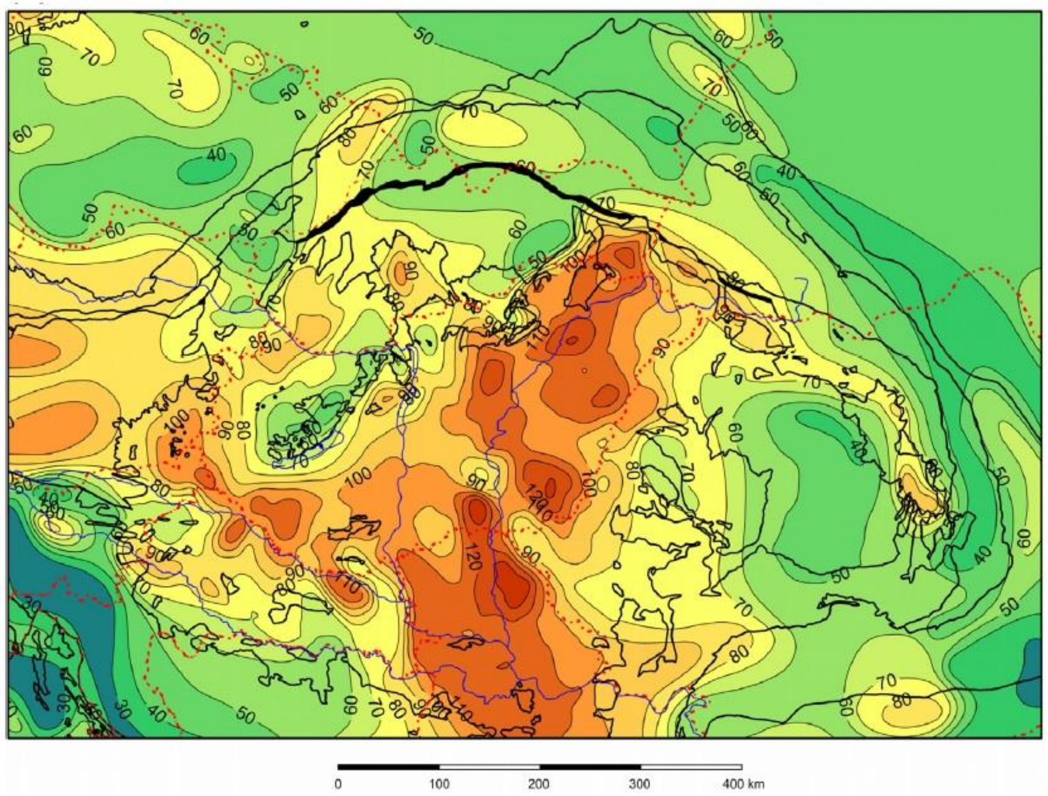
Thermal Conditions of the Pannonian Basin

Geothermal Gradient



Surface Heat Flow

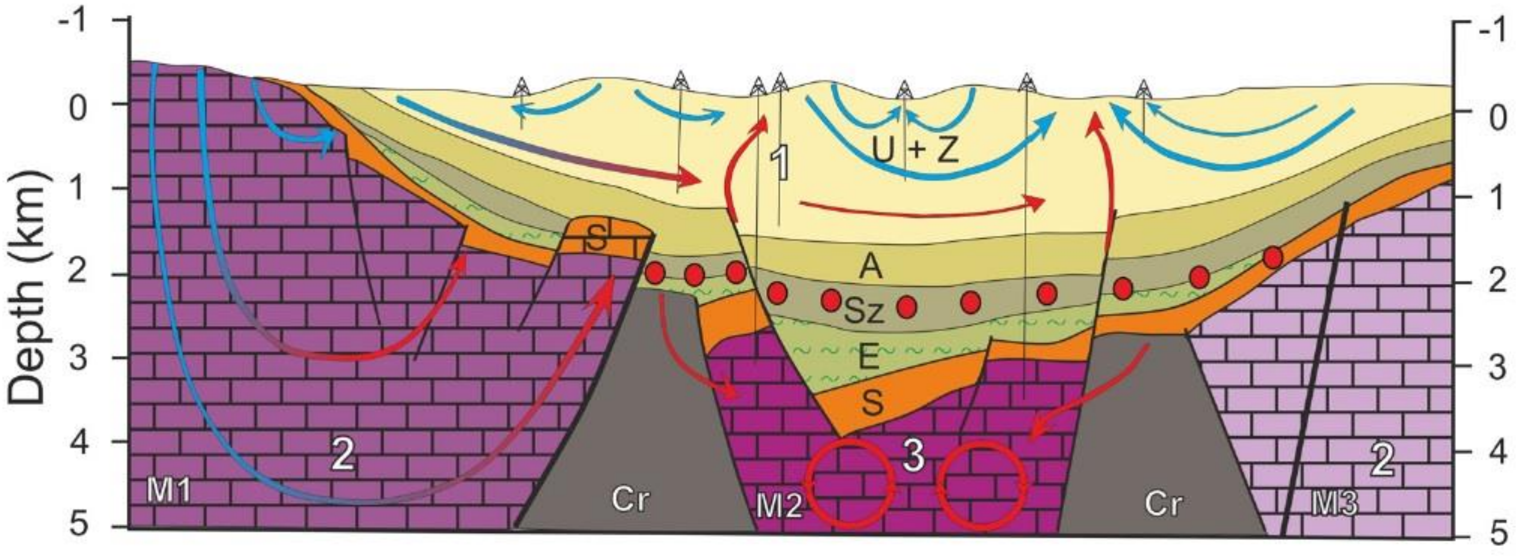
(corrected for the variation in the sedimentation rate and the change in the thermal properties of sediments due to compaction)



Horváth et al., 2015, Geothermics, 53

- Temperature gradient in the Pannoinian basin varies between 40 and 50 mK/m (200°C at ~5 km).
- The heat flow distribution in the Pannonian basin shows values ranging from 50 to 130 mW/m^2, with a mean value of about 100 mW/m^2. The Carpathians and the Bohemian Massif show heat flow values of 50–70 mW/m^2, while the Outer Dinarides exhibit extremely low heat flow (about 30 mW/m^2) due to cooling by meteoric water inflowing at the high karst plateau (Mesozoic carbonate).

Pannonian Basin Hydrothermal Setting

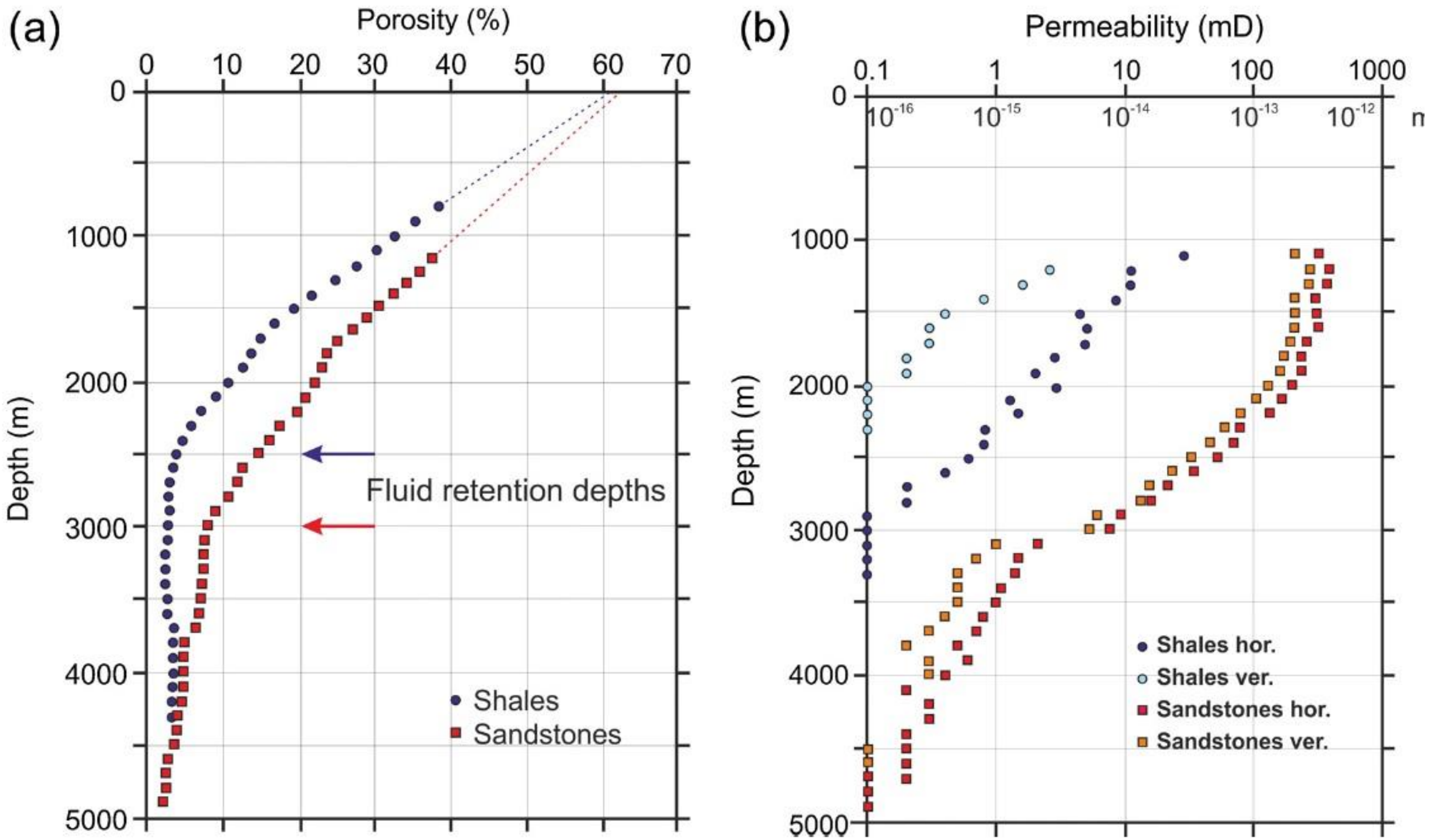


1 and 2 = gravity-driven flow system in the porous basin fill rocks, and Mesozoic (locally Eocene and/or Miocene) carbonates. 3: an overpressured system below a pressure seal (red ellipses) including lower Pannonian and Early to Middle Miocene basin fill, and fractured basement rocks. U + Z = Újfalu and Zagyva; A = Algyo; Sz = Szolnok; E = Endrod; S = Synrift; M1,2,3 = Mesozoic rocks belonging to the Transdanubian Range unit, the MHFZ and Villány - Mecsekunit; Cr = Crystalline basement.

Horváth et al., 2015, Geothermics, 53

- **Two superimposed hydraulic systems can be distinguished in the Pannonian basin representing an upper and a lower domain (spatially separated, but not isolated)**: (1) In the upper domain a gravity-driven water flow system exists, which is regionally unconfined, hydrostatically pressured and recharged from precipitation.
- The first system, in which water circulation is gravity-driven caused by the topography gradient, prevails in the porous sedimentary rocks (marls, sandstones) of the basin fill from the surface down to a depth of about 2000 m. In addition, exposed older permeable rocks (mostly Mesozoic carbonates) represent another type of gravity-driven flow system.
- The lower hydraulic domain is a regionally confined system characterized by remarkable overpressures below a low permeability pressure seal and contains highly saline waters expelled from sedimentary rocks during compaction.

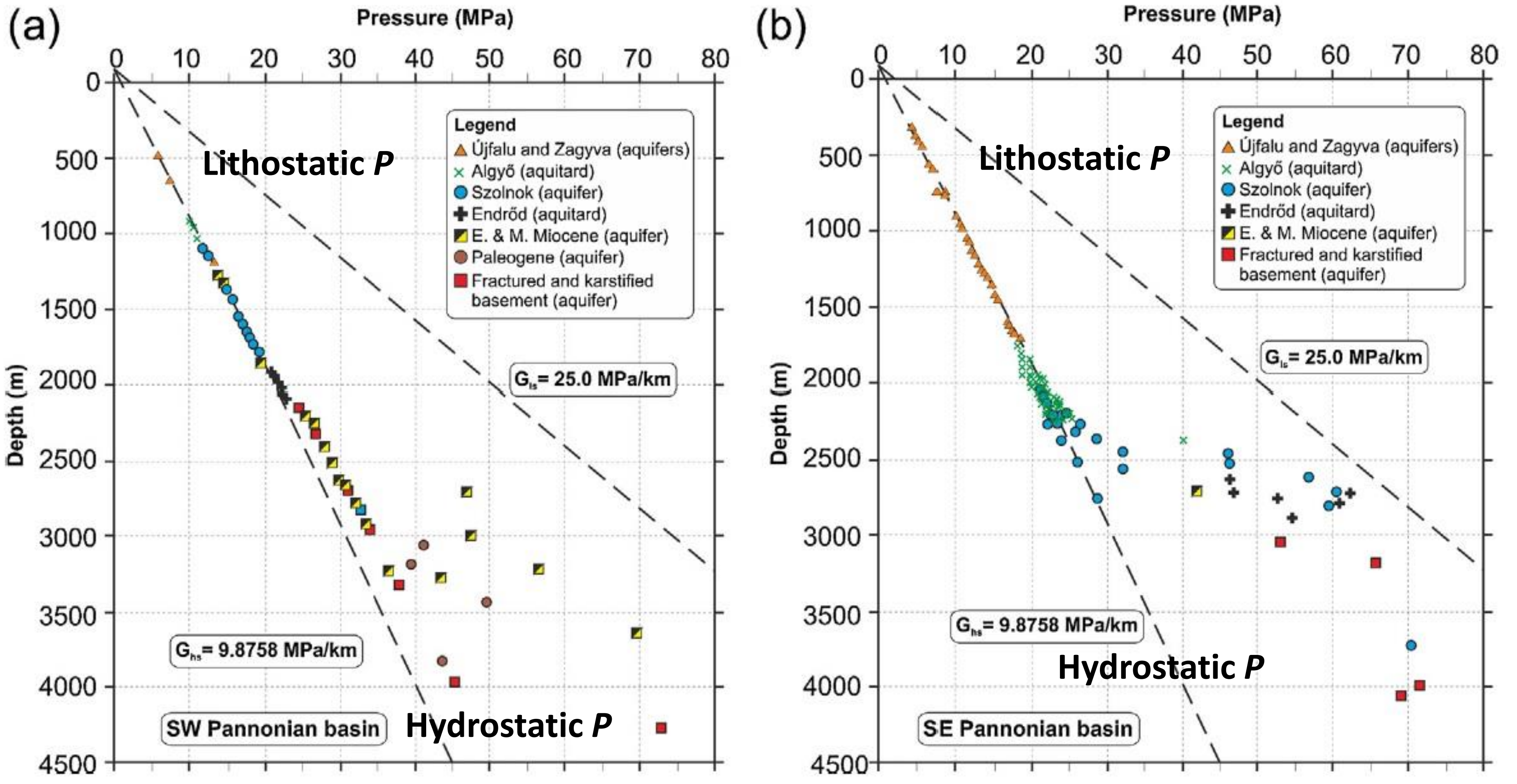
Pannonian Basin Hydrothermal Setting



Horváth et al., 2015, Geothermics, 53

Overpressure conditions in the Pannoinian Basin are likely caused by: **(1) disequilibrium compaction mechanism** (fast sedimentation): As displayed in Fig. (a), from 2500 to 3000 m depth disequilibrium compaction leads to isolation of pores and fluid retention; **(2) neotectonic changes from extensional to compressional stress** (largest overpressures developed at the elevated parts of the basement, rather than in the deepest grabens).

Pannonian Basin Hydrothermal Setting

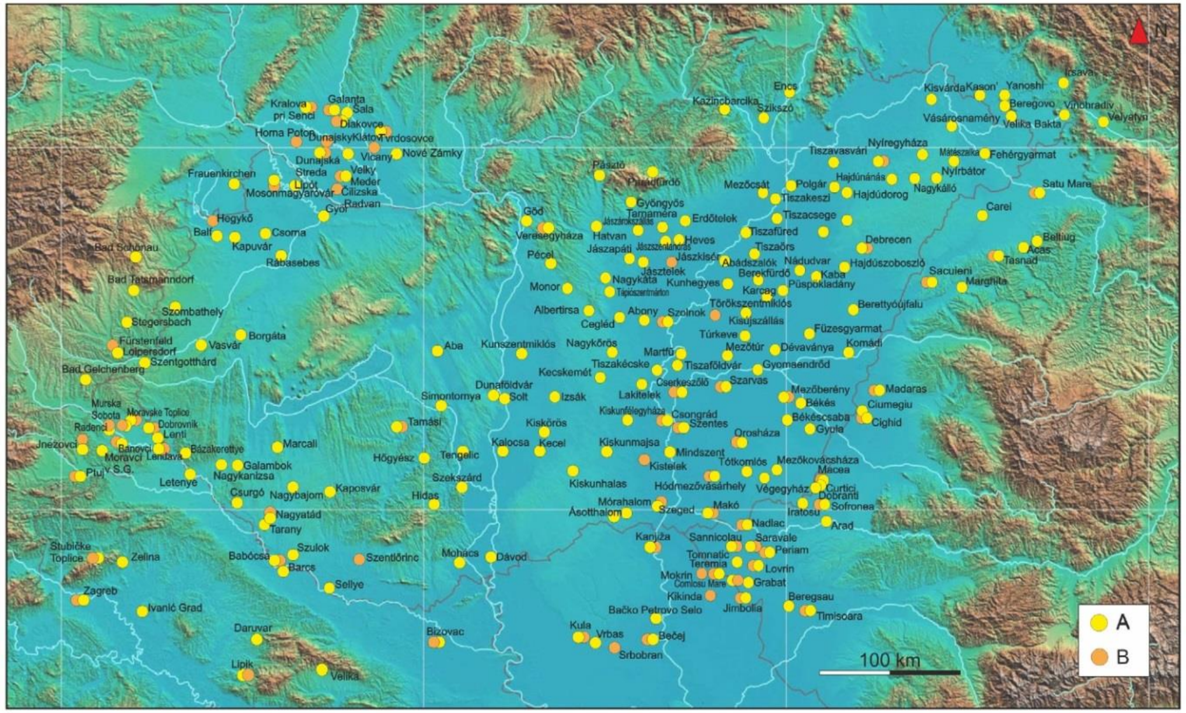


Horváth et al., 2015, Geothermics, 53

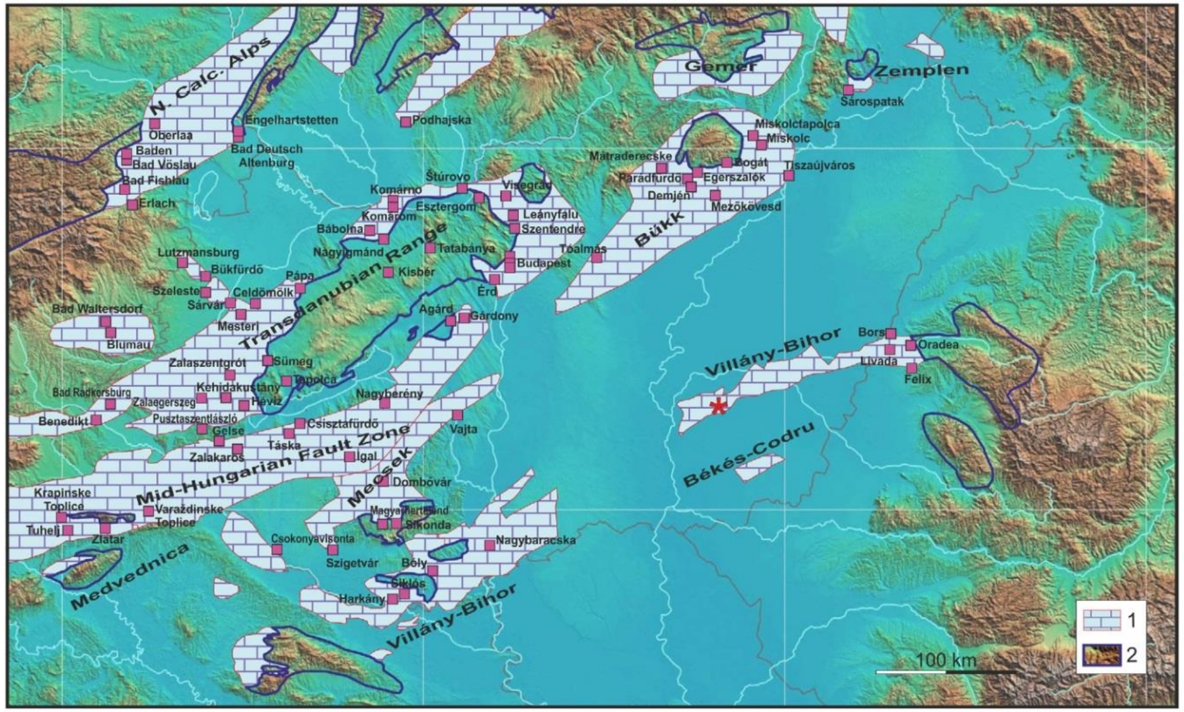
- In the southwestern Pannonian basin overpressure conditions appear with a pressure gradient slightly above normal between 2200 and 3000 m.
- In the southeastern Pannonian basin the increase of *P* seems to be very sharp in the 2200–2900 m depth range.

Geothermal installations in the Pannoinian Basin

Main geothermal installations in the Pannonian Basin



Main Mesozoic karstic reservoirs below the surface (symbol 1) in the Pannonian basin and their recharge areas (symbol 2)



Horváth et al., 2015, Geothermics, 53

- The porous aquifer produce thermal waters typically of 200–300 m^3/day, and *T* in the range of 30–100°C (increasing with depth).
- There is no utilization of the high-*T*-high-*P* geothermal system developed beneath a pressure seal in the deeper part of the basin and fractured basement.