Course of Geothermics

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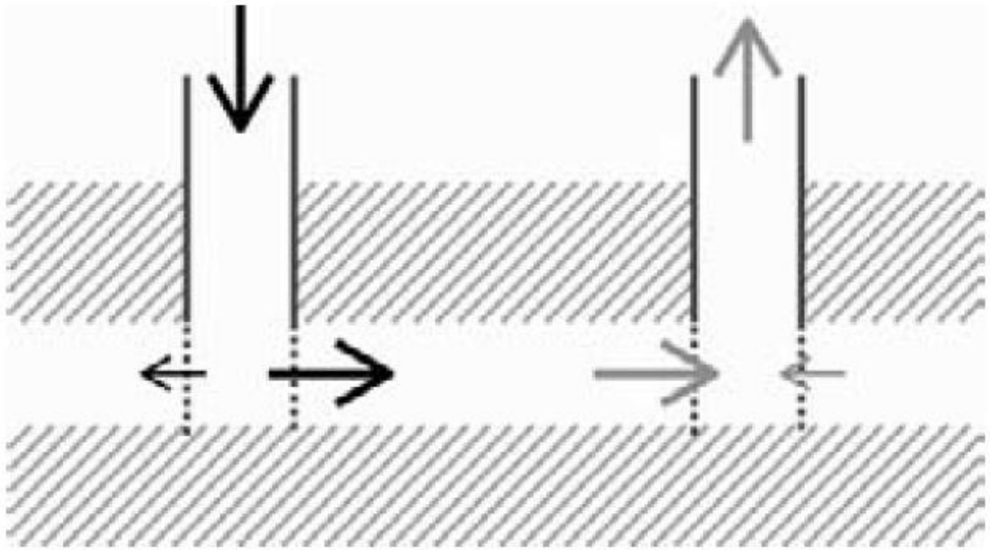
Course Outline:

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
12. **Geothermal Systems**

Geothermal Resources

- A region of potential geothermal resource may be associated with high heat flow.
- The near surface temperature gradients can be extrapolated through impermeable strata to obtain deep temperatures, but cannot be continued through permeable strata because convection/advection may determine the temperature distribution.
- The general reduction of permeability with depth implies that successful production from greater than a few km requires anomalously high permeability, and the chance of such anomaly decreases with increasing depth.
- Deep sedimentary aquifers heated by the normal thermal gradient are found in many continental environments (e.g., a carbonate/sandstone aquifer in the Paris basin).
- Many warm springs are found along major fault and fracture lineations throughout the world, which provide the channels for the flows of warm water that feed the springs. These channels are a form of convective system.
- Lithology, fault scale and geometry, self-sealing processes, fluid chemistry and degassing rates, *P*-*T* history and the stage of fault evolution may also be important controlling factors in the fault behaviour and the fluid circulation at depth.



Injection and Production well in sedimentary aquifer

Once exploitation occurs, fluid flow to and from wells is generally much greater than the natural flow: With large-scale production and reinjection, the primary induced flow is from the injection wells to the production wells.

Geothermal Resources

Catalog scheme of geothermal resources by temperature according to different authors

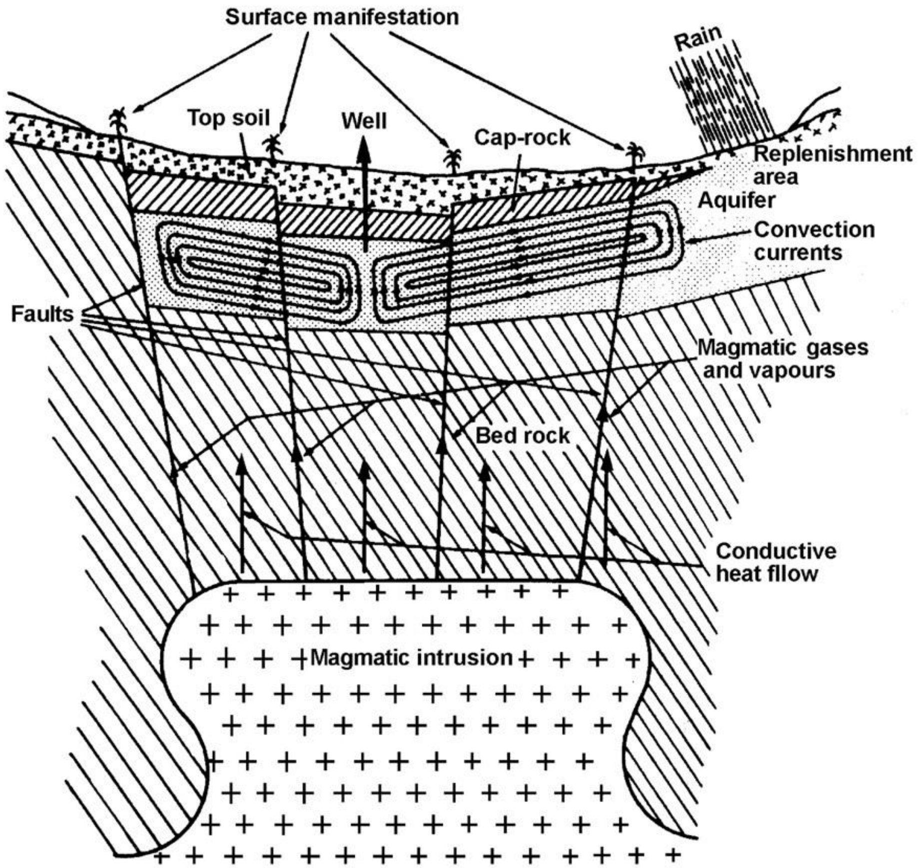
	Muffler [8] (°C)	Hochstein [9] (°C)	Benderitter and Cormy [12] (°C)	Haenel et al. [10] (°C)		
Low enthalpy	< 90	< 125	< 100	< 150		
Moderate enthalpy	90–150	125–225	100–200	–		
High enthalpy	> 150	> 225	> 200	> 150		
Sanyal [13]	Non-electrical (°C)	Very low (°C)	Low (°C)	Moderate (°C)	High (°C)	Ultra high (°C)
	< 50–100	100–150	150–180	180–230	230–300	> 300

Moeck, 2014, Renewable and Sustainable Energy Reviews, 37

- At 40–60 ◦C they are primarily used in greenhouse or covered ground heating, aquaculture, as well as bathing and swimming applications.
- Beyond 60–70 ◦C, space and domestic water heating is the most common application (washing, cooking, sterilising, drying).
- Geothermal binary plants for electricity production require primary fluids at surface having a temperature of at least 100 ◦C.

What is a Geothermal System?

- The basic components of the geothermal system are: (1) an aquifer or fracture network containing hot fluid, (2) a path through which cold water or an input of magmatic fluid can flow to recharge the system, (3) a cap rock, i.e. a low permeability layer which restrains the main fluid flow at a depth where the *T* is high and (3) a heat source (e.g., a magma chamber).
- The *P* drive that sustains the hot recharge or upflow is the buoyancy difference between the columns of descending cold and ascending hot water.

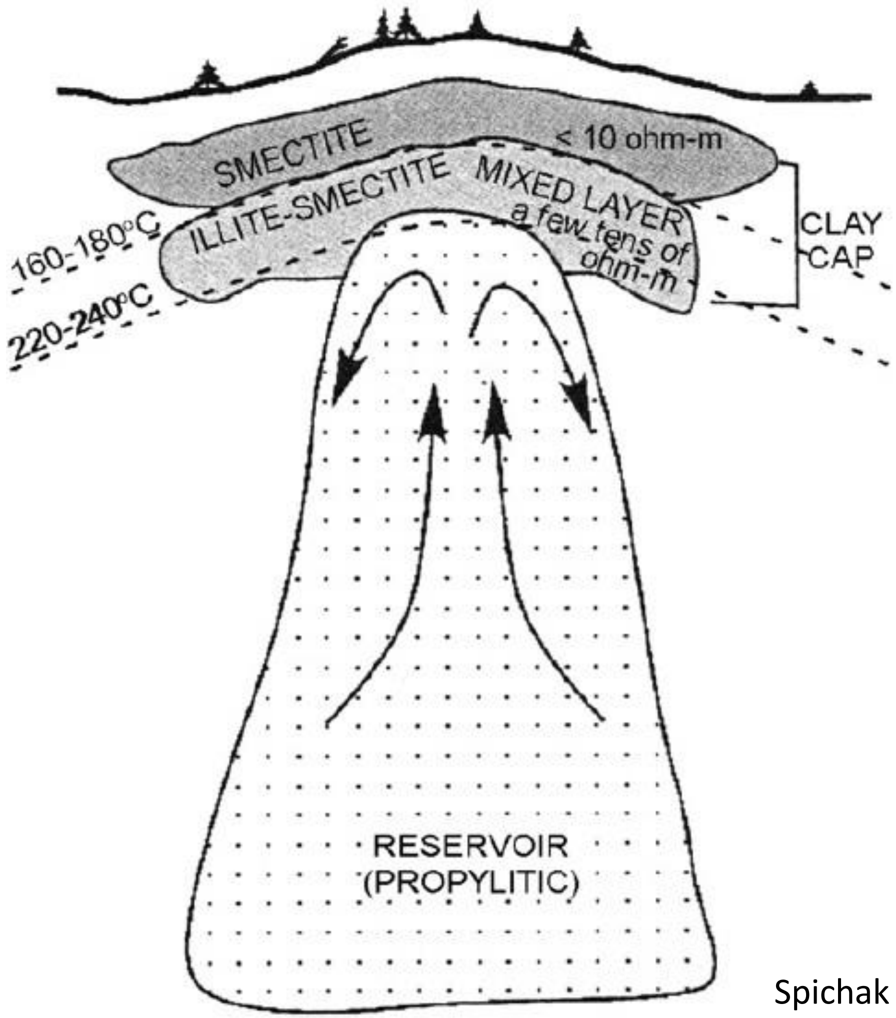


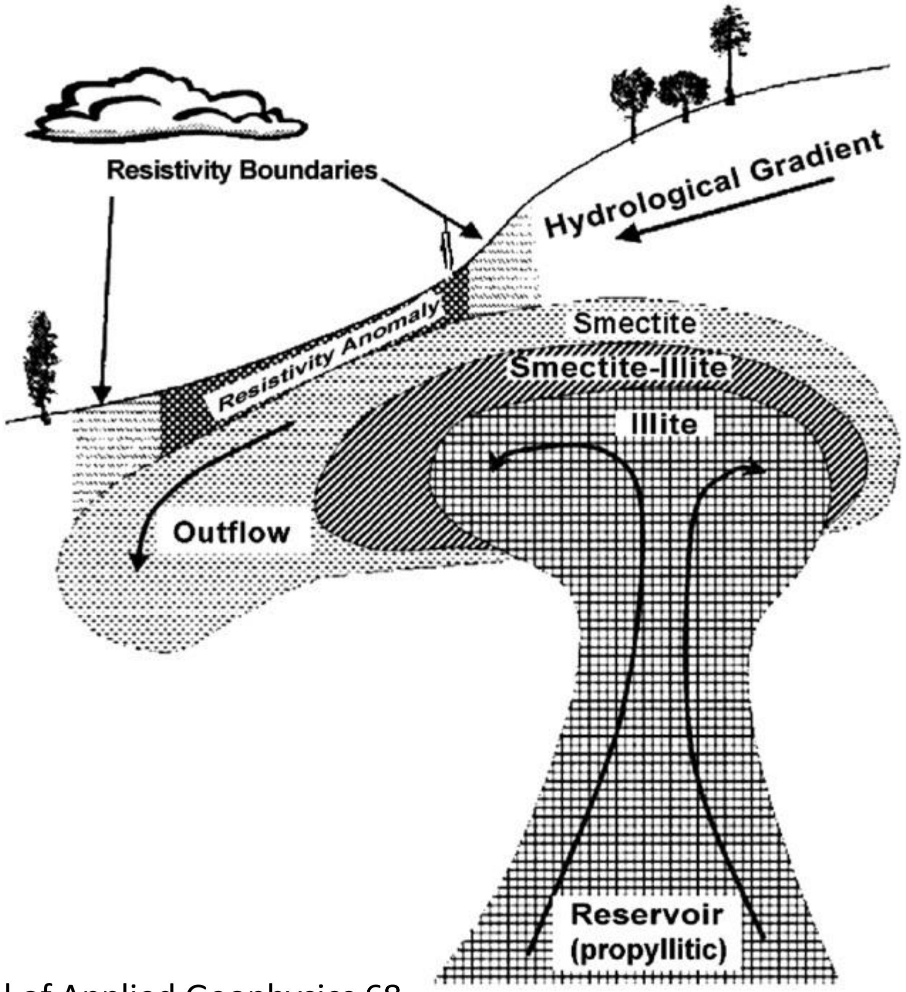
Spichak and Manzella, 2009, Journal of Applied Geophysics, 68

What is a Geothermal System?

- In geothermal areas where the permeability is high and alteration pervasive, the lowest resistivity corresponds to a clay cap overlying the geothermal reservoir, which in turn has a much higher resistivity (e.g., Iceland, New Zealand, Indonesia).

- When topography is steep, the overall structure of the geothermal system is more complex: The conductive clay layer may be quite deep over the system upflow and much closer to the surface in cooler outflow areas.





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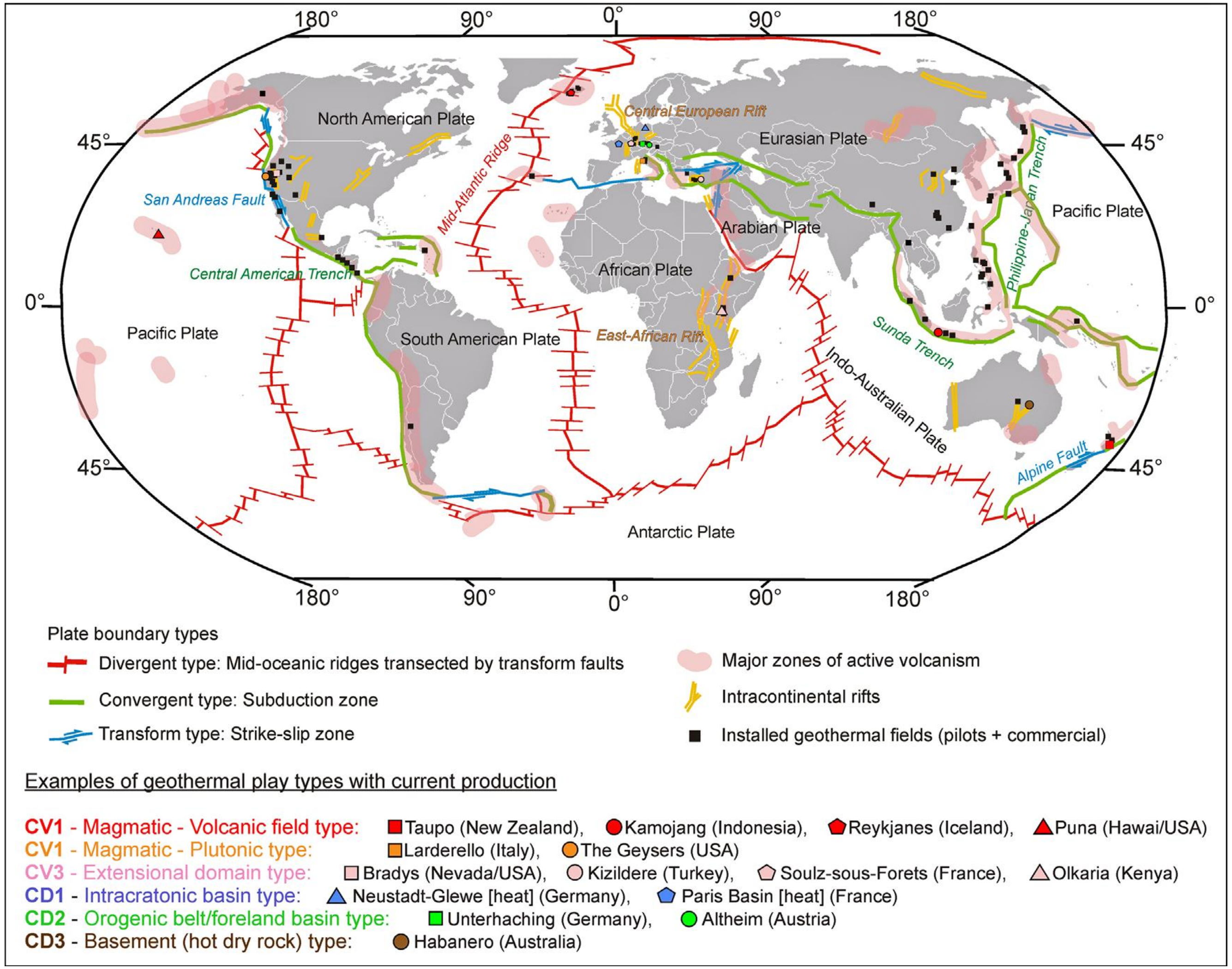
Geothermal Systems Types



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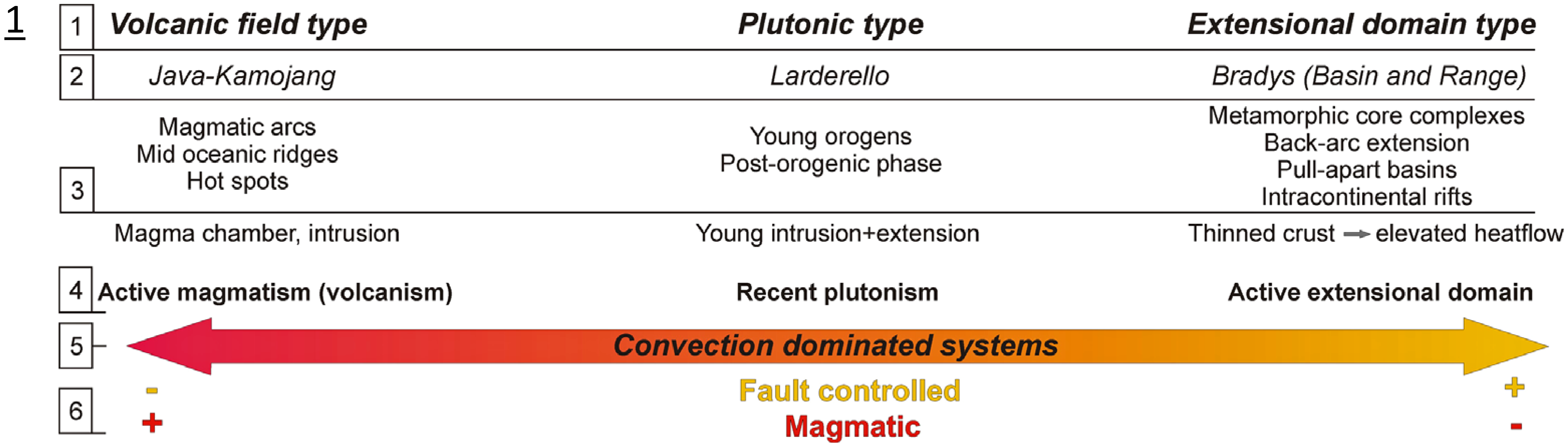
- Convection-dominated geothermal plays host high enthalpy resources and occur at plate tectonic margins, or settings of active tectonism or volcanism.
- Conduction-dominated geothermal plays host low to medium enthalpy resources, are located predominately at passive tectonic plate settings, where no significant recent tectonism or volcanism occurs.

Convection Dominated Geothermal System



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- Convection-dominated geothermal systems are controlled by either an igneous activity like a magma chamber in volcanic areas, or faults in extensional terrains, or both, such as intrusive bodies at fault zones.
- Plutonism controlled geothermal systems are typically located along continent-continent convergent or transform margins with recent magmatism and with (e.g. Larderello) or without (e.g. the Geysers) recent recharge of meteoric water. In Laderello, granite intrusions are associated with young (0.3–0.2 Ma) magmatism generating a fluid-dominated (K-horizon) layer above the granite and a vapor-dominated (H-horizon) layer above the fluid-dominated layer.

Conduction Dominated Geothermal System

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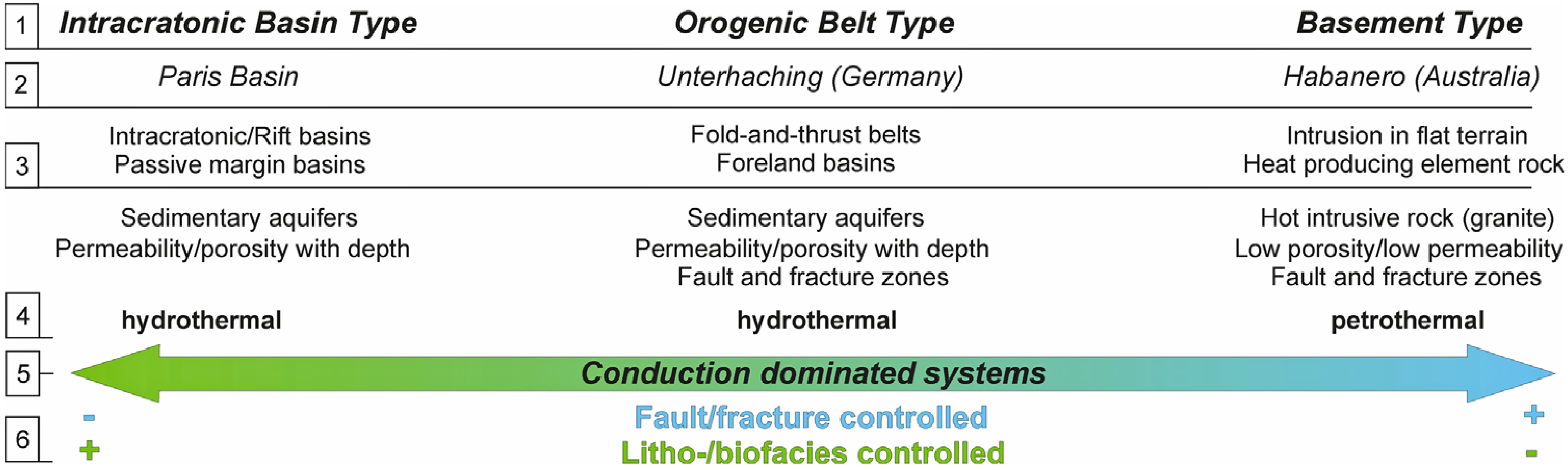

- These types of settings are further considered with respect to the porosity-permeability ratio of the reservoir rock and the absence or presence of producible fluids in the reservoir.

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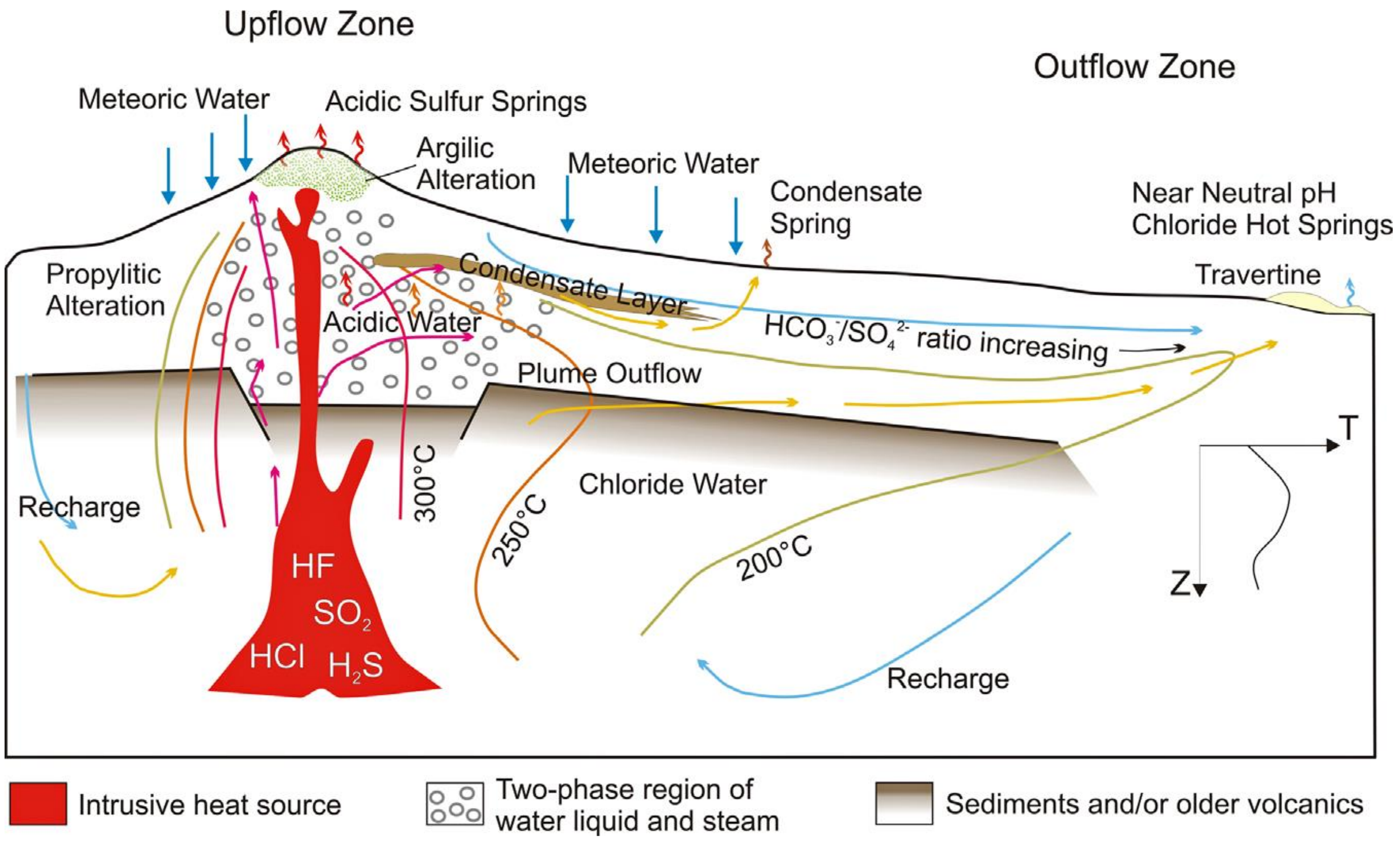
Convection Dominated Geothermal System



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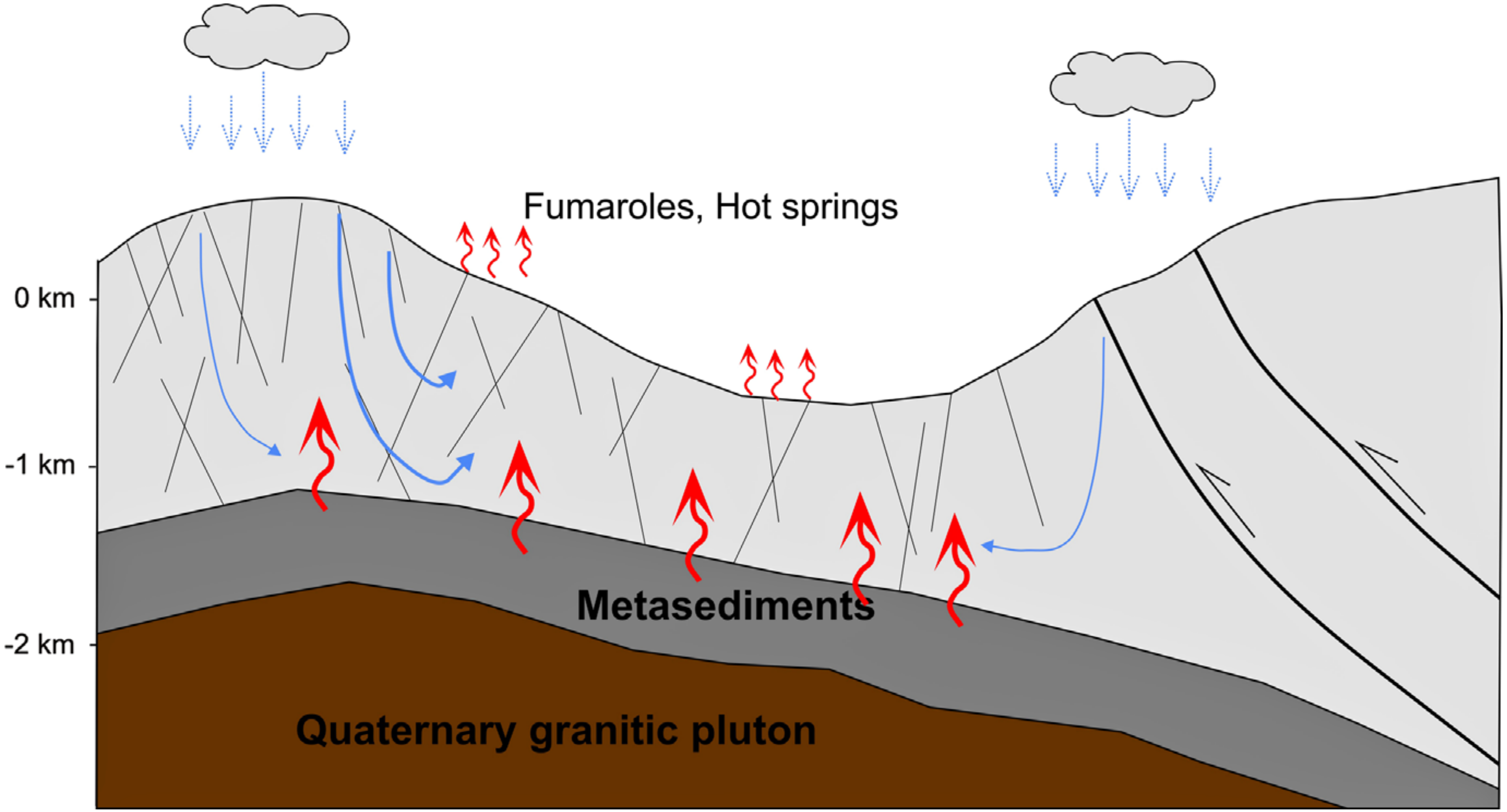
- Geothermal manifestations in the upflow zone are acidic springs associated with thermo-chemically altered rock forming alteration clays that indicate high temperature systems below the spring.
- The upflow zone commonly consists of a vapor-dominated part above a liquid-dominated part.
- Condensate layers in steep terrain, such as volcanoes, can be generated by upwelling fluids condense above the heat source and can neutralize initially acidic fluids.

Convection Dominated Geothermal System



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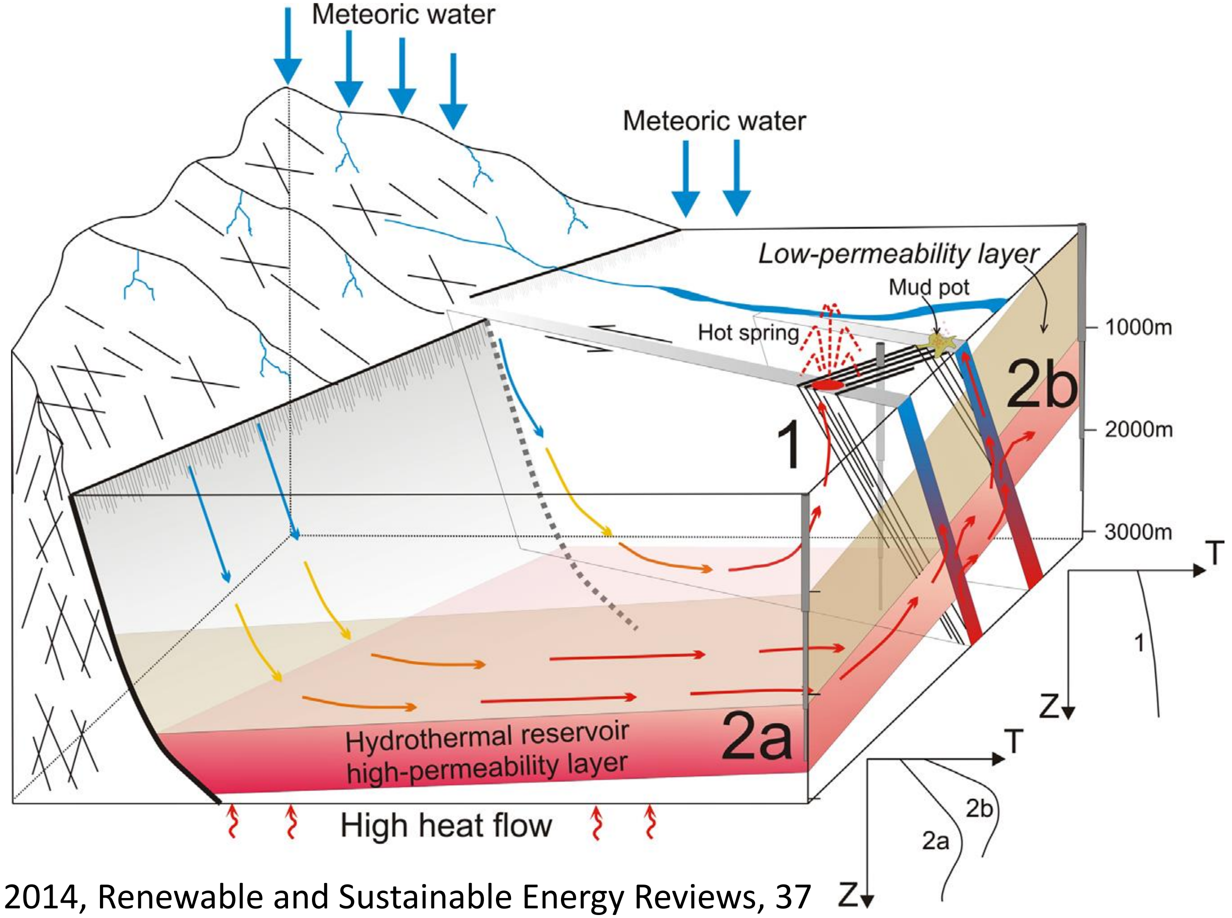
- Geothermal systems related to recent plutonic fields are typically found at intrusions along continent-continent or continent-oceanic convergent or transform margins (e.g., Lardarello, associated with recent plutonism and extension).
- The Geysers field is an example of these systems, being charcaterized by a large felsic pluton provides the heat source for a vapor dominated fluid in porous meta-sedimentary reservoir capped by a low permeability serpentinite/clay.

Convection Dominated Geothermal System



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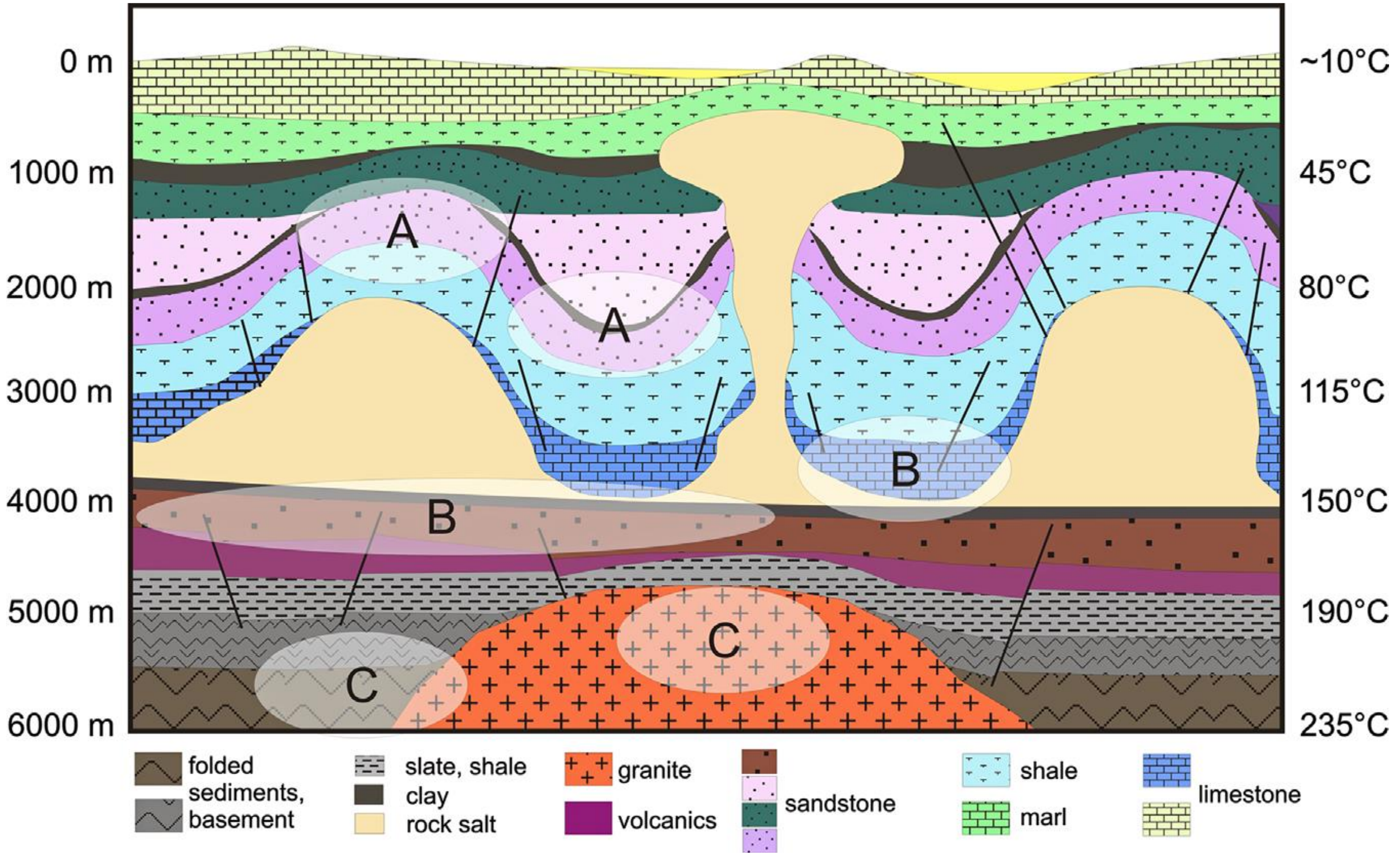
- Type1 is a convection cell from infiltration to discharge along one fault. Temperature gradient is gradually increasing at well site 1. Type 2a and 2b are fault leakage controlled systems. The temperature gradient of a well drilled in to such an area rises up to the permeable layer and drops below the layer (well2a and2b).
- As thermal fluids move away from the upwelling zone along a fault zone, they mix with cooler ground water or meteoric water, as indicated by an increase of bicarbonate and magnesium and decrease of boron, sulfate and chloride.

Conduction Dominated Geothermal System



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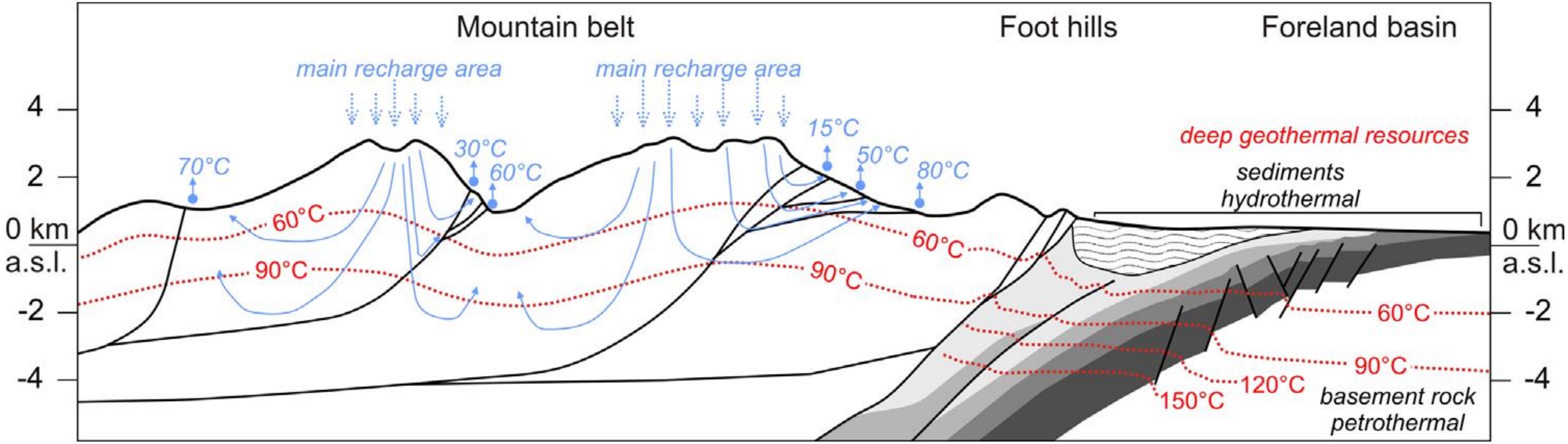
- Geothermal Systems at different depths and temperatures in intracratonic basins settings: A – Geothermal systems above 3 km depth with temperature suitable for district heating, B – Deep geothermal systems below 3km depth suitable for heating and electricity, C – Very deep geothermal below 4 km depth as potential Hot Dry Rock (HDR) systems.
- Geothermal reservoirs are located in different basin portions depending on the internal present-day structure of the basin (e.g., crustal regions above salt formations are suitable for district heating), as well as the heat content of the fluid in the porous layer setting is strongly affected by the basin geometry.

Conduction Dominated Geothermal System



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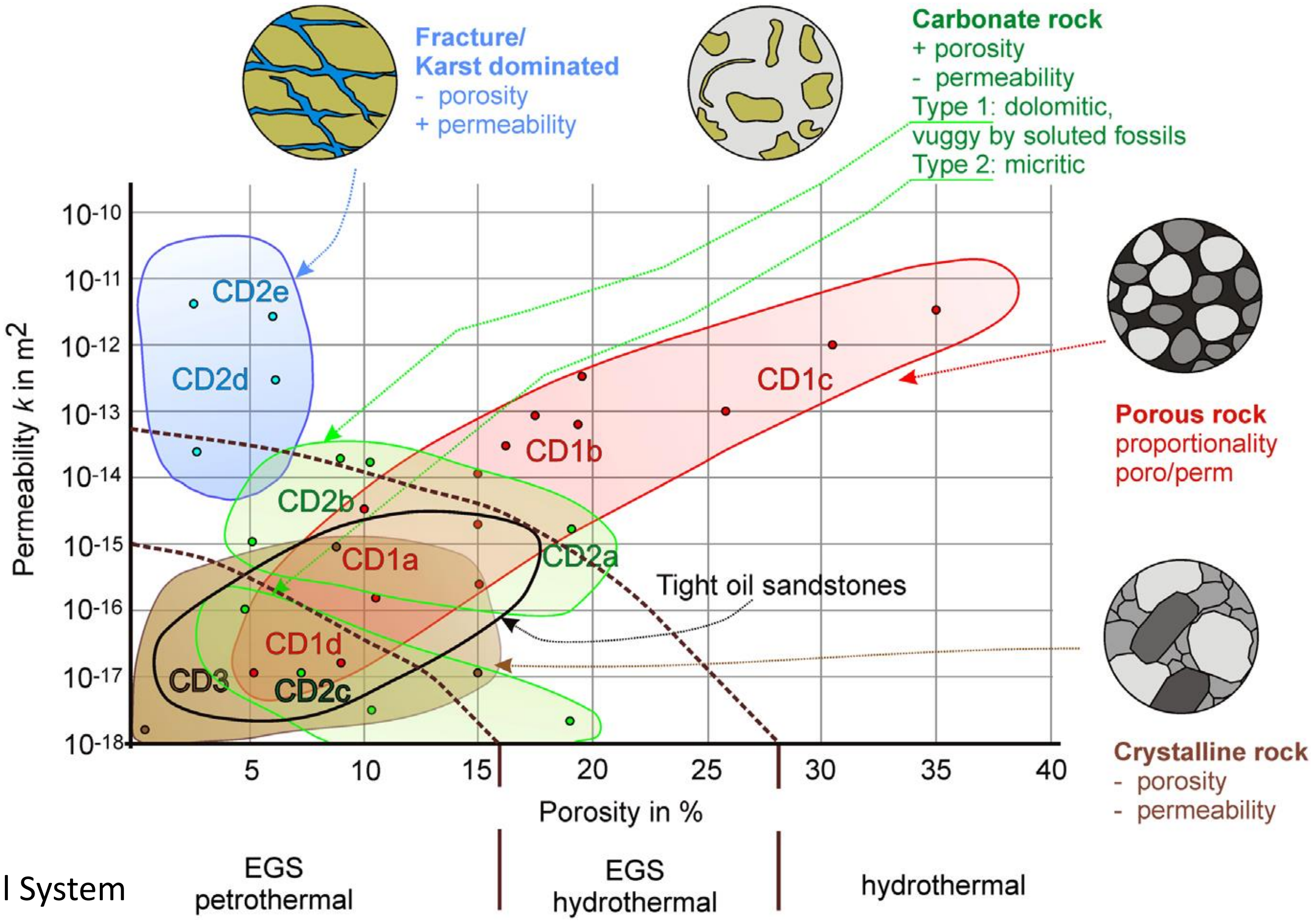
Geothermal system in orogenic belt and adjacent foreland basins: Red lines: Schematic isotherm distribution, recharge locations, fault geometry and basin geometry. Blue lines: Water flow lines result from heat advection and topography controlled hydraulic head (blue arrows).

- In the adjacent mountain belt, groundwater flow and thermal gradient are strongly influenced by large hydraulic heads resulting from the pronounced topographic relief.
- Thermal highs occur underneath high mountains and thermal lows beneath the valleys, resulting in varying local thermal gradients, due to meteoric water circulation. Beneath high mountains at about 15–20°C and beneath deep valleys 30–50°C.

Porosity and permeability relation of different geothermal reservoirs



EGS: Enhanced Geothermal System

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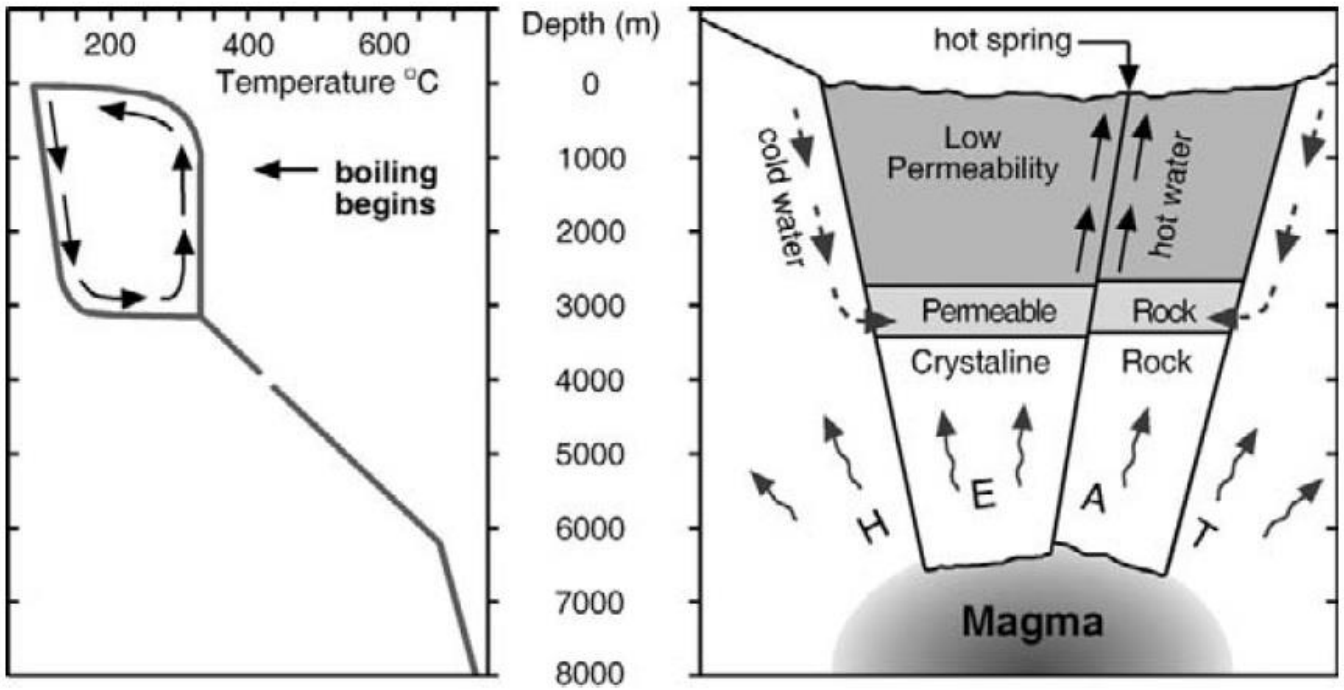
Porosity and permeability relation of different geothermal reservoirs with data points:

- Carbonate rock with partly dissolved biogenic content and filled pore space due to secondary dolomitization has a high porosity versus a low permeability (type CD2a and b), while karstic and highly fractured carbonate rock, such as reef formations, have a low porosity versus high permeability (type CD2e).
- Sandstone formations are characterized by a proportional ratio of porosity to permeability, controlled by depth.

Geothermal Reservoir

- Hydrothermal convective systems are geothermal systems with high temperatures and usually with surface activity.
- Natural flows play a dominant role in establishing the state of the fluid within the reservoir, and an understanding of them provides information about reservoir parameters such as vertical permeability.
- In low-*T* systems the reservoir fluid is always liquid water, while in higher-*T* systems, steam can also be present. The dominant mobile phase (liquid or steam) controls the pressure distribution, although the other phase may be present in significant amounts.



Model of the large-scale circulation of fluid in the natural state of a geothermal system.

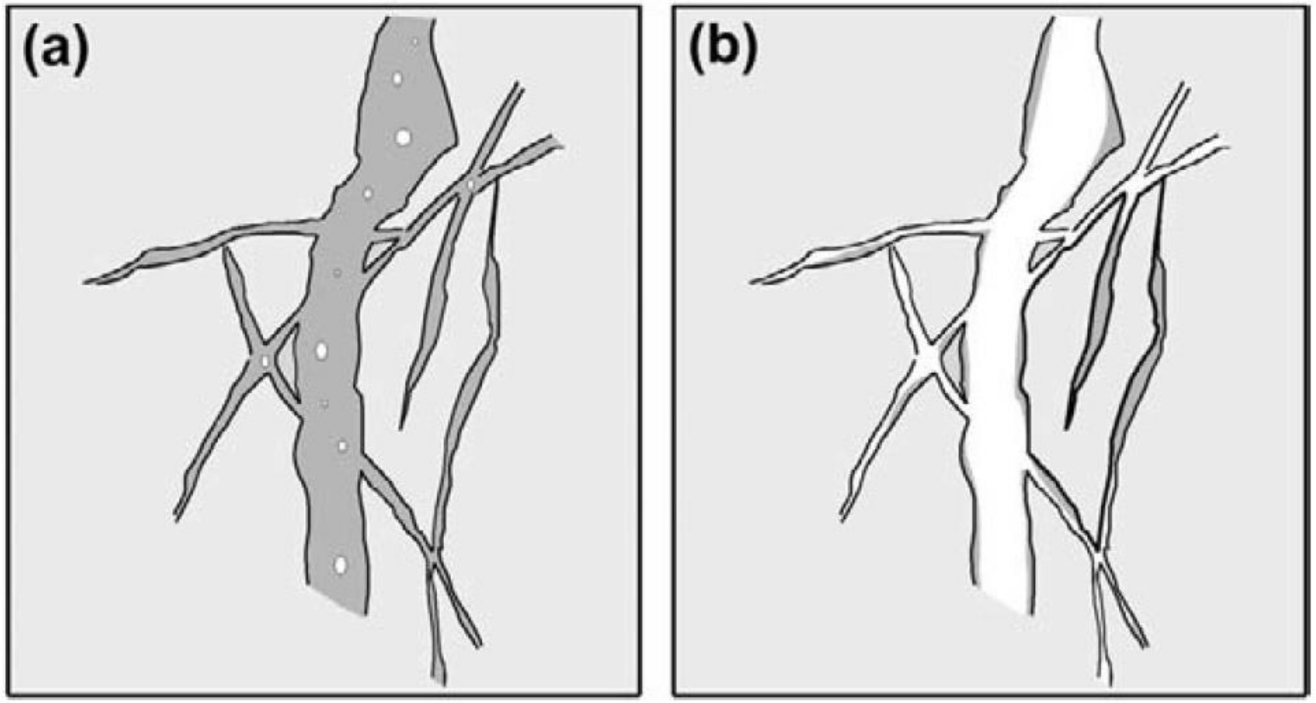
- The water is heated at depth, probably by close contact with some magmatic body, to the high temperatures encountered in reservoirs associated with such systems.
- Long lifetimes of geothermal systems cannot be sustained by a single emplacement of magma (thermal disturbance lasts only about 10,000 years). The magma source itself must be convecting, so molten rock remains near the zone, where heat is exchanged between the magma and the fluid in the geothermal system.

System with vertical upflow

- The upflow at great depth consists of water or supercritical fluid. As this fluid ascends, the P decreases, and at some point the ascending fluid forms two phases, gas and liquid, which both rise to the surface.
- From conservation of mass and energy we have to assume that the fluid at depth is liquid and boils when it reaches its saturation pressure.

Below the boiling level, the temperature distribution is given by: $T = T_b$ (constant base temperature).
Above the boiling level, T is given by the saturation relation: $T = T_{sat}(P)$.
The pressure gradient at any depth is equal to the local hydrostatic gradient plus the dynamic gradient caused by the upflow (generally less than 10% of the static gradient).
Ignoring the dynamic gradient, the boiling pressure for depth (BPD) is given by:

$$\frac{dP}{dz} = \rho_w \, g$$

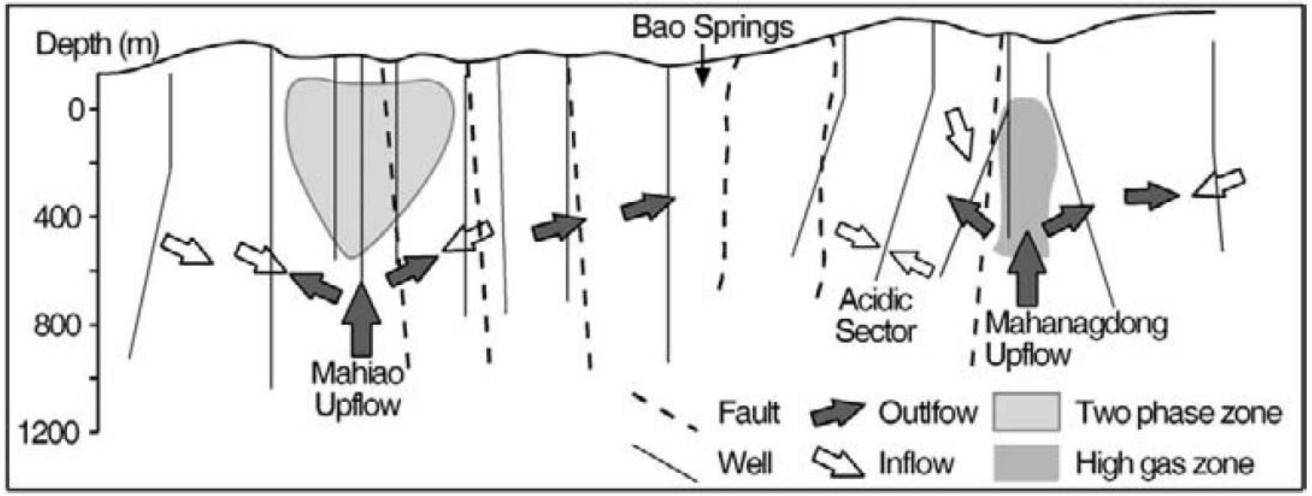


- The BPD approximates the P and T in the reservoir (T is at saturation for the local P) and also specifies that little mobile steam is present in the boiling zone.
- The rising fluid is cooled by dilution and conduction as much as it would be by boiling. Boiling conditions will be found in the core of the upflow, with cooler conditions toward the margins.
- If non-condensable gases and salts are present in the reservoir fluid boiling starts deeper (P at which the liquid phase first boils is greater).

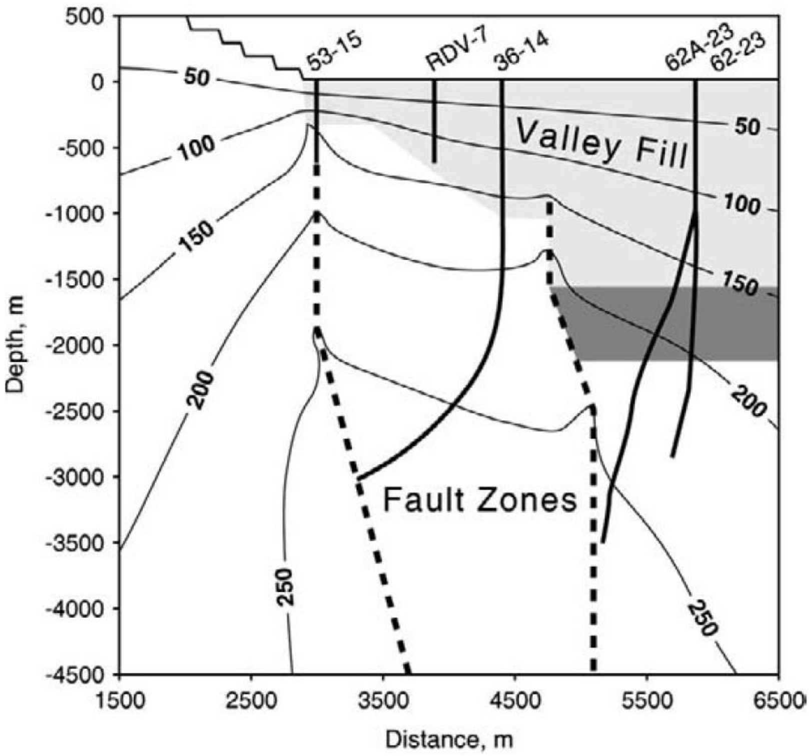
Distribution of liquid (dark shading) and steam (no shading) in liquid-dominated and vapor-dominated reservoirs.

System with lateral outflow

- Lateral flow generally occurs on occount of permeability variations and topographic effects, but boiling conditions can only be maintained if there is some continued upflow of fluid.
- If the natural flow is horizontal or turns downward, boiling ceases and liquid reservoir conditions are encountered.
- Within the reservoirs are present upflow regions, where boiling conditions should be expected, and outflow regions, where high-temperature liquid conditions with temperature inversions (declining temperatures with depth) can be expected.



Conceptual model of Greater Tongonan (Philippines)

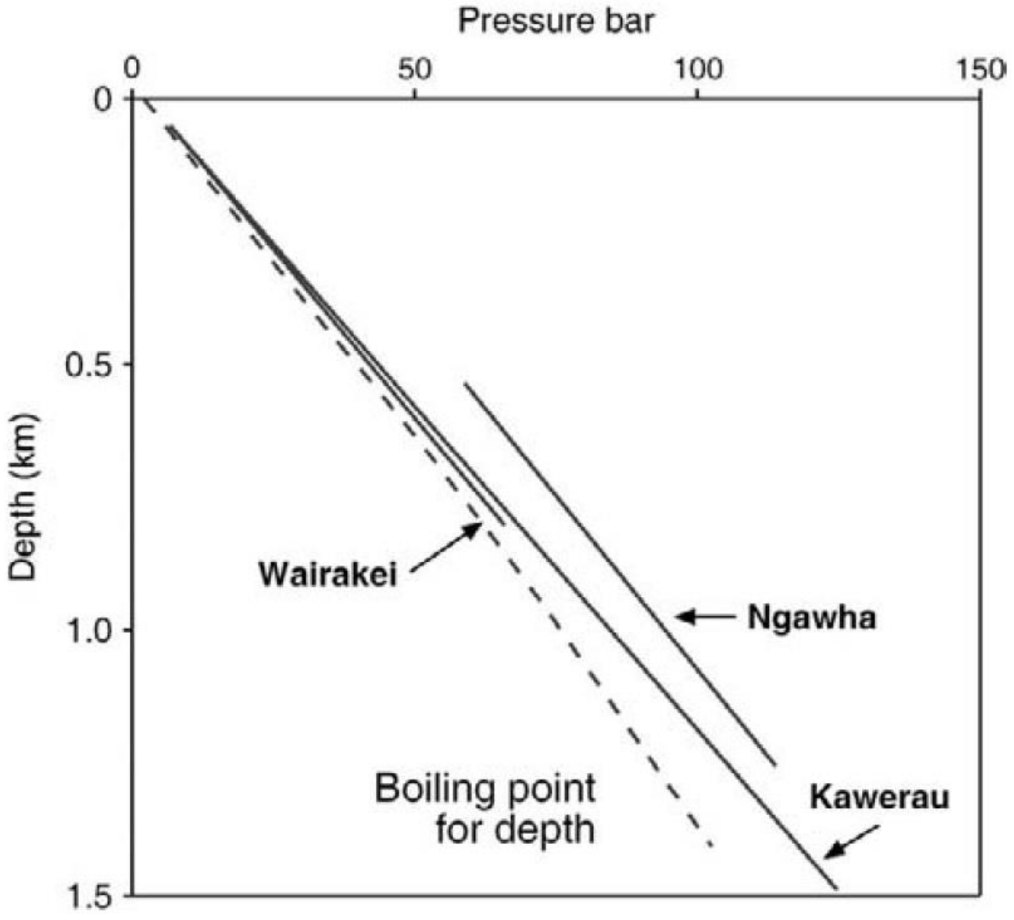


Conceptual model of Dixie Valley (USA)

- In the Tongonan Systhem water flowed laterally away from the upflow area to ultimately discharge at the Bao Springs and elsewhere.
- Dixie Valley is a high-temperature field in the Basin and Range province associated with permeability developed along a fault zone that provides the primary source of deep recharge

Pressure distribution

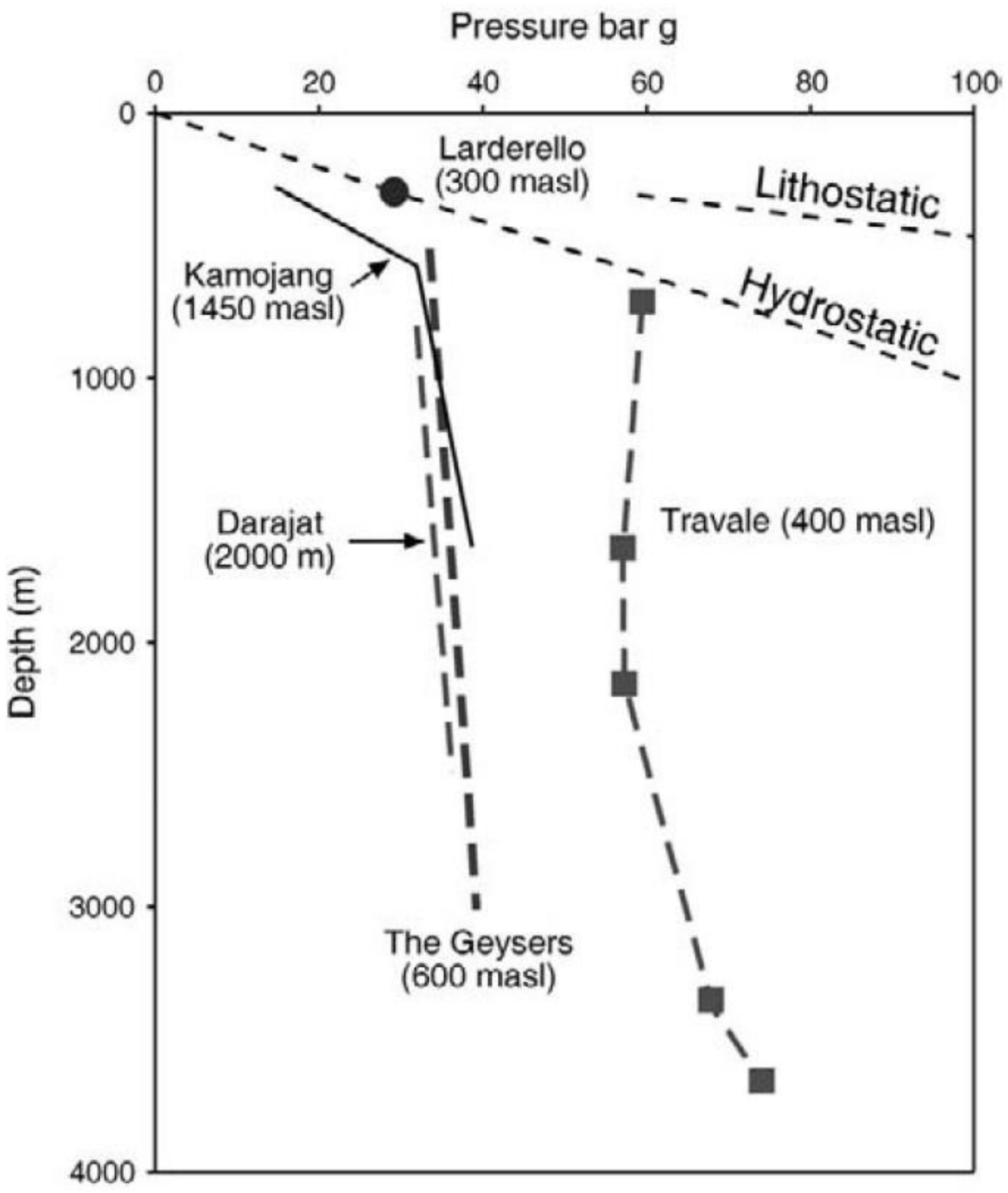
- The *T* profiles measured in exploration wells, together with geological information, provide a guide to the permeability structure of the reservoir, but *P* distribution gives more definite and direct information about how different parts of the reservoir are interconnected.
- A pressure difference between two geothermal reservoirs at different depths that is significantly different from hydrostatic does not necessarily imply that the two reservoirs are not connected by permeable structures: since the natural state of a geothermal reservoir is dynamic, its pressure distribution is as well, and pressure differentials may be caused by the natural flow, temperature difference, or some combination of these factors.



Pressure distribution with depth in three New Zealand geothermal fields.

- At Wairakei and Kawerau pressures deep in the reservoir are overpressured compared to surface (for a hydrostatic column allowing for *T*). This overpressure is caused by the natural upflow through the reservoir, not by any confining bed.
- Data from Ngawha show a confined geothermal field: The reservoir is overpressured with respect to ground surface, due to the confining layer rather than the dynamic pressure gradient due to vertical flow.

Vapor-dominated reservoir



Reservoir pressure in vapor-dominated systems (Travale is part of greater Larderello).

- A vapor-dominated reservoir appears to contain a static column of steam, and any associated surface thermal activity consists of steam or steam-heated water (examples fields: The Geysers; Lardarello; Kamojang; and Darajat).
- In a vapor-dominated reservoir *P* increases only slowly with depth, implying that a low-permeability boundary exists above the reservoir and around the sides of the reservoir, isolating the relatively low pressures inside the reservoir from regions outside the reservoir, where *P* near hydrostatic are expected.

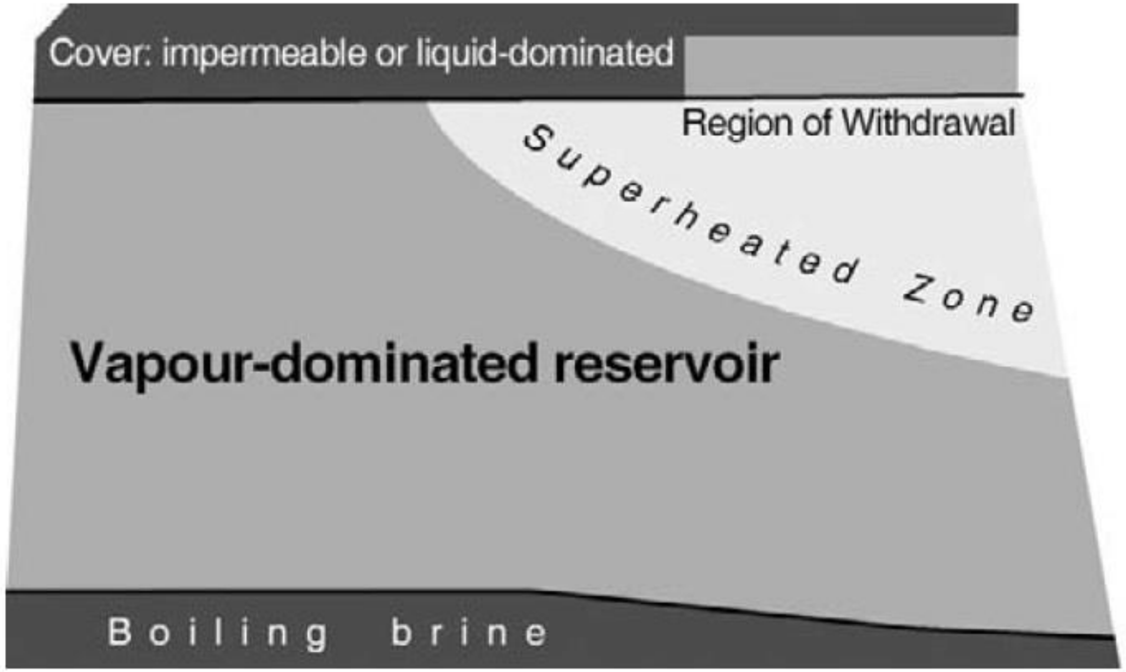
Model of vapor dominated reservoir:

1. In a steam-dominated reservoir, the counterflow system of steam moving up and water moving down controls the fluid distribution. The flowing steam occupies most of the fracture space, and water occupies the remaining pore space.
2. The steam spreads laterally through the reservoir and heat is lost through the caprock, and some of the steam condenses, absorbing some of the gas.
3. This condensate migrates down through the reservoir under gravity, wetting the matrix rock and increasing the permeability by dissolving some of the rock minerals.
4. Liquid-mobile species present in the steam are removed with condensate, and vapor-mobile species are concentrated in the remaining steam.

Vapor-dominated reservoir

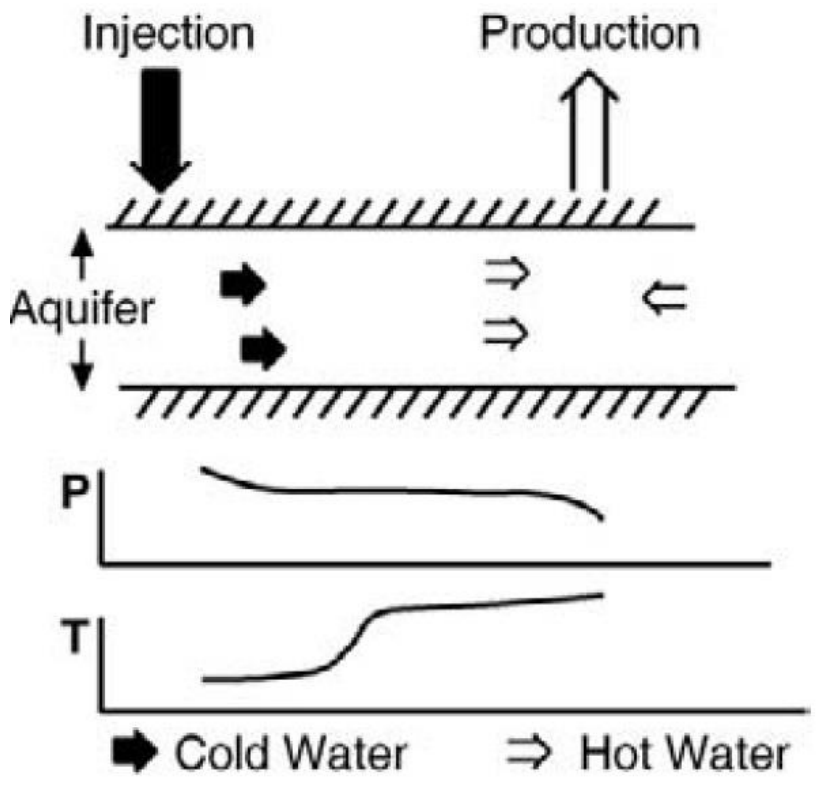
The essential components of a vapor-dominated reservoir are: (1) the stored steam and immobile water, (2) the heat stored in the rock, (3) an overlying condensate layer, and (4) a possible deep zone of boiling brine.

- Under exploitation, the immobile water is gradually evaporated and the vapor-dominated reservoir becomes depleted of water, eventually forming a dry (superheated) zone.
- As production continues, the superheated zone around the area of exploitation expands into the (saturated) vapor-dominated region.



Liquid-dominated reservoir

- If the reservoir is at a fairly uniform *T*, the flow is isothermal. If there is a distribution of *T* or if there is reinjection of cooled fluids, it is necessary to compute the motion of thermal changes along streamlines.
- The reservoir may be unconfined rather than confined, in this case, with the possible entry of surface groundwaters, or with reinjection of cooled fluids, the efficiency of thermal sweep through the reservoir becomes important.



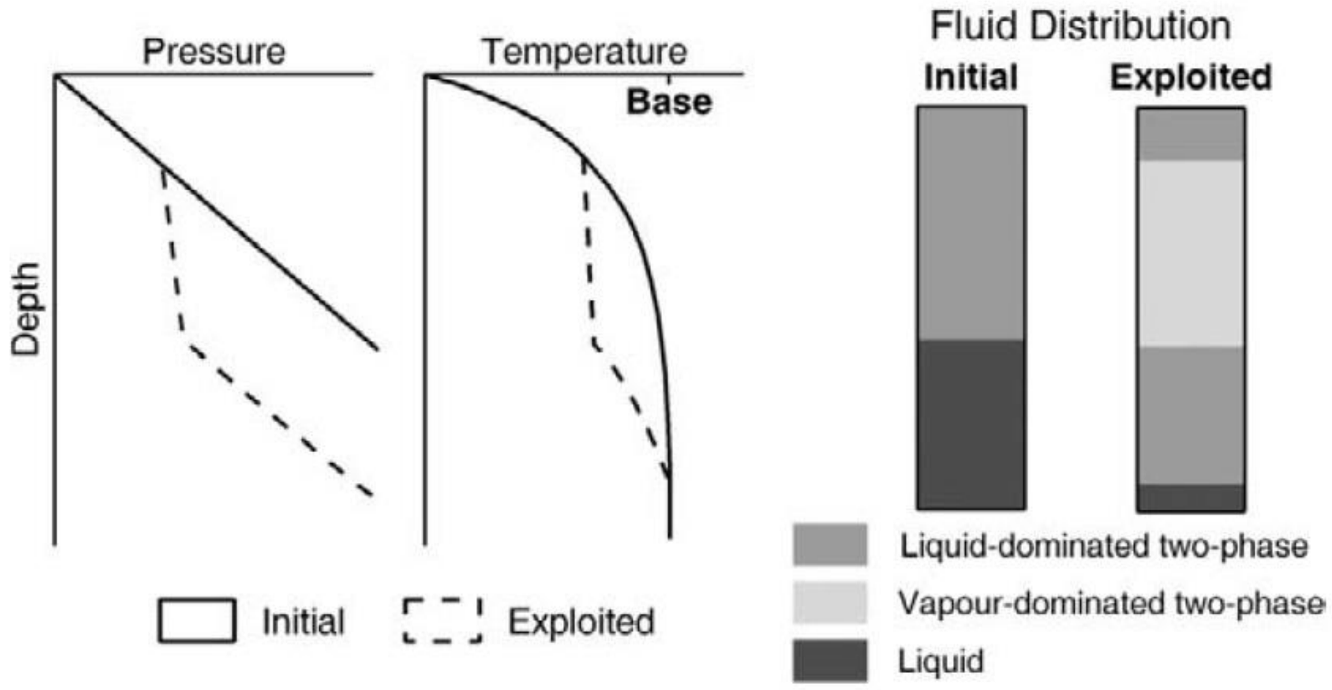
Flow of liquid in an exploited liquid reservoir.

Liquid-dominated reservoir: changes by exploitation

In high-temperature geothermal reservoirs, a decline in pressure caused by exploitation may initiate boiling in part or all of the reservoir, producing changes in the steam/water ratio, as well as *P* and *T* changes.

1. In its initial state there was a near-hydrostatic pressure profile; 2. With production, pressure in the reservoir falls, and above the level of production the vertical gradient is less than hydrostatic; 3. Water will now drain downward, causing a decrease of liquid saturation and favouring mobility of steam, which drains upward; 4. In the upper part of the underlying liquid dominated region, boiling occurs as the pressure declines, and steam is formed; 5. This steam also drains upward, forming a distinct vapor-dominated region at the top (a steam cup), limited above by a layer of lower permeability.

- The net heat loss of the reservoir is the cooling of the rock as a result of water boiling and the advance of peripheral cooler waters.

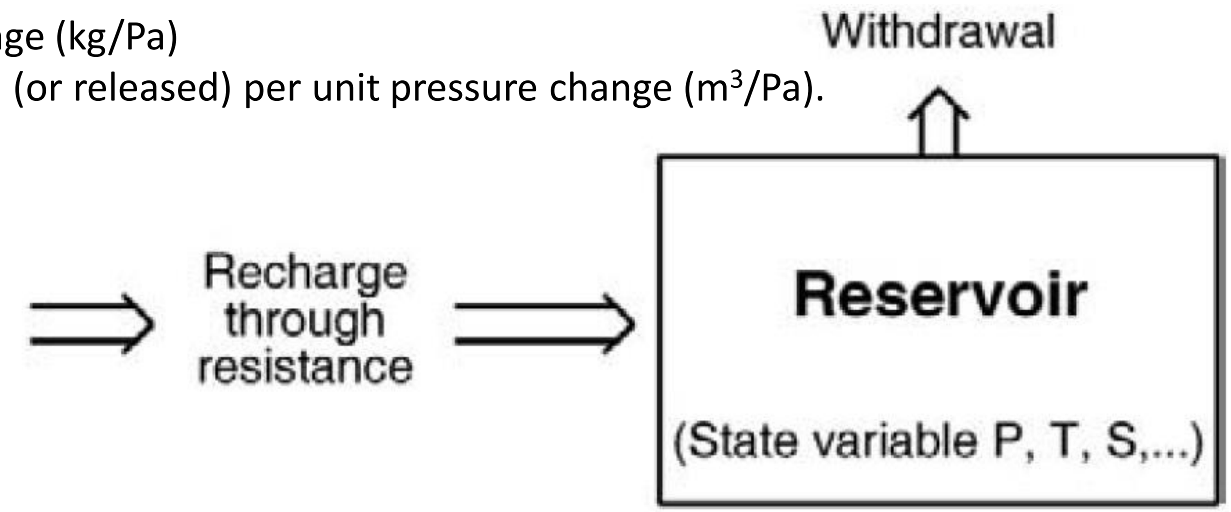


Fluid distribution in natural and exploited states of a liquid-dominated reservoir.

Geothermal Reservoir

- The reservoir is described as a single box containing homogeneous rock and fluid, which is withdrawn from the box, and the box is recharged from external sources (laterally adjacent aquifers, or overlying groundwater).

W= Mass flow withdrawn from the box (kg/s)
W_r= Mass flow recharge (kg/s)
S_M=Mass that is stored or released per unit pressure change (kg/Pa)
S_V="storage coefficient," the volume of fluid that is stored (or released) per unit pressure change (m^3/Pa).
α=recharge coefficient, kg/Pa s



$$S_V = V\varphi c$$

c=compressibility (Pa^{-1})
ϕ=porosity

If the box undergoes a *P* change ΔP, it expels a fluid mass S_M ΔP. Equating this to the net fluid loss (W e W_r)Δt gives conservation of mass for the block:

$$S_M \frac{dp}{dt} + W - W_r = 0$$

Recharge is proportional to the pressure difference between the box and the recharge source:

$$W_r = \alpha(P_o - P)$$

$$W = \alpha(P_o - P) - S_M \frac{dP}{dt}$$

Geothermal Reservoir

The response to a discharge at constant rate *W*, beginning at time t = 0, is an exponential approach to equilibrium:

$$P_o - P = \frac{W}{\alpha}\left[1 - e^{-\frac{t}{\tau}}\right] \qquad \tau = S_M/\alpha$$

τ = system time constant or relaxation time

For short times, pressure falls linearly with time: $$P_o - P = Wt/S_M = qt/S_V$$

q=flow rate (m^3/s) W = ρq (kg/s)

Within this time scale, recharge has negligible effect on the *P*. The recharge, if any, cannot be determined until there are observations over a period of time comparable with the system time constant τ.

For time values significantly in excess of the relaxation time, the pressure stabilizes at a value determined by a balance with the recharge:

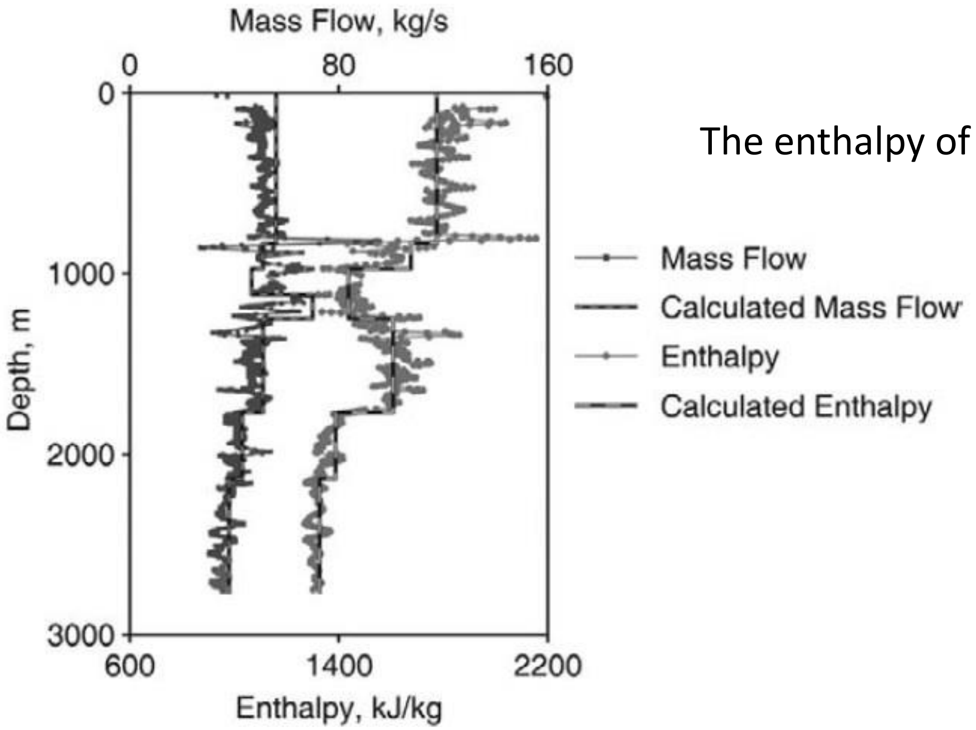
$$P_0 - P \approx W/\alpha \quad t >> \tau$$

This balance is independent of the storage coefficient ($\boldsymbol{S_v}$), fluid withdrawn is replaced by the recharge.

Well models

In the exploration stage, we must have an overview of the resource in its natural (undisturbed) state, the extent of the resource, and the distribution of *T*, *P*, and rock characteristics (permeability, porosity, fracture patterns, etc.) using the following parameters:

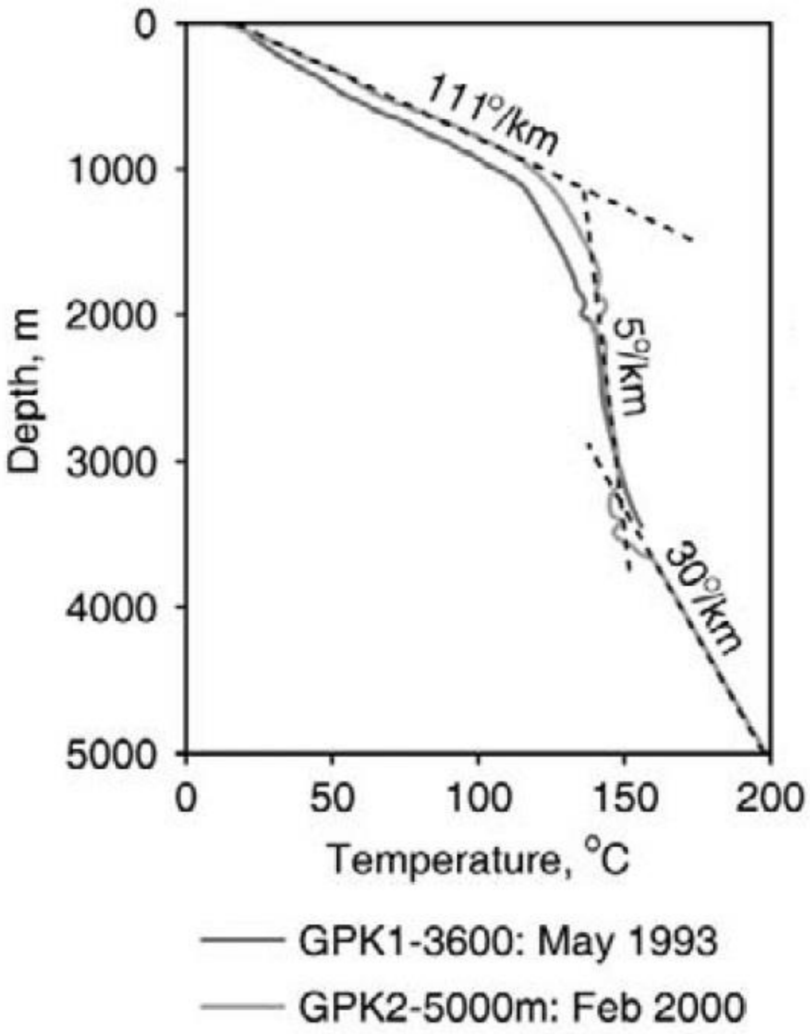
- Temperature distribution
- Pressure distribution
- Permeability distribution
- Reservoir state (liquid/vapor dominated, single-phase/two-phase)
- Fluid chemistry
- Non-condensable gas content



The enthalpy of the flowing mixture is computed from the pressure gradient and flow rate.

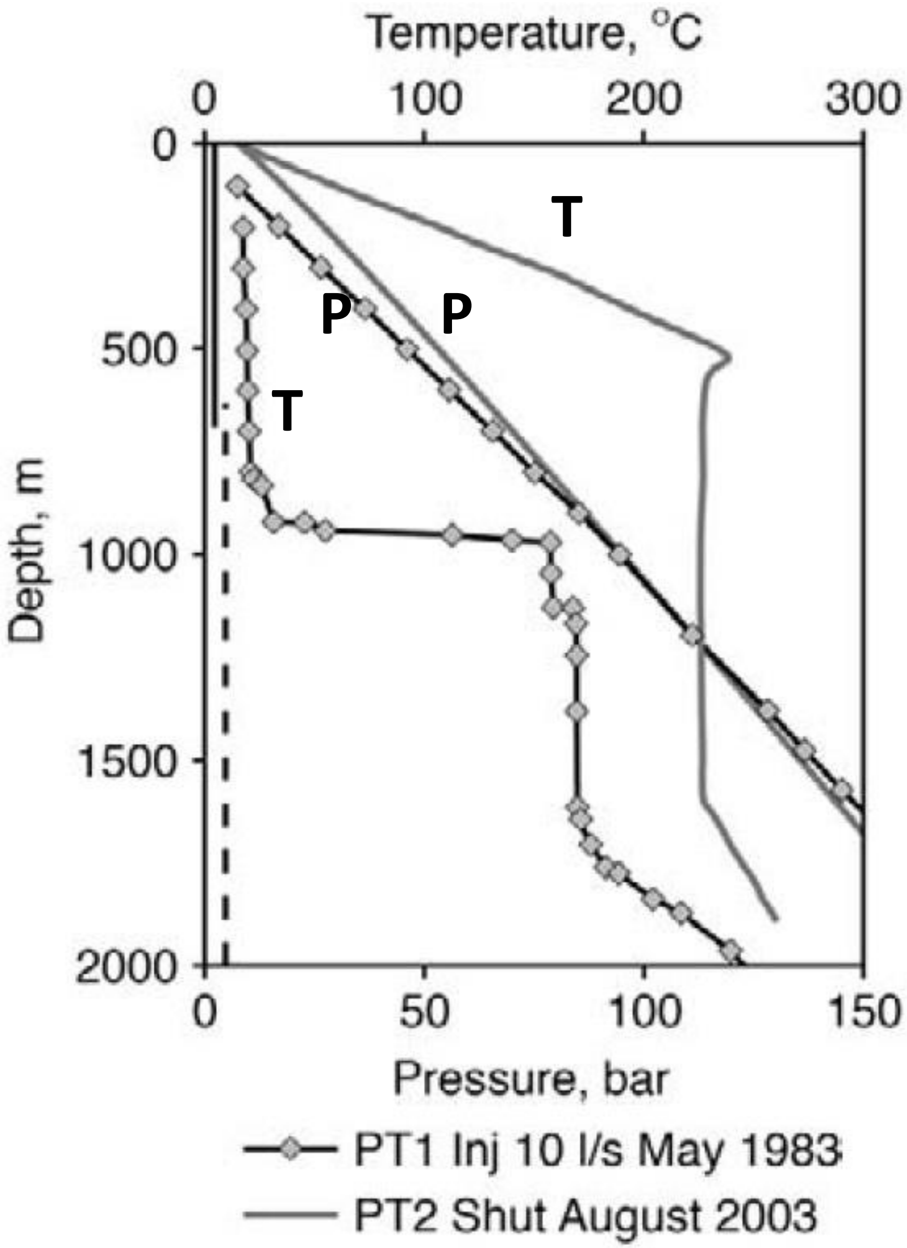
Temperature profile in a well

- When rock is impermeable, heat is transported by conduction (temperature increases linearly with depth).
- Once there is some permeability in the rock, the fluid motion controls the temperature distribution (convection transport).
- Convective profiles can take a considerable variety of forms, with isothermal sections, inversions, boiling sections, and mixtures of all of these.



1. The first kilometer has a high gradient and linear profile, indicating conductive transport.
2. From 1 km to 3.3 km there is a much lower gradient, which is attributed to a convective system along faults and fissure zones.
3. Below 3.3 km there is again a high linear gradient, indicating conductive heat transport and consequently lower permeability in the surrounding formations.

Temperature profile in a well

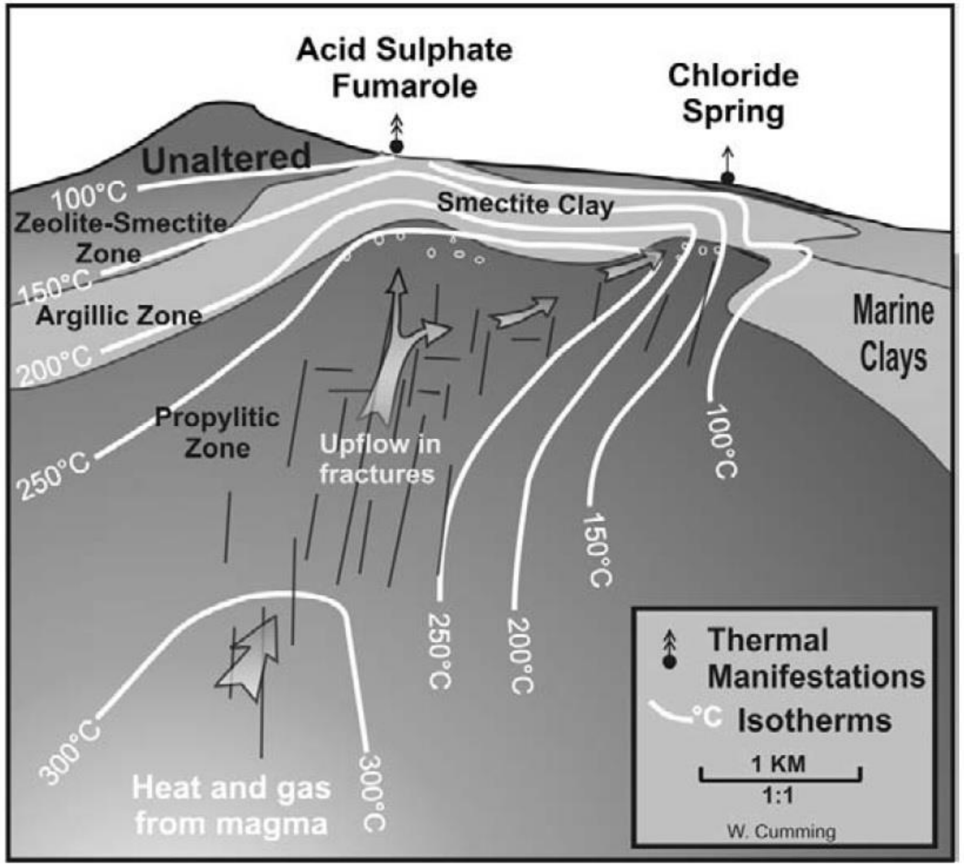


Temperature and Pressure profiles change during injection and shut conditions:

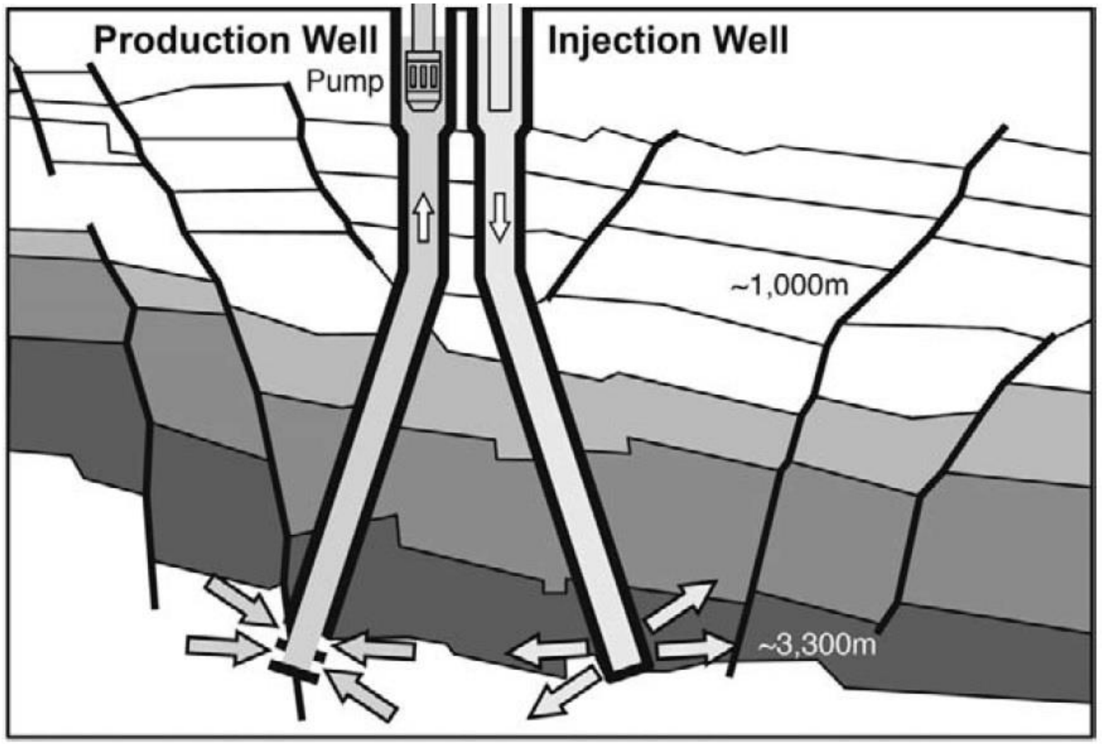
- During injection there are 1. inflows to the well at 960 m; 2. At the inflow depth, the *T* increases rapidly over a short interval as the cold water being injected into the well mixes with the hot inflow; 3. The mixture flows down the well and exits at a lower feed zone at 1600 m.
- In the shut-in condition, the temperature profile reflects conduction and convection conditions of the reservoir.
- The pressure profiles show that during injection, the *P* at 960 m is less than the stable shut *P* and more than stable shut *P* at 1600 m, confirming that during injection there is inflow at the upper zone and outflow at the lower zone.

Conceptual model of a Geothermal System

- The characteristic pattern of steam-heated features at higher altitude and chloride springs at lower altitude identifies the point where the liquid upflow reaches the ground surface at elevations lower than the steam discharges.
- In the other example the reservoir is defined as the aquifer, and the geology that defines its depth and the increase of temperature with depth.



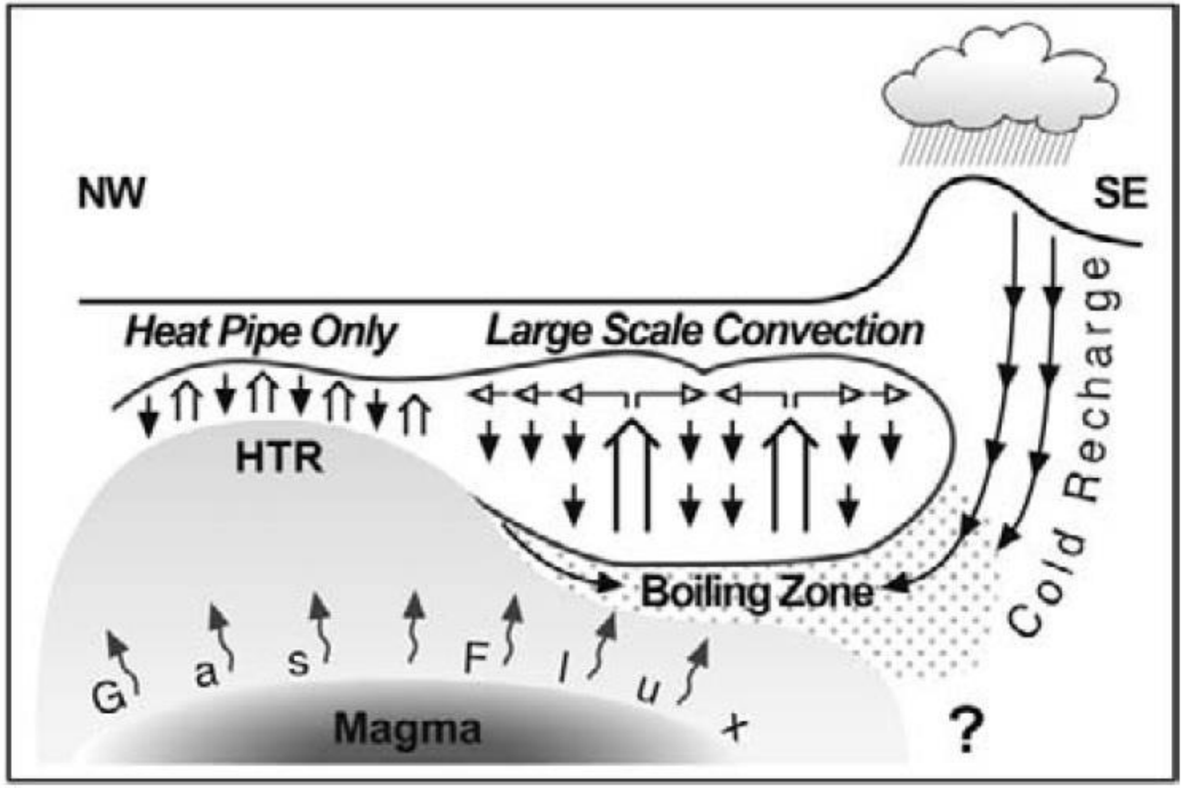
High-temperature system



Low-temperature system

Geothermal Systems: The Geysers

- The Geysers geothermal field in northern California is the largest electric power producing field in the world.
- The reservoir is vapor-dominated and no deep-water level is present beneath the steam reservoir.
- In the NW, there is at depth the high temperature reservoir (HTR), where *T* reach 345°C in superheated steam, while the water is absent.
- In the SE, steam rises from a conjectured deep boiling zone of liquid water.
- Within the reservoir, there is an upflow of steam and return downflow of condensate.



Main Properties of the Reservoirs

The main properties that have to be mapped are:

1. The reservoir geology
2. Surface and downhole geophysics
3. Reservoir temperature
4. Reservoir pressure
5. Permeability distribution
6. Zonation within the reservoir: liquid, two-phase, or steam zones
7. Fluid chemistry
8. Natural discharges
9. Hydrothermal alteration
10. Well discharges
11. Surface deformation

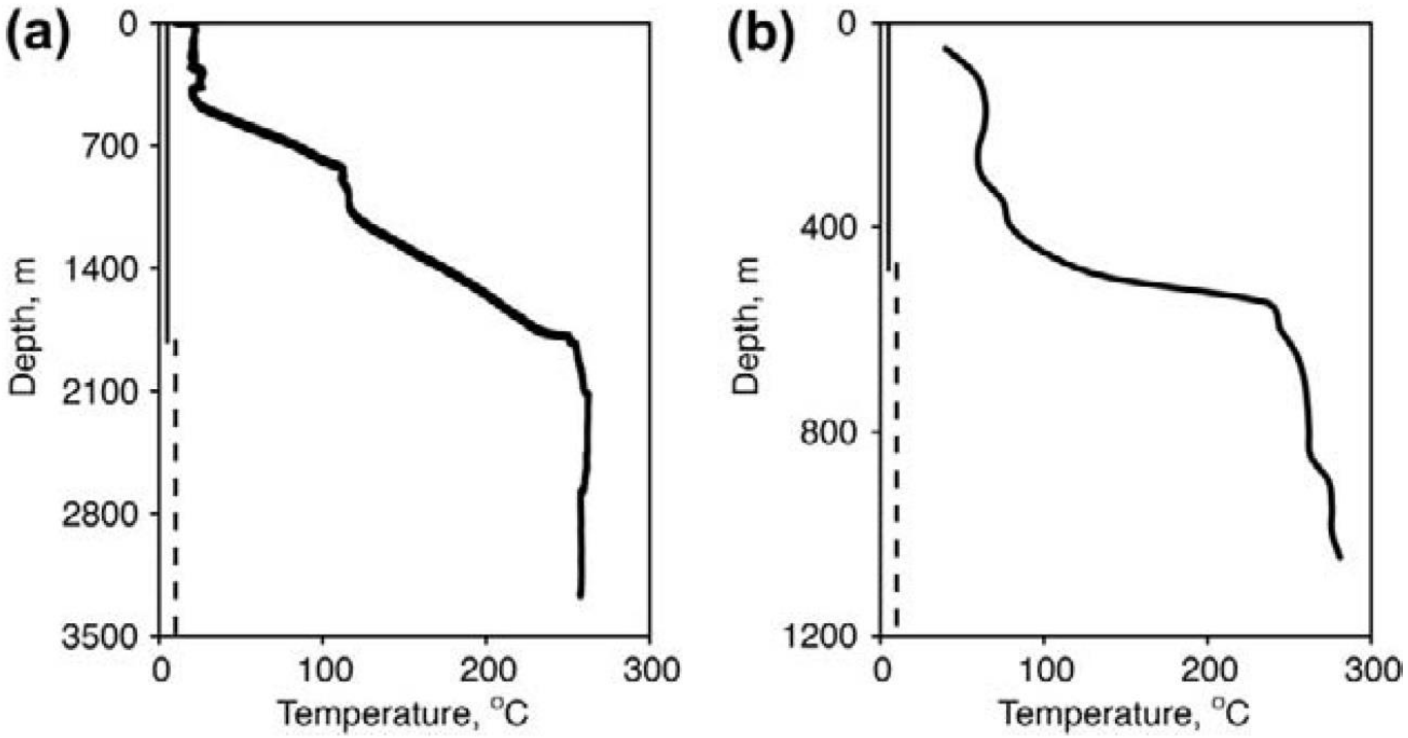
- Temperature gradient provides information on the natural flow: 1. A temperature reversal implies a flow of colder water entering the reservoir. 2. A zone of constant temperature implies convective mixing of fluid. In all cases, maximum or minimum zonal temperatures imply some flow and thus some permeability.

Temperature Profiles

- The upper layers above the high-temperature reservoir may be quite cold, while in the area of surface discharge hot or boiling conditions may extend to the surface.
- Two forms of temperature profiles are often found in such regions: a linear conductive gradient indicating poor permeability and roughly isothermal cold *T*, or large *T* inversions indicating cold aquifers.

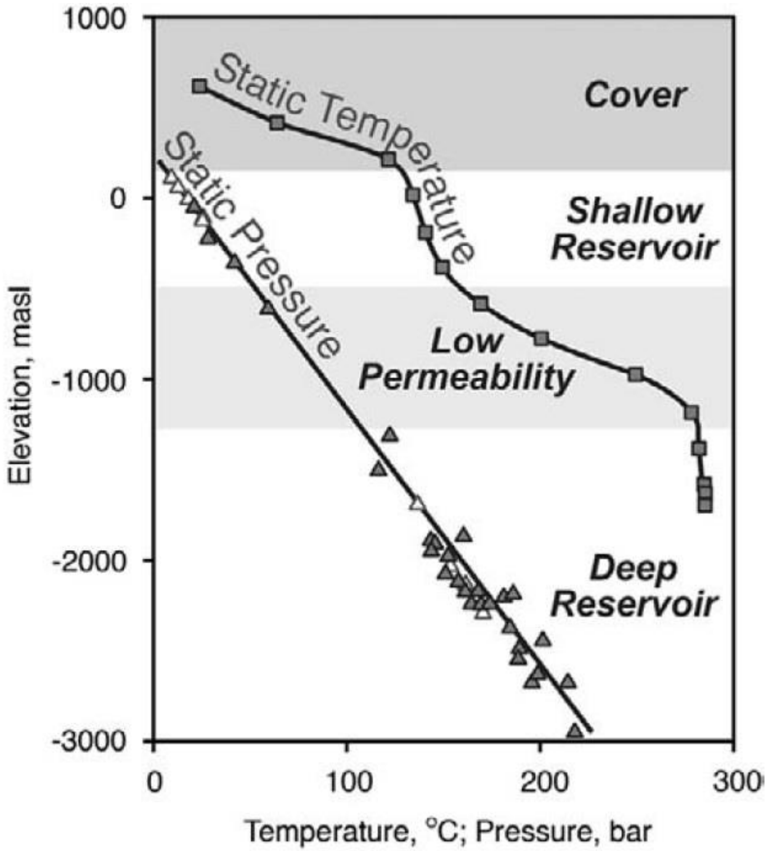
Example (a): The reservoir top is at 1800 m, below which there is little temperature change in the convecting reservoir. Above this depth, there is a region of roughly linear gradient to 500 m, interrupted by an aquifer at 1000 m. Above 500 m are cold aquifers.

Example (b): The reservoir top is at 550 m. The steep gradient between 400 and 550 m implies low permeability at this depth, whereas above 400 m there must be fluid movement in relatively cool rock.



Pressure-Temperature Profiles

- Permeability of a reservoir can be sufficient for convection, but not enough for production.
- The *T* profile shows that the shallow and deep reservoir of the Mt. Amiata system are separated by a conductive zone between the two convective temperature zones. However, there is, within the data scatter, a common pressure gradient implying a hydrologic connection.
- Therefore, it should be determined how low the permeability must be to show the conductive gradient, and how high it must be to create the common pressure gradient.



P-T profile in Mt. Amiata system

Mapping Geothermal Reservoirs

Geothermal reservoirs can be mapped and monitored using electromagnetic data (EM):

- Geothermal waters have high concentrations of dissolved salts, that result in conducting electrolytes within a rock matrix. As a result, there is a large reduction in the bulk resistivity to increasing temperatures.
- The resulting resistivity is also related to the presence of clay minerals, and can be reduced considerably when clay minerals are broadly distributed.
- Although water-dominated geothermal systems have an associated low resistivity signature, the opposite is not true, and thus we need to include geological and geophysical data.
- Due to theoretical relations between the electric resistivity, on the one hand, and temperature, porosity, and permeability on the other hand, the former is often used for the indirect estimation of these parameters.

The resistivity of a solid phase correlates with *T*:

$$\rho = \rho_0 e^{E/kT}$$

where ρ_0 is the resistivity at a theoretically infinite T, E is an activation energy, k is a Boltzman constant, and T is an absolute temperature (K).

$$\rho_T = \rho_0 / (1 + \alpha(T - T_0))(1 + \beta(T - T_0))$$

(for altered basalts)

where ρ_T and ρ_0 are the resistivities at temperature T and reference temperature T_0. Empirical constants α and β are determined at T_0

Other empirical relations exist between the resistivity of rocks and their porosity:

$$\sigma_r = (1/F)\sigma_w + \sigma_s \quad \text{where} \quad F = \phi^{-m}$$

m is the cementation factor often taken equal to 2

σ_r is the bulk electric conductivity of a rock, σ_W is the electric conductivity of a saturating fluid, F is the formation factor of the sample.

Surface conductivity σ_s is due to an extra conduction pathway via the electrical double layer that forms at the interface of the clay mineral and the water. It is negligeble for concentration of the ionic solution > 0.01 M.

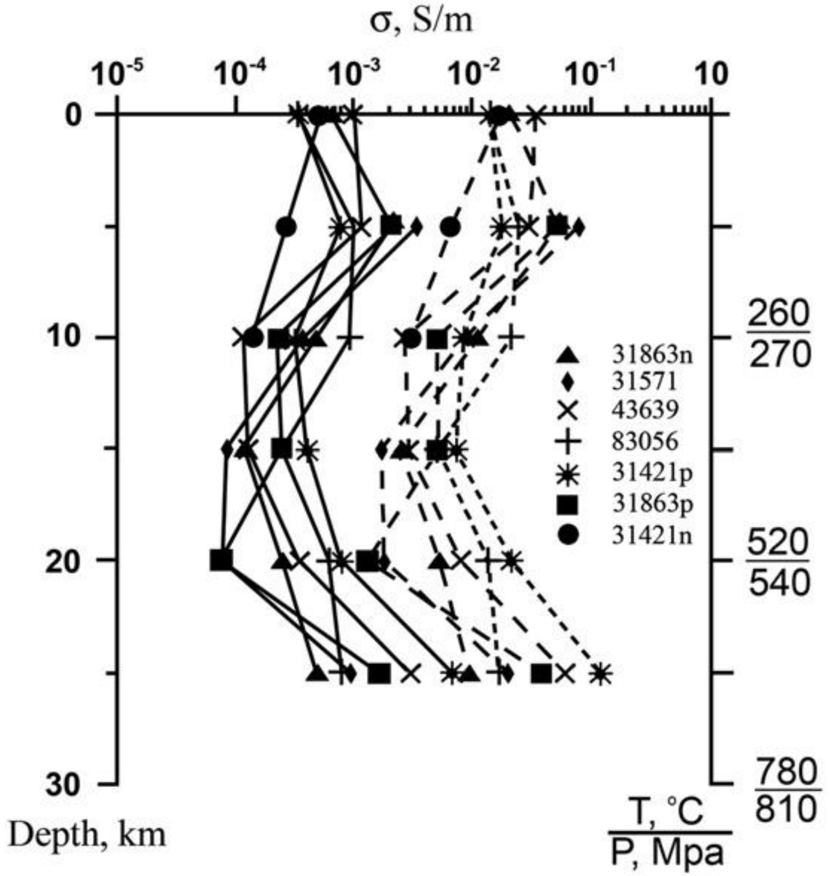
The hydrologic permeability k can be expressed as: $k = \alpha\phi^3$ where α has a value of 1×10^{-12} m^2

$$\sigma_r = \sigma_w \phi^2$$

$$\sigma_r = \sigma_w (k/\alpha)^{2/3}$$

Electric conductivity variations

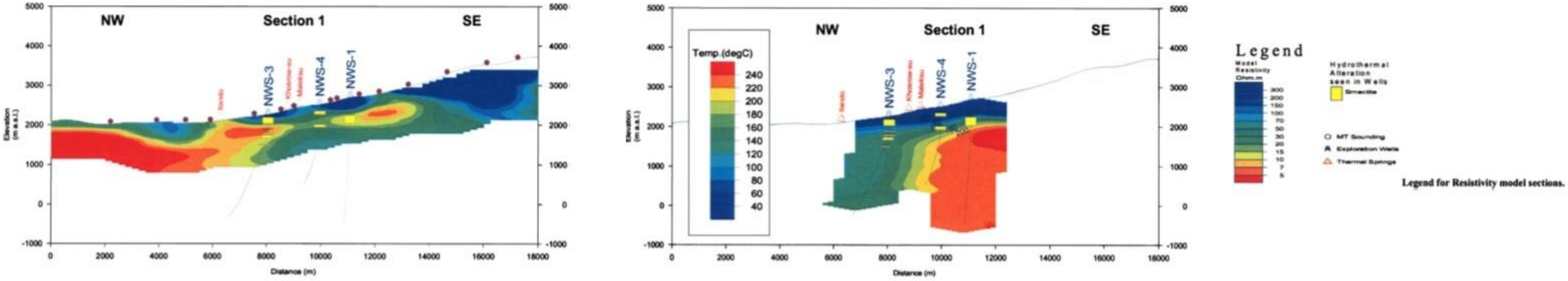
- Electric conductivity of a fluid increases as a function of *T* and *P*, whereas permeability can both decrease and increase with respect to these variables.
- At depths of more than 15 km the rock electric conductivity often decreases, since the permeability of rocks decreases due to compaction.
- Electric conductivity values of the order of 10^{-1} S/m can be reached in regions with high heat flows at depths of about 20 km where *T* higher than 500 °C occur.



Spichak and Manzella, 2009, Journal of Applied Geophysics, 68

Calculated values of electric conductivity of rocks with a fluid represented by a 0.1 M solution of NaCl (solid lines) and 3.3 M solution of KCl (dash lines) for regions with high heat flows. Symbols *p* and *n* at rock sample numbers indicate that the data on permeability are in the directions parallel and perpendicular to the bedding, respectively.

Resistivity vs Temperature



Spichak and Manzella, 2009, Journal of Applied Geophysics, 68

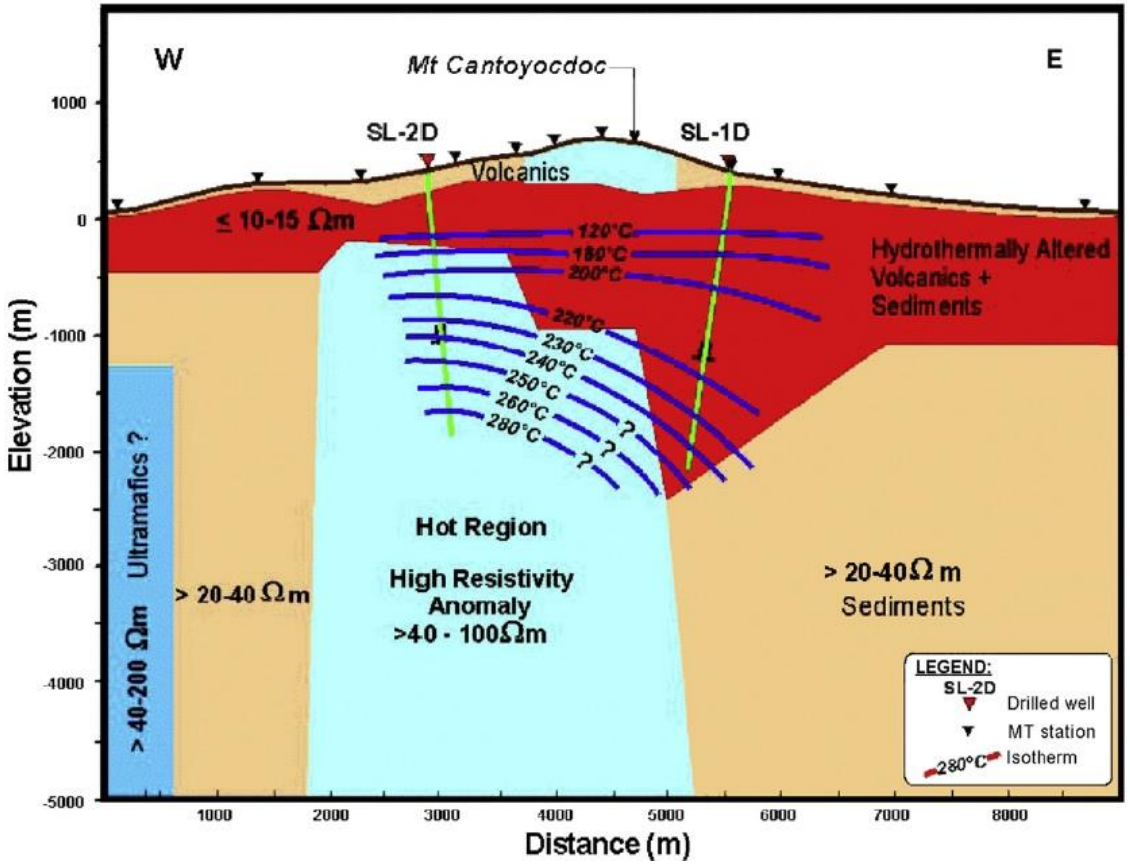
- Magnetotelluric (MT) sounding reveal anomalously low resistivity zones in the Takigami geothermal area (Japan), indicating a potential geothermal reservoir beneath the volcanic area.
- The low resistivity patterns often correlate with the distribution of smectite clay alteration products and are generally consistent with higher temperatures revealed from exploration wells.
- By comparing the resistivity distributions with the *T* distributions based on fluid flow calculations at a steady state, the validity of the location and dimensions of the estimated reservoirs could be confirmed.

Temperature is the major control of clay mineralogy:

- Low temperature clay minerals such as smectite and zeolites are electrically conductive and formed at *T*> 70 °C.
- At higher *T*, chlorite and/or illite may appear, the proportion of chlorite or illite increases with *T*, especially above 180 °C.
- Zeolites and smectite disappear at 220–240 °C and pure chlorite and/or illite usually appear at *T* higher than 240 °C.

Resistivity vs Temperature

- Geothermal areas located in andesitic volcano-sedimentary systems are often more resistive than surrounding sediments being characterized by high-T, but less conductive minerals (e.g., smectite, illite, epidote) present within the sediments.
- Correlation between low resistivity and fluid concentration is not always correct since alteration minerals produce comparable and often a greater reduction in resistivity.
- Low resistivity anomalies are not always suitable as geothermal targets (they should be accompanied by low-density anomalies.



Spichak and Manzella, 2009, J. Applied Geophysics, 68

MT resistivity structure correlated with lithology and T at Southern Leyte, Philippines.

Iceland Deep Drilling Project (IDDP)

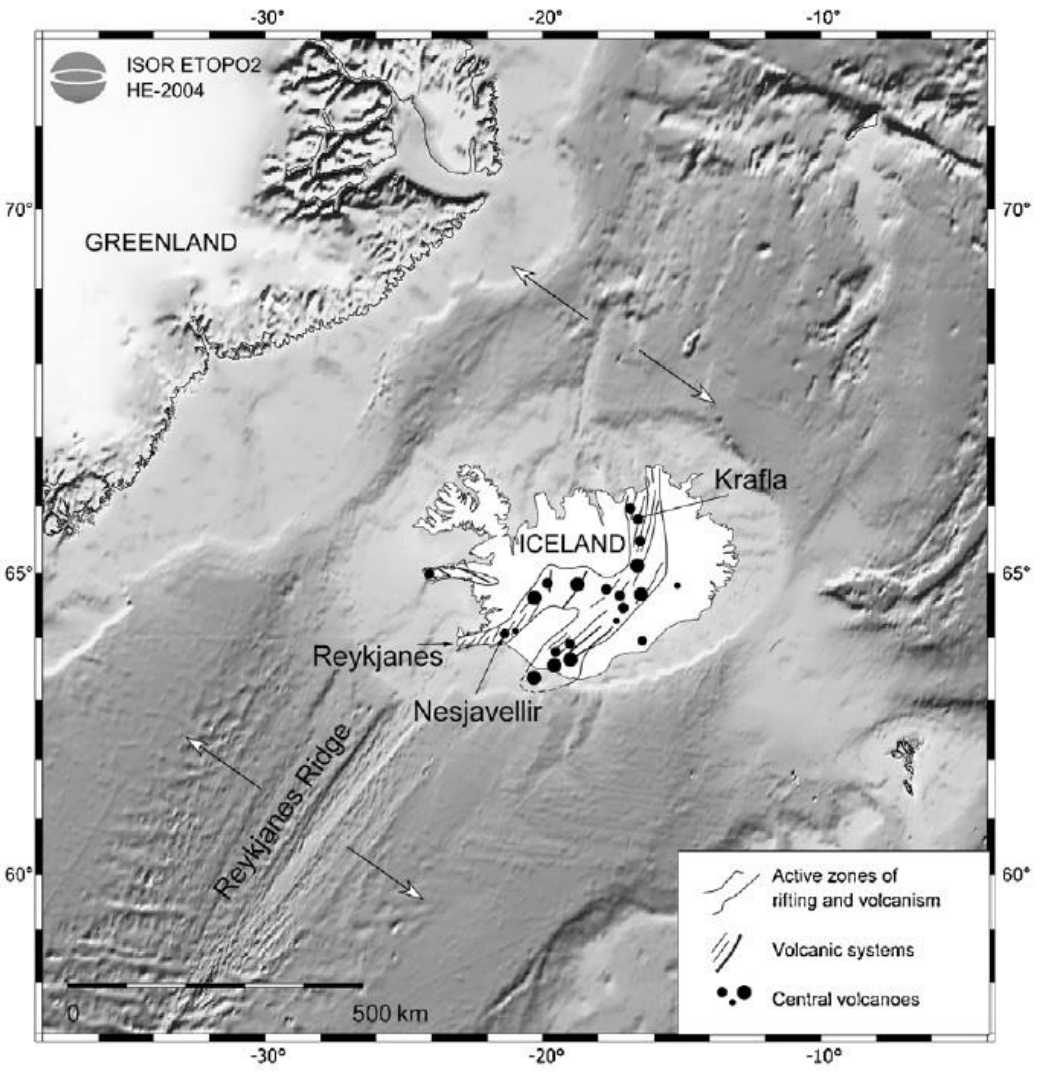
IDDP is a long term program by an industry-government consortium established in 2000, which aims at investigating unconventional, very high-temperature, geothermal systems, to improve the economics of geothermal resources, minimize the environmental impact of harnessing geothermal reservoirs, evaluate 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 25 MPa 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?



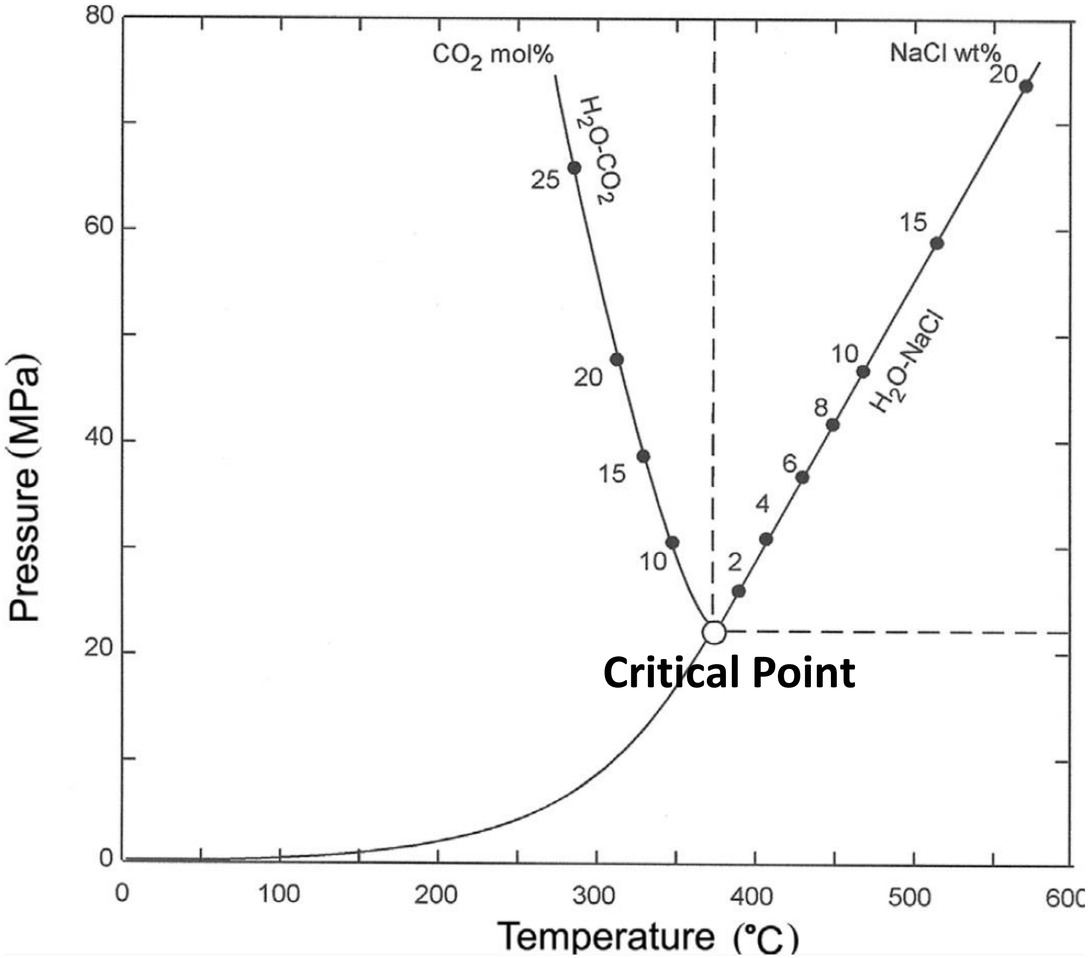
Friðleifsson et al., 2014, Geothermics, 49

- 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

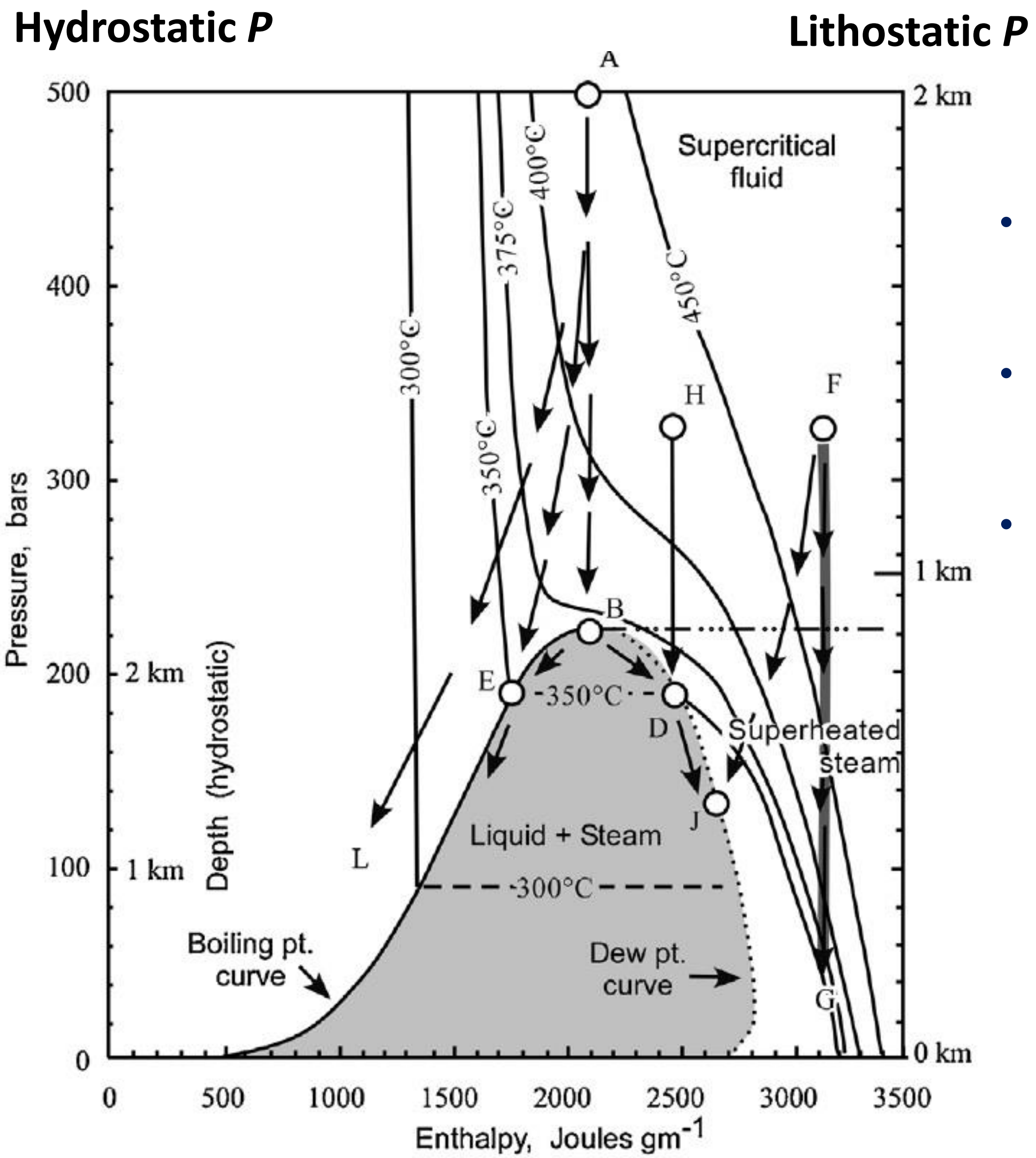


Elders et al., 2014, Geothermics, 49

- 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)

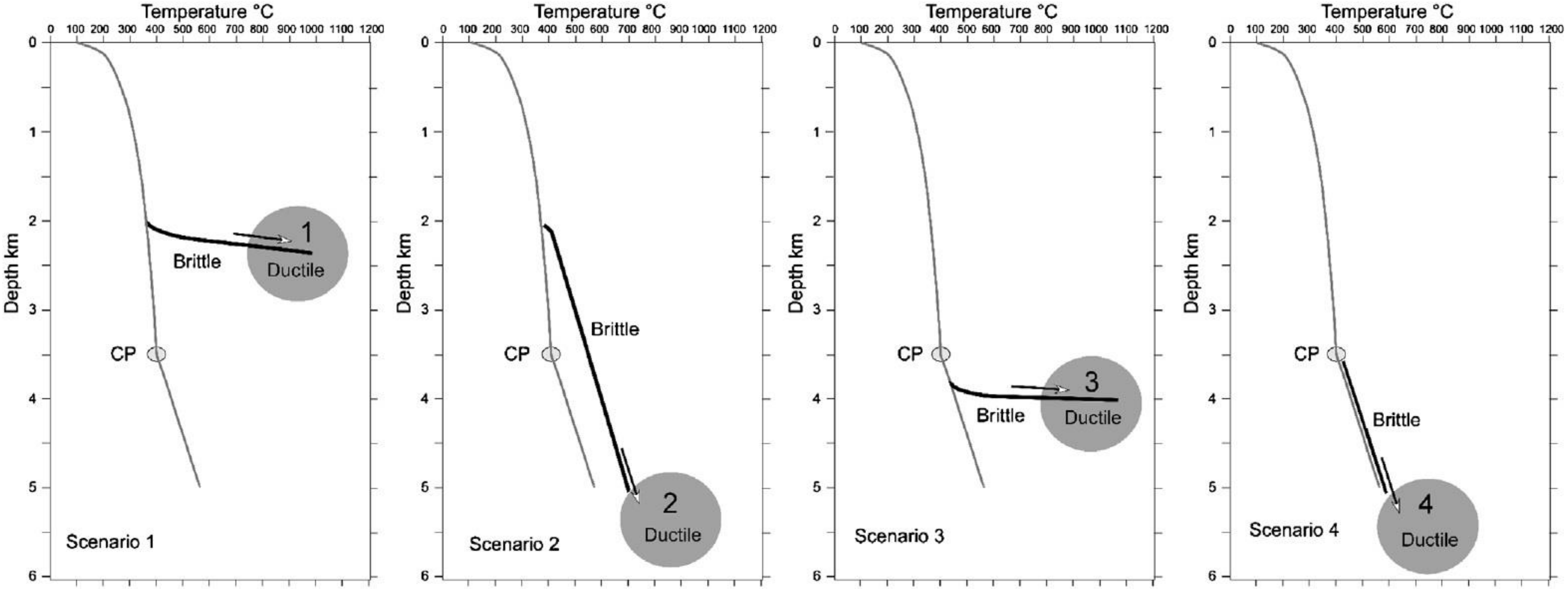
Pressure-enthalpy diagram (pure water, selected isotherms)



- 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

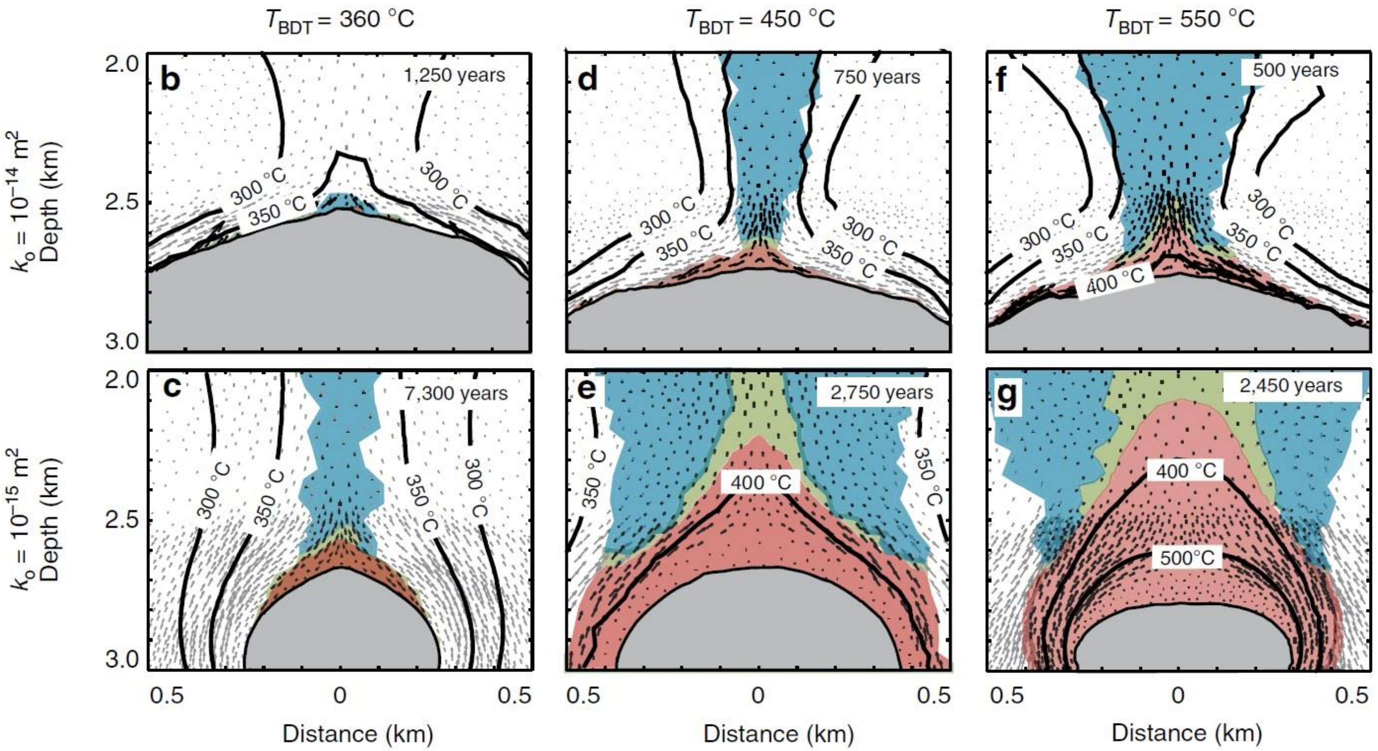
1. *T* path if we drill into magma at ~2 km depth, **2.** *T* path if we drill downwards along a contact aureole of a large magma intrusion, **3.** *T* path if we drill into magma at 4 km depth, **4.** *T* path if we drill beyond the critical point within the amphibolite facies rocks at supercritical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

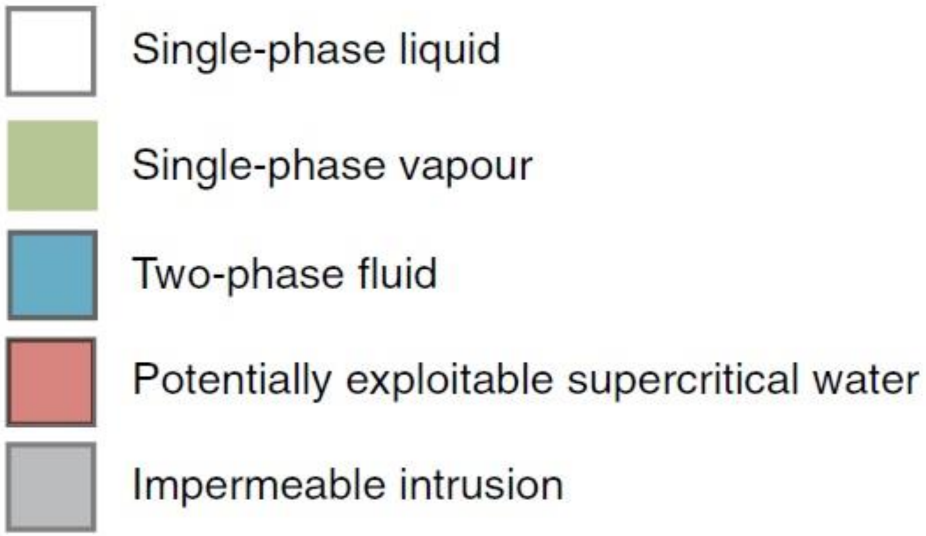
- Molten or recently crystallized, shallow intrusion responsible for superheated conditions are fairly widespread at Krafla.
- Hydrous phases may exist in the crust at depths where the average *T* exceeds 400°C. Expected *T* at all IDDP drill fields of Iceland, range from 550°C to 650 °C at 5 km.
- Geothermal 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 about 4–10 MW 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} m^2 ('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)

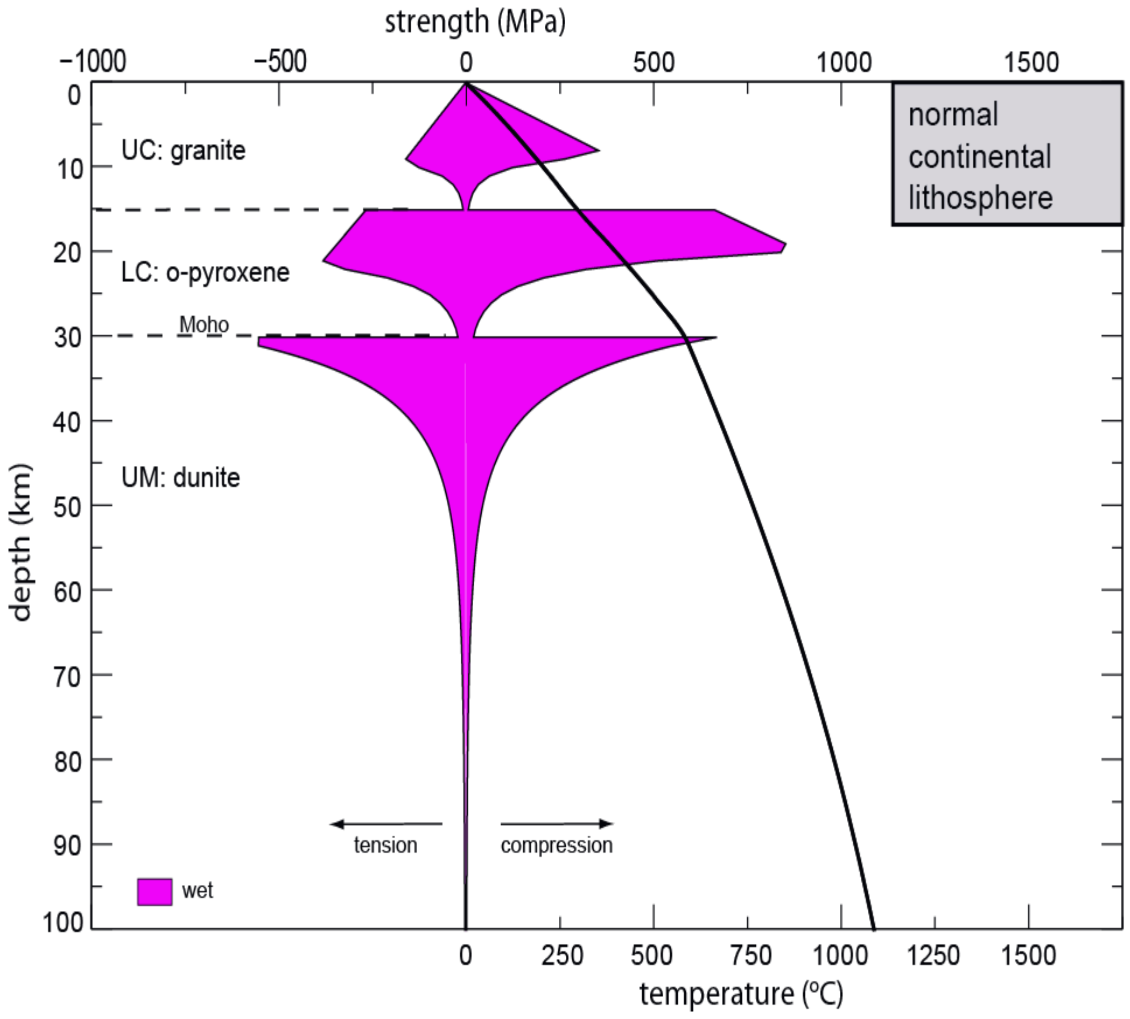
Scott et al., 2015, Nat. Communications

- 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)

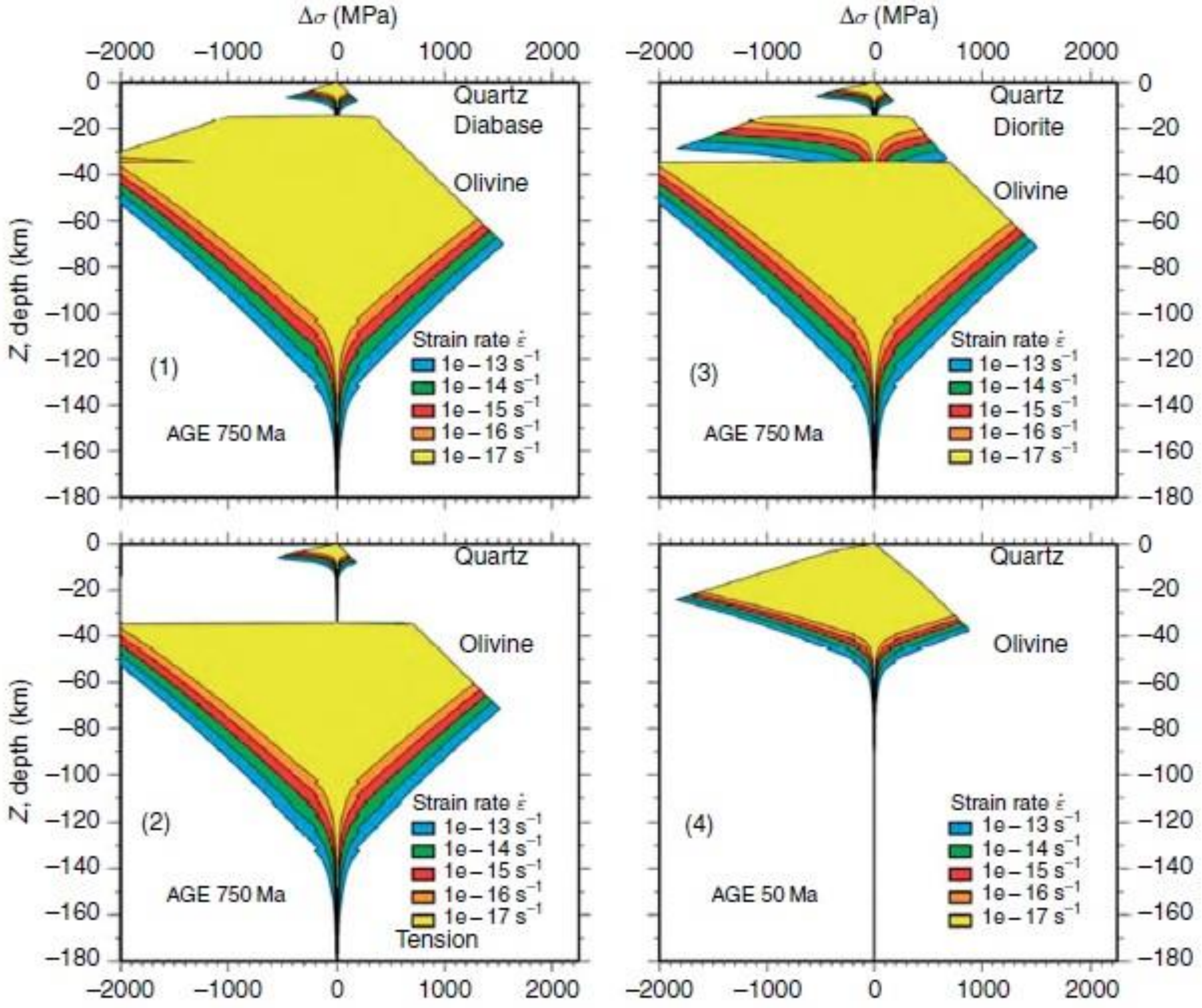
Continental lithosphere



- 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.

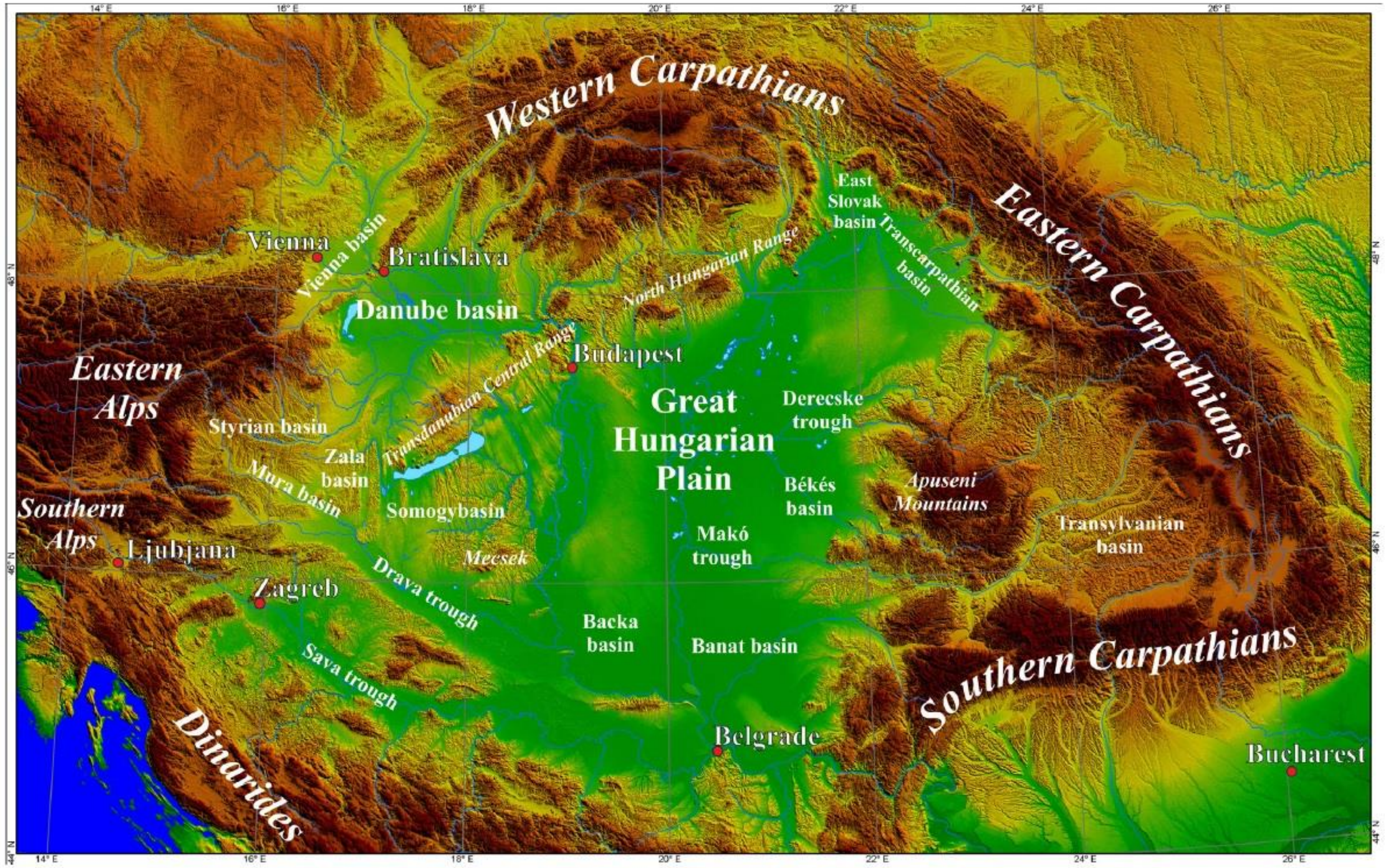
Brittle-Ductile Transition (BDT) and strain rates

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.

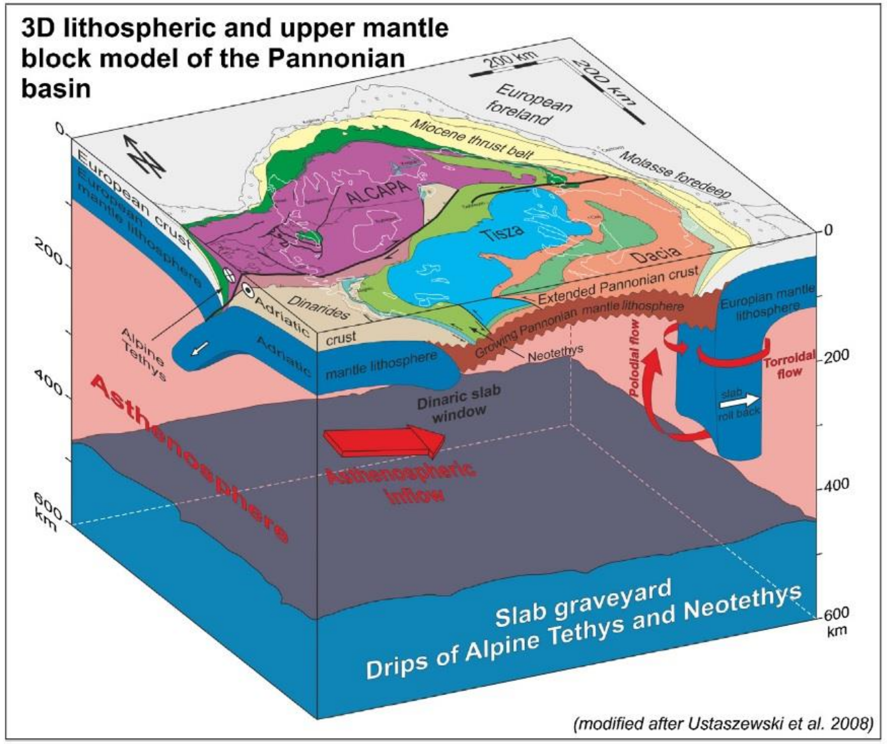
Geothermal Systems: Pannonian Basin

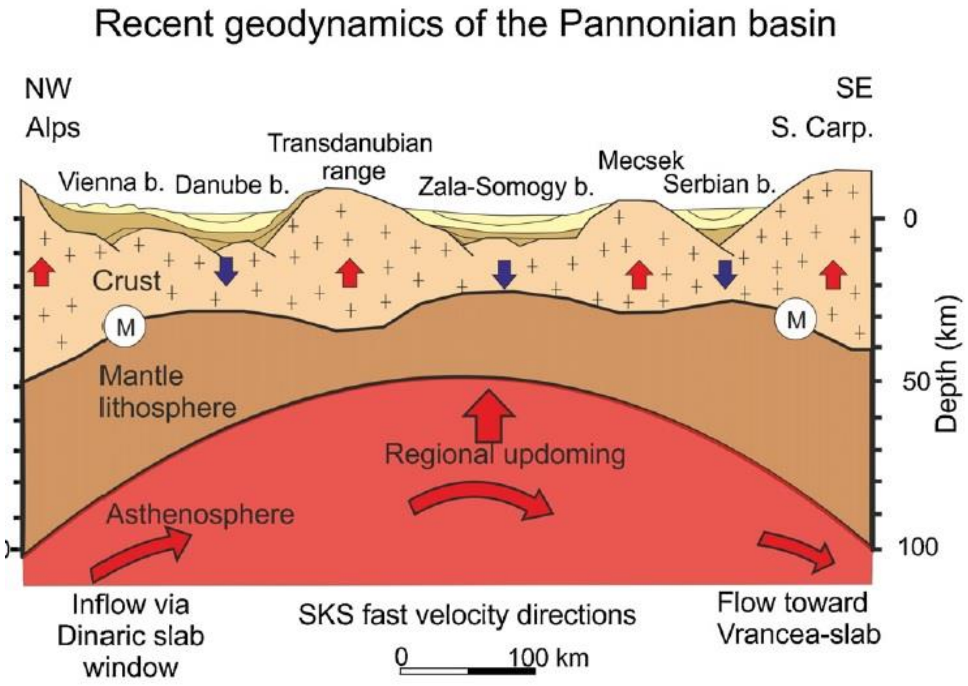


Horváth et al., 2015, Geothermics, 53

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



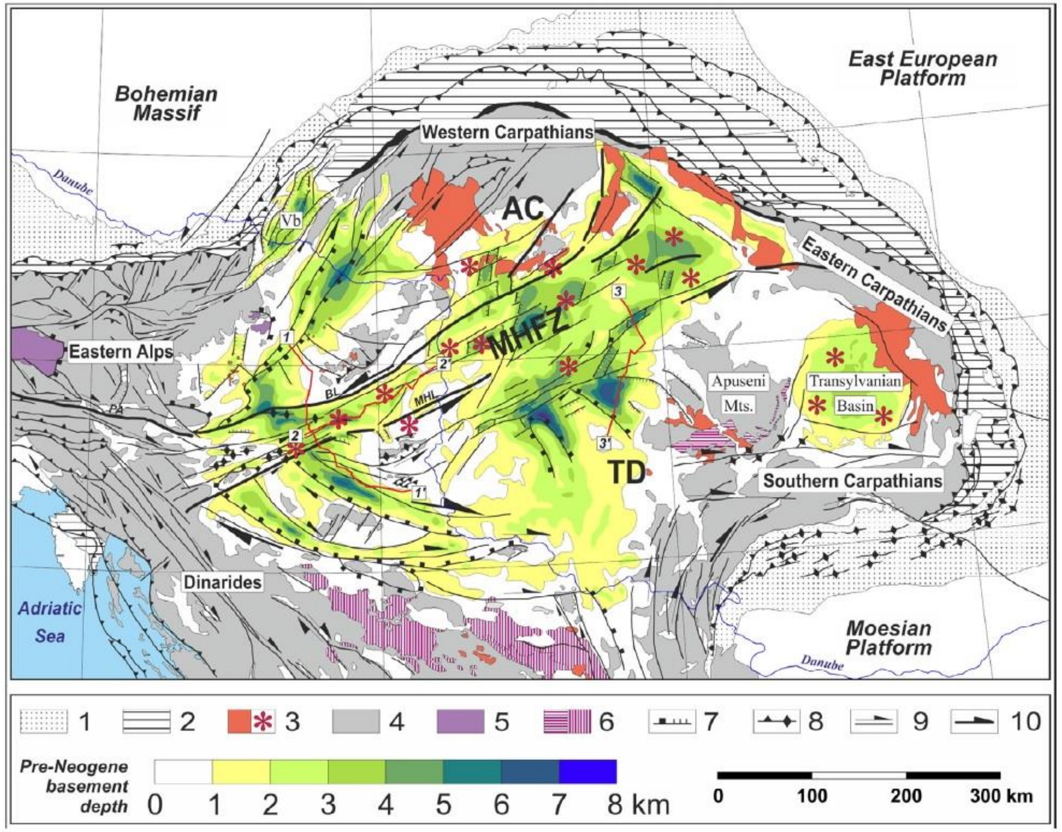


Horváth et al., 2015, Geothermics, 53

- 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.

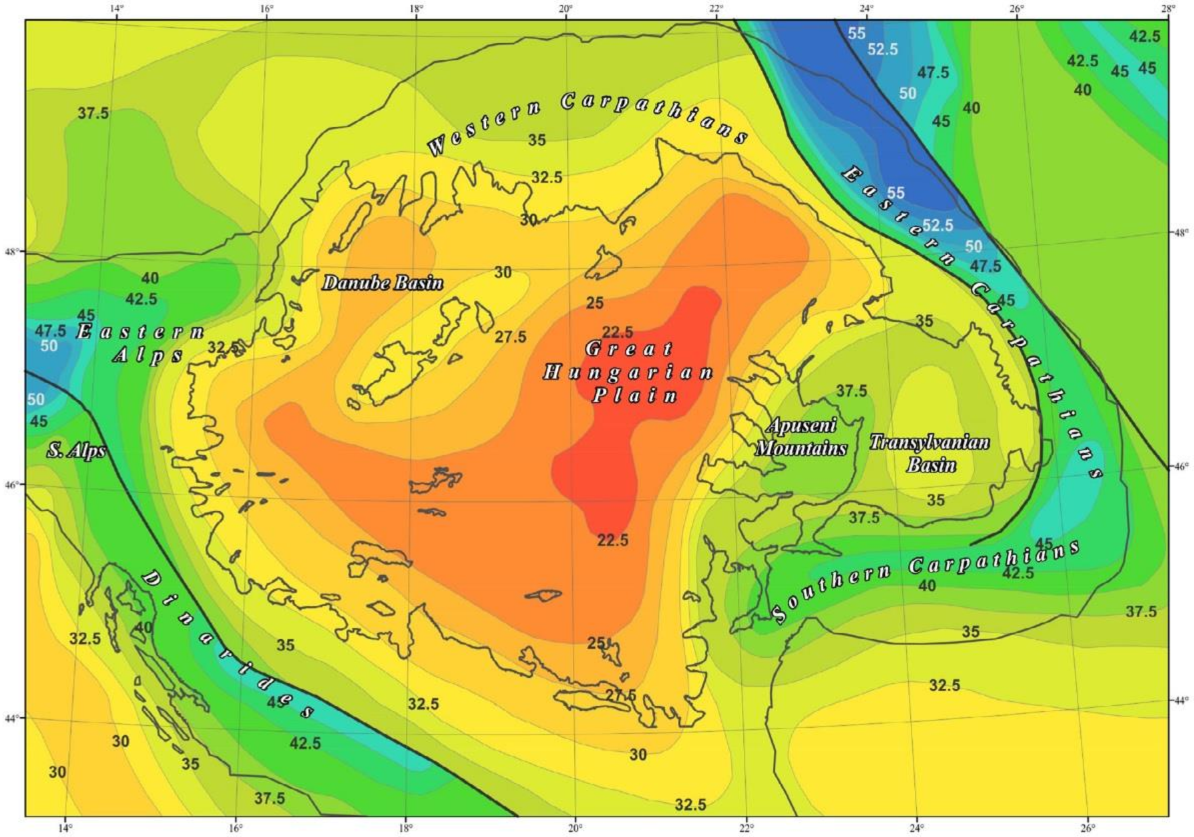
Pannonian Basin

Depth of the Basement



Horváth et al., 2015, Geothermics, 53

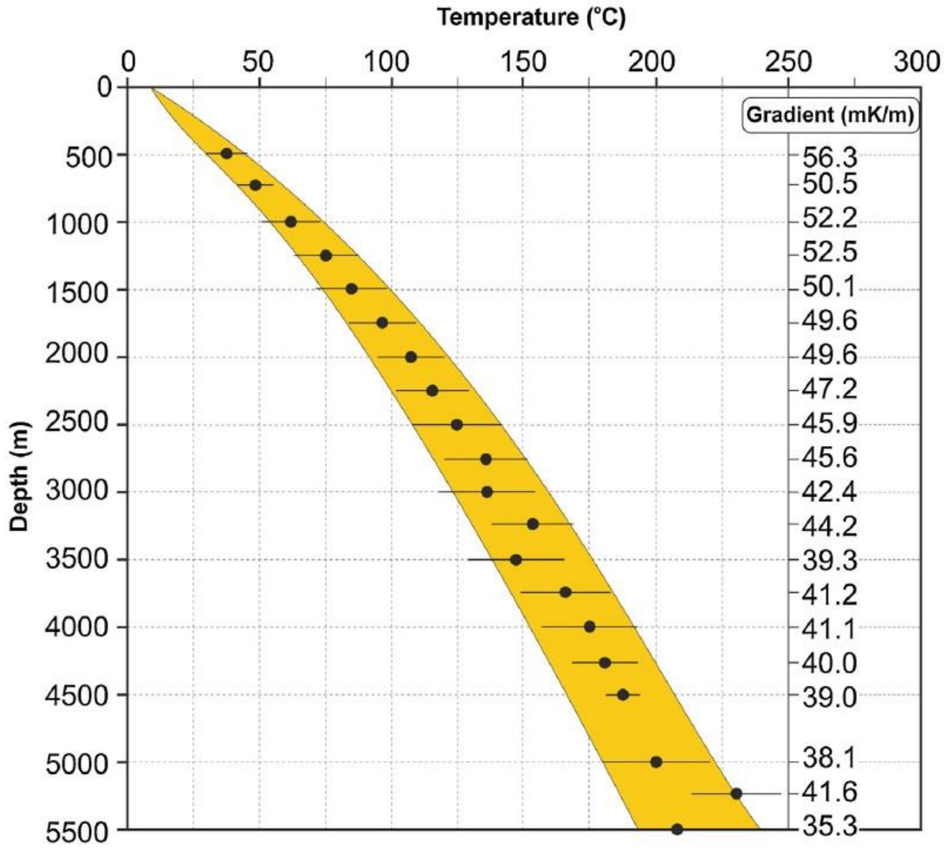
Crustal Thickness



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

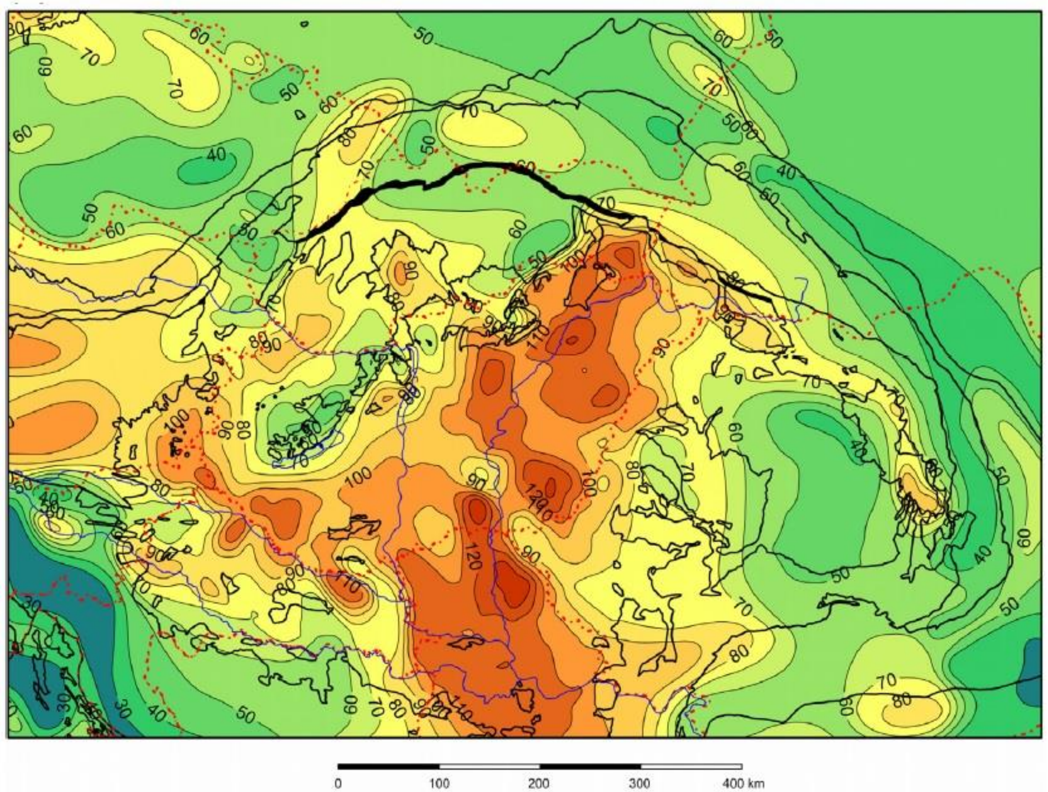
Geothermal Gradient



Horváth et al., 2015, Geothermics, 53

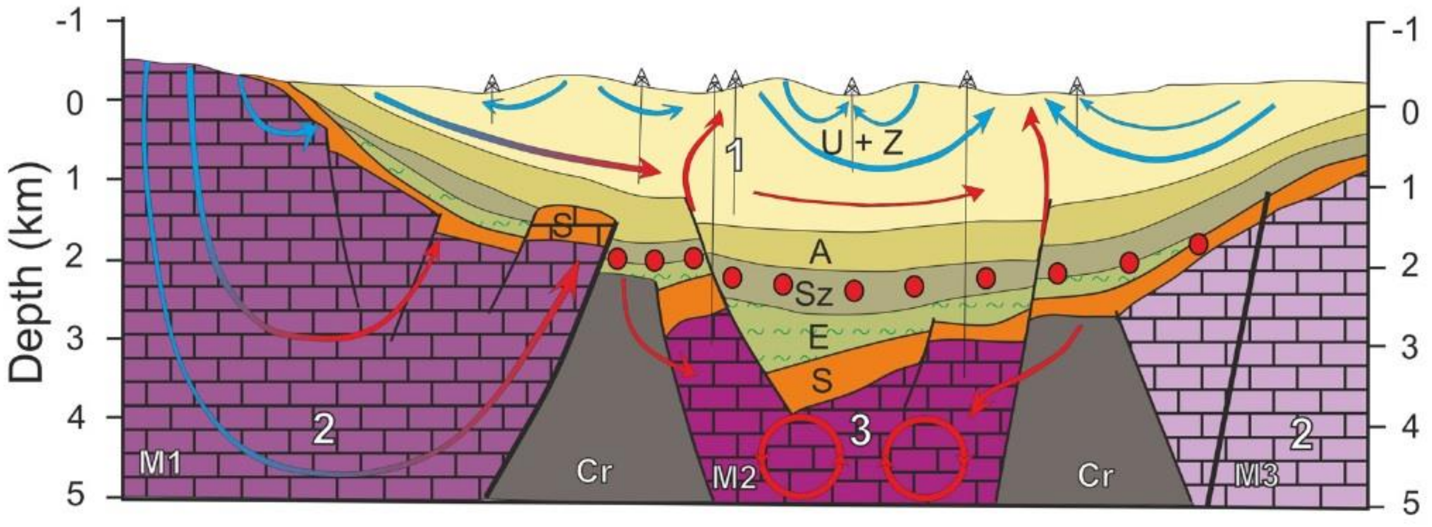
Surface Heat Flow

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



- 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

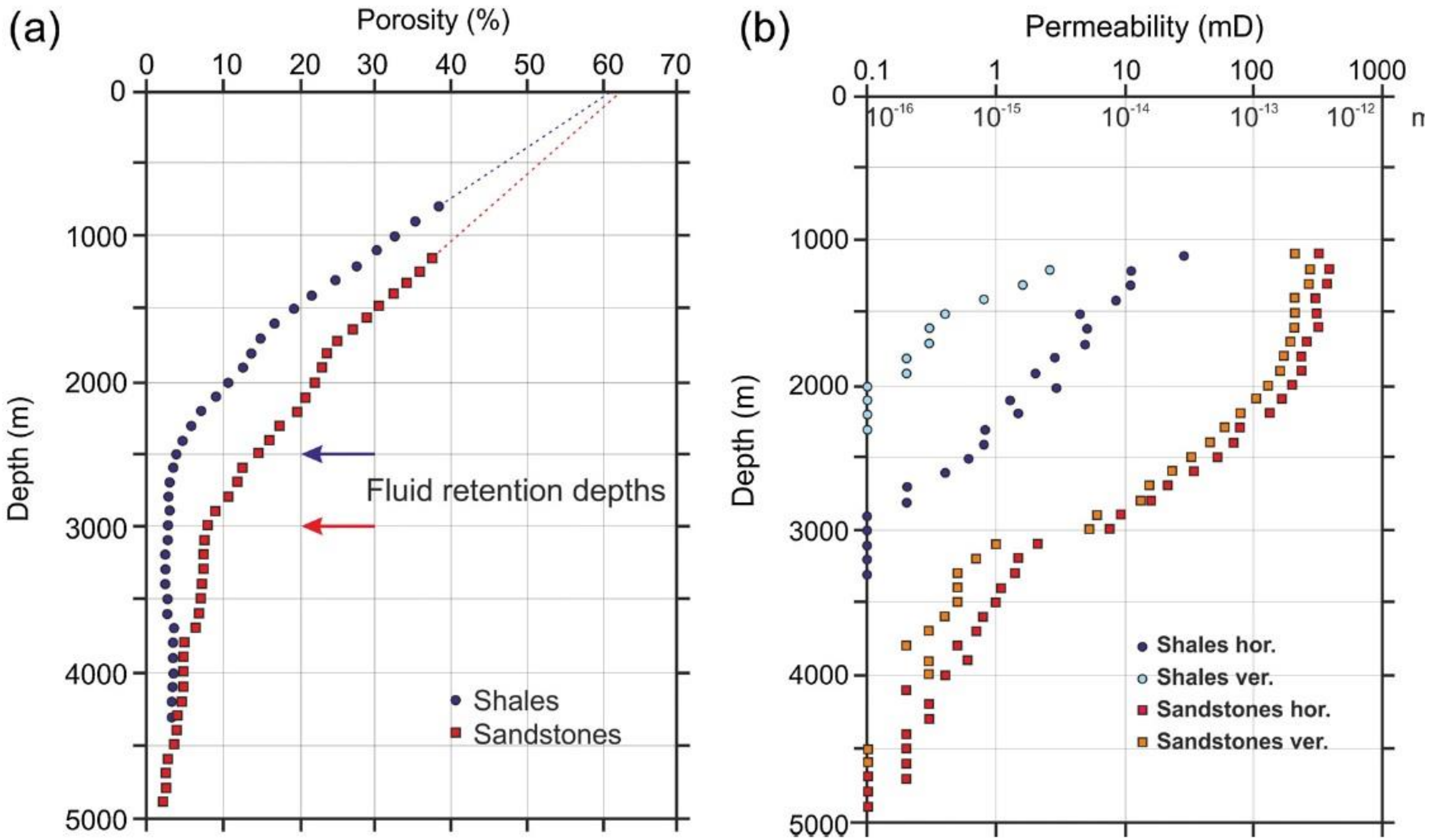


1 and 2 = gravity-driven flow system in the porous basin fill rocks, and Mesozoic (locally Eocene and/or Miocene) carbonates. 3: an overpressured system below a pressure seal (red ellipses) including lower Pannonian and Early to Middle Miocene basin fill, and fractured basement rocks. U + Z = Újfalu 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.

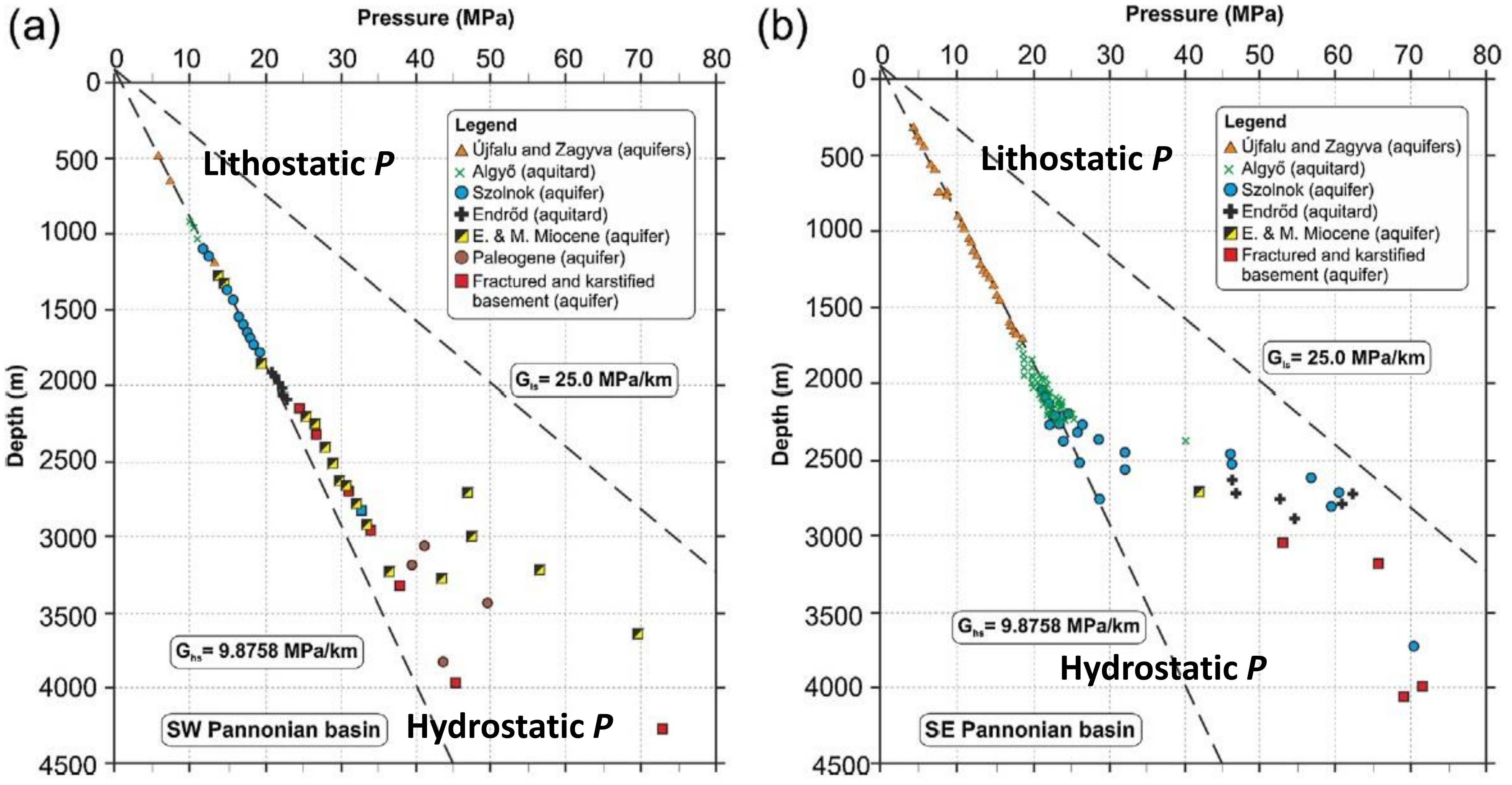
Pannonian Basin Hydrothermal Setting



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

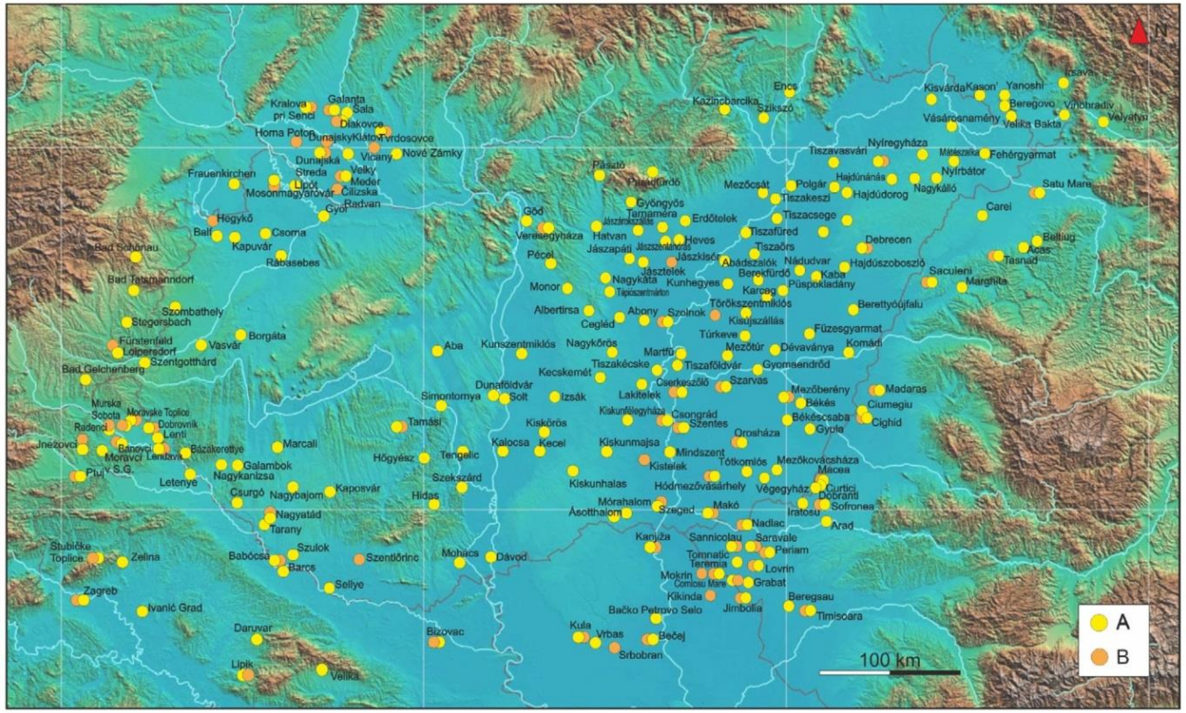


Horváth et al., 2015, Geothermics, 53

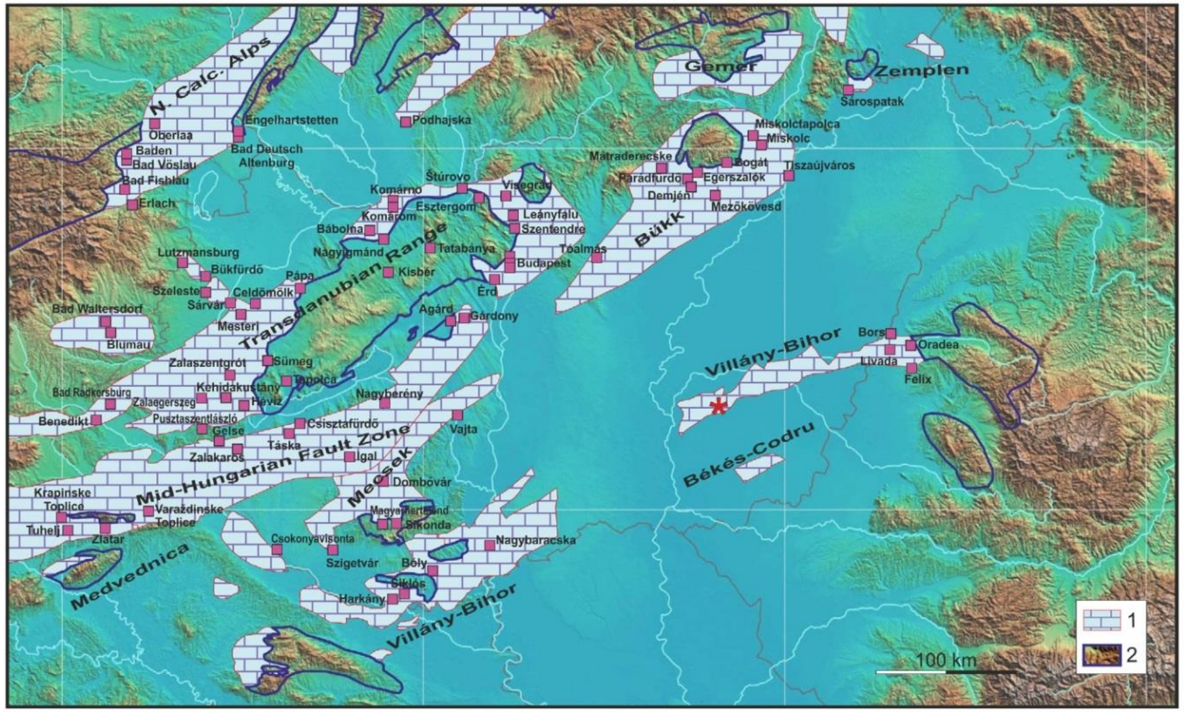
- In the southwestern Pannonian basin overpressure conditions appear with a pressure gradient slightly above normal between 2200 and 3000 m.
- In the southeastern Pannonian basin the increase of *P* seems to be very sharp in the 2200–2900 m depth range.

Geothermal installations in the Pannonian Basin

Main geothermal installations in the Pannonian Basin



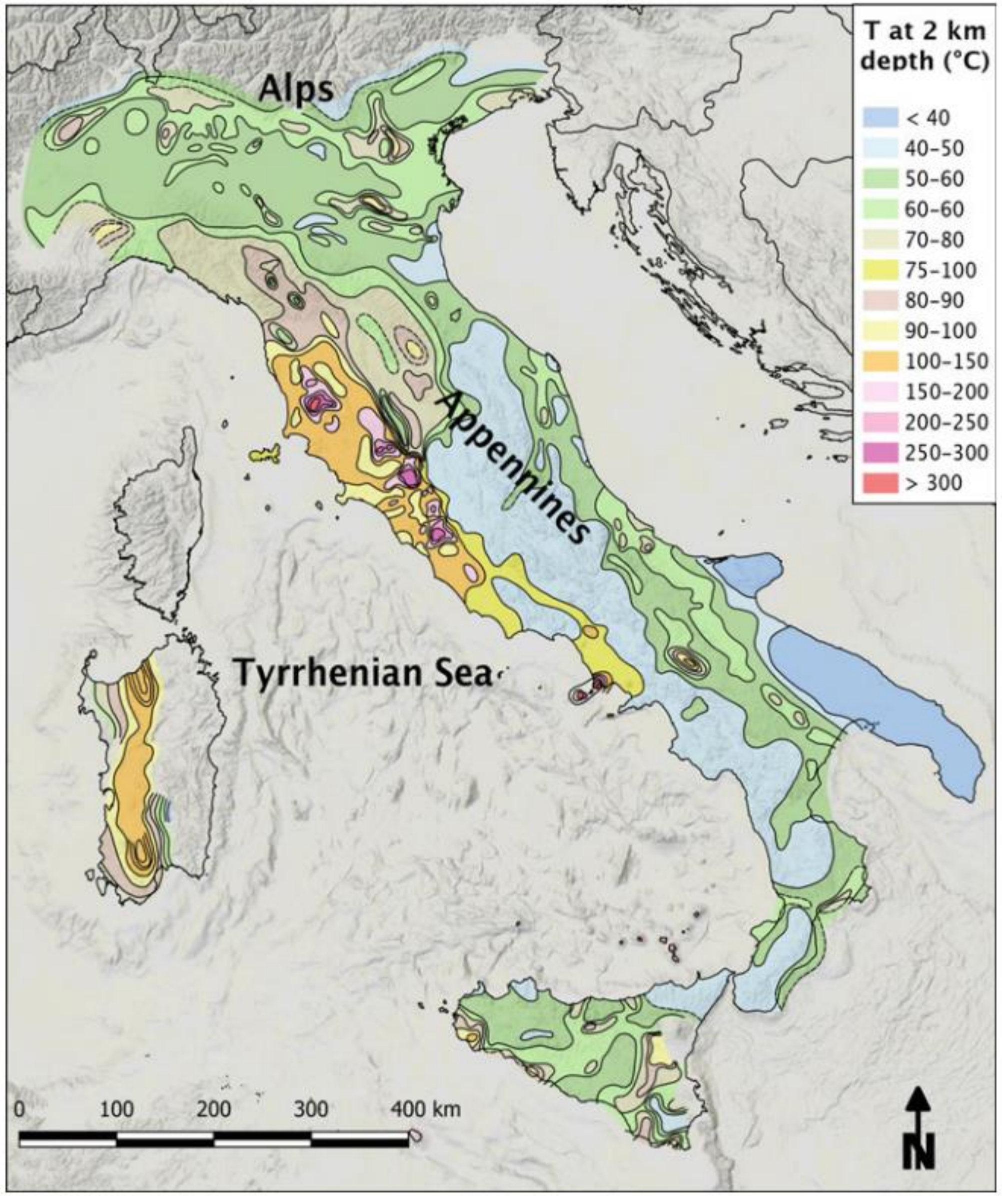
Main Mesozoic karstic reservoirs below the surface (symbol 1) in the Pannonian basin and their recharge areas (symbol 2)



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.

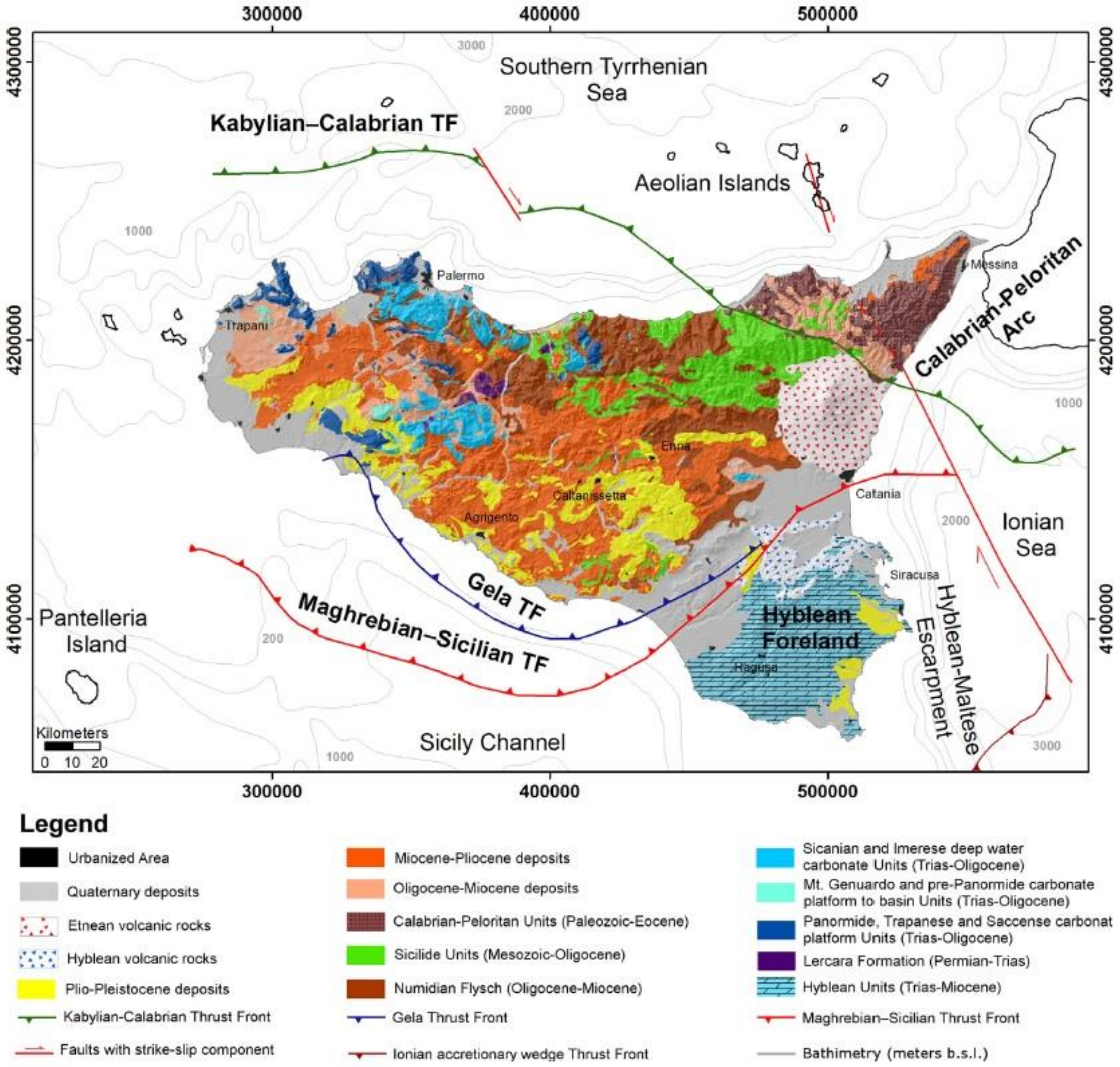
Italian Geothermal System



Temperature at a depth of 2 km

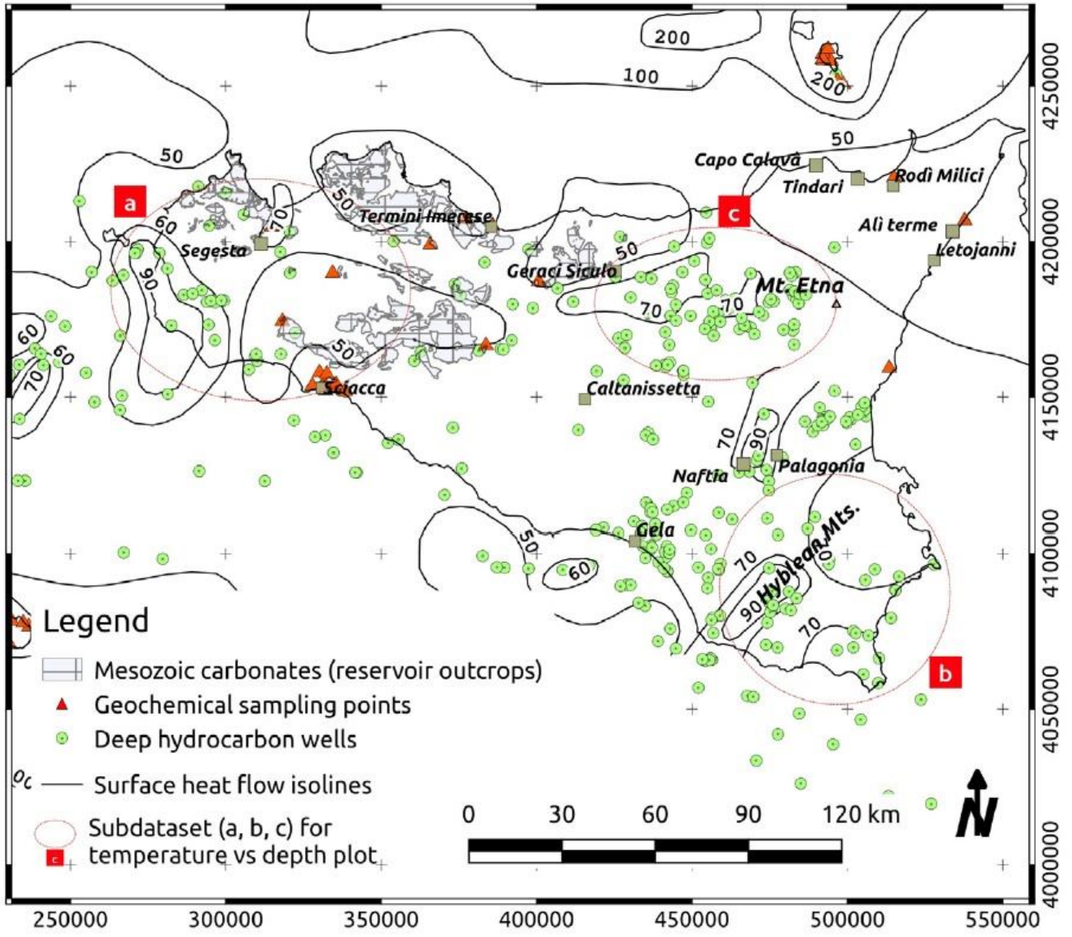
Cataledi et al., 1995, Geothermics 24

Geothermal Systems: Sicily

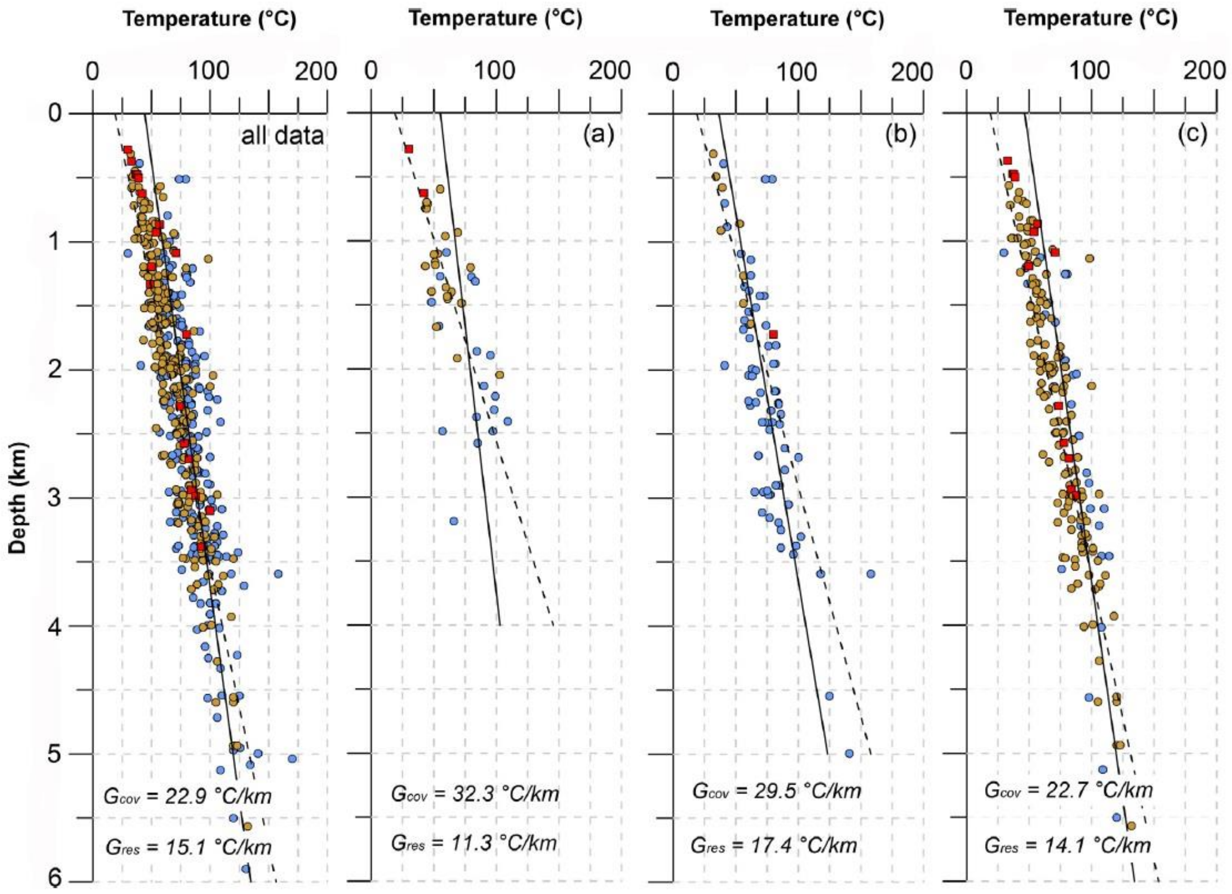


Trumpy et al., 2015, Geothermics, 56

Geothermal Systems Cases of Study: Sicily



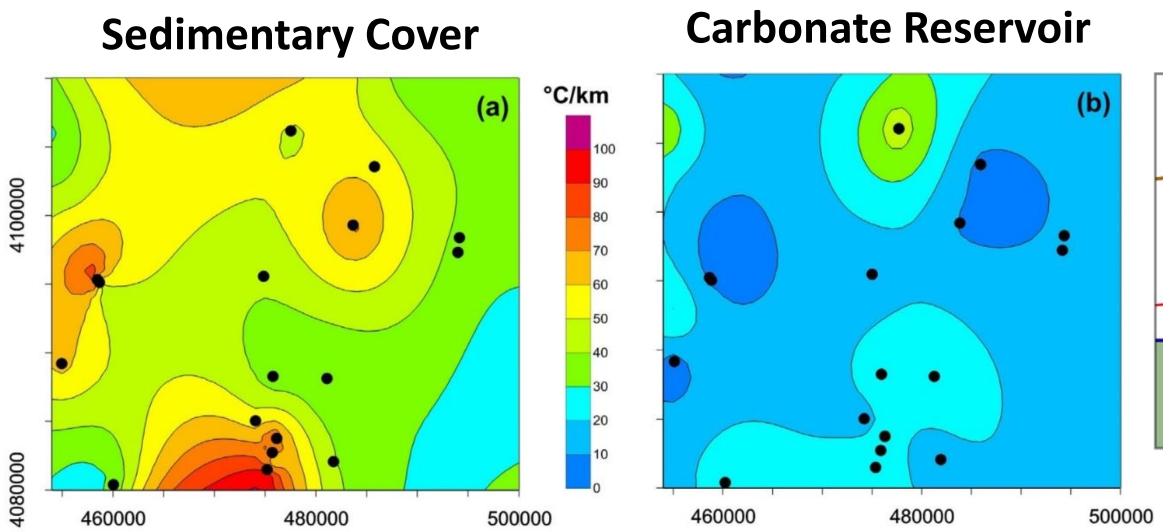
Trumpy et al., 2015, Geothermics, 56

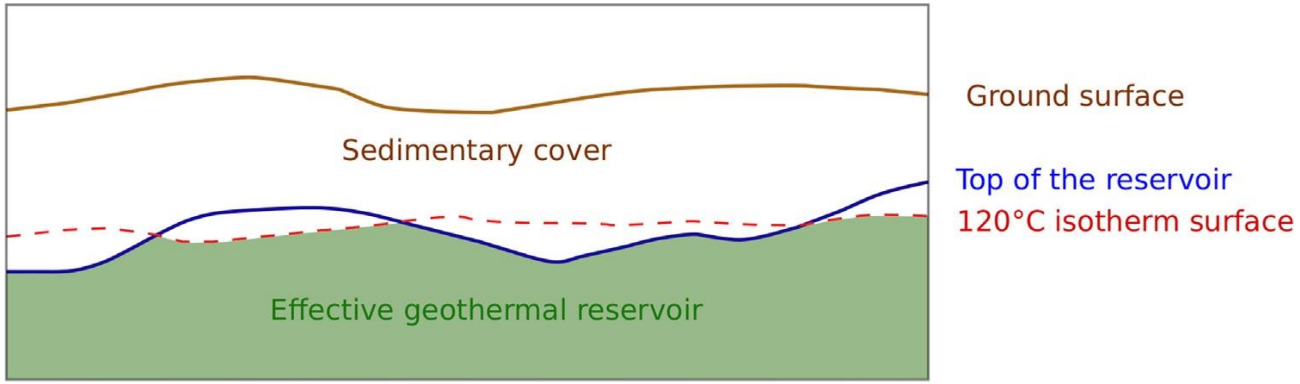


G_{COV}= Cover Gradient G_{Res}= Reservoir Gradient

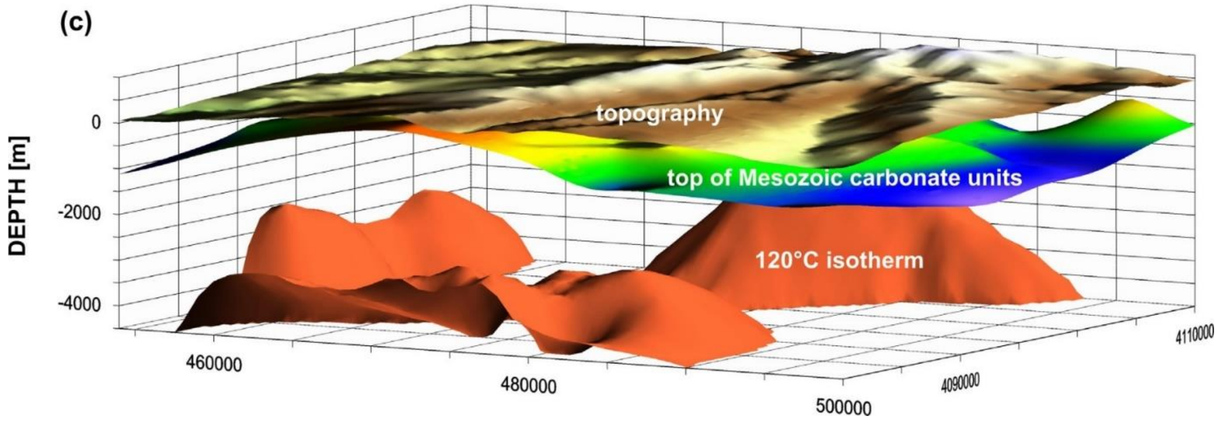
- In eastern and southwestern Sicily the heat flow values are anomalously high (up to 90 mW/m^2), while low values occurring across the outcrops of Mesozoic carbonate may be due to infiltration of meteoric waters.
- The resulting thermal gradients reflect the fluid movements, they are anomalously high in the cover formations and move toward an isothermal profile in the reservoir units where the convective processes take place.

Geothermal Systems: Sicily





Trumpy et al., 2015, Geothermics, 56



The "effective reservoir layer" is derived by the intersection between the depth of the reservoir top and the 120◦C isotherm depth.

- The geothermal gradient maps are used to compute Rg:

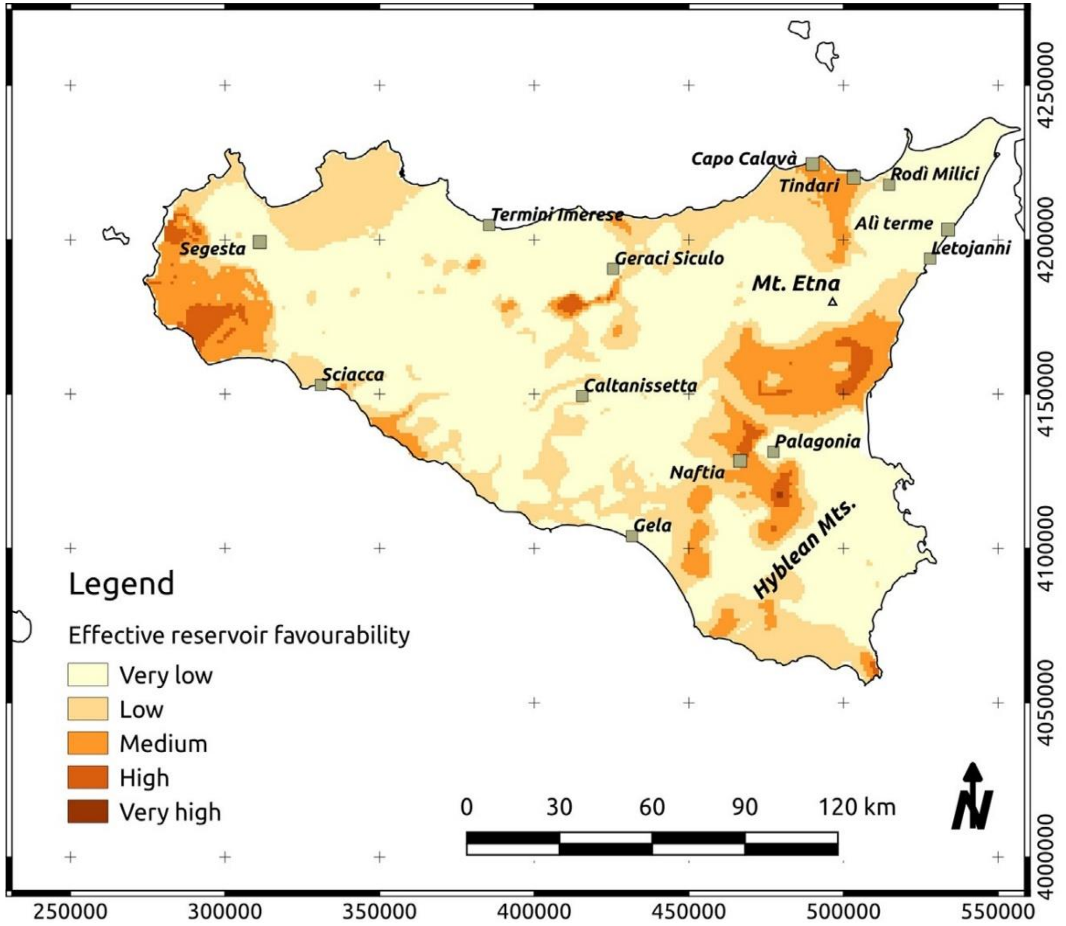
$$Rg = \frac{k_2}{k_1} = \frac{G_1}{G_2}$$

$K_{1(cover)}$=2.4 W/mK
$K_{2(Res)}$=3.6 W/mK

- Rg values, ranging from 1.2 to 1.8, can be explained by the thermal conductivity differences between the clay-rich terrigenous rocks and the overlying Mesozoic carbonates.
- Higher ratios can be interpreted as due to ascending/descending water movements occurring in the permeable reservoir, while lower ratios relate to ineffective cap-rock.

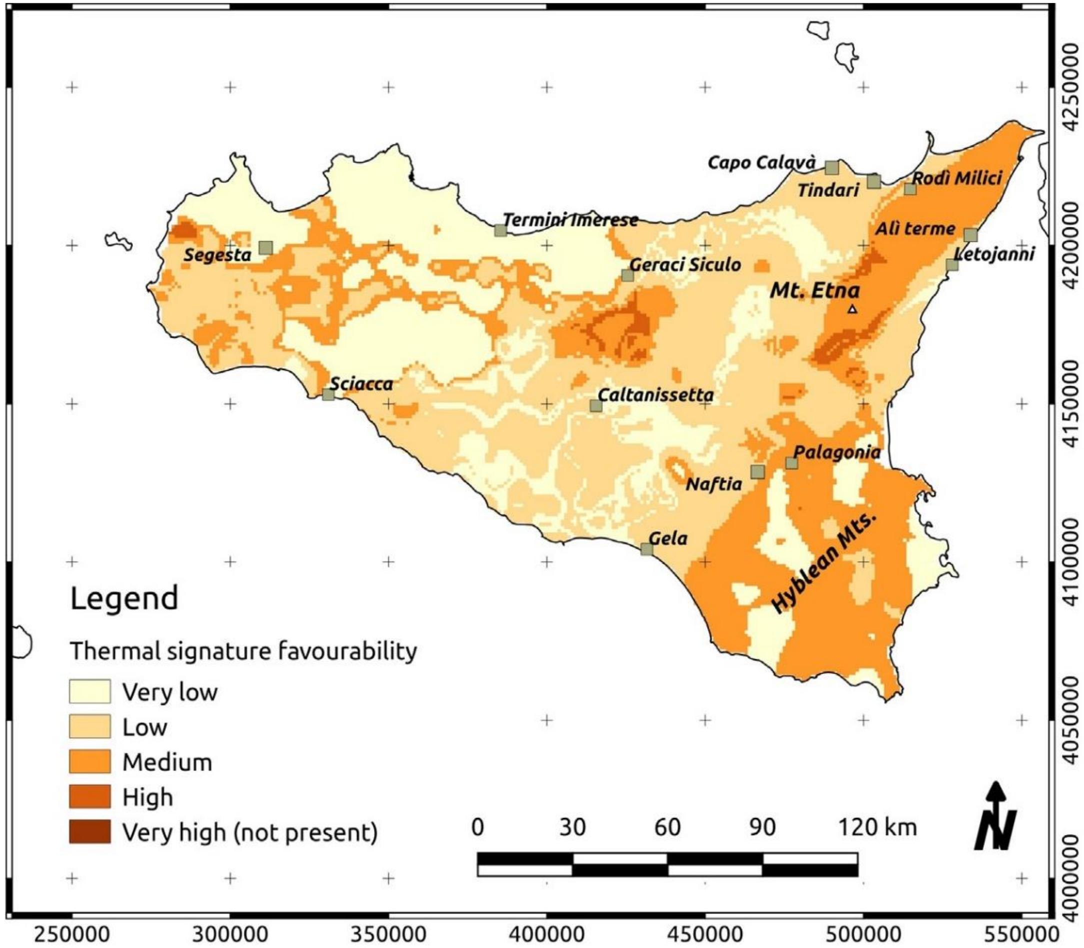
Geothermal Systems: Sicily

Effective reservoir layer



Trumpy et al., 2015, Geothermics, 56

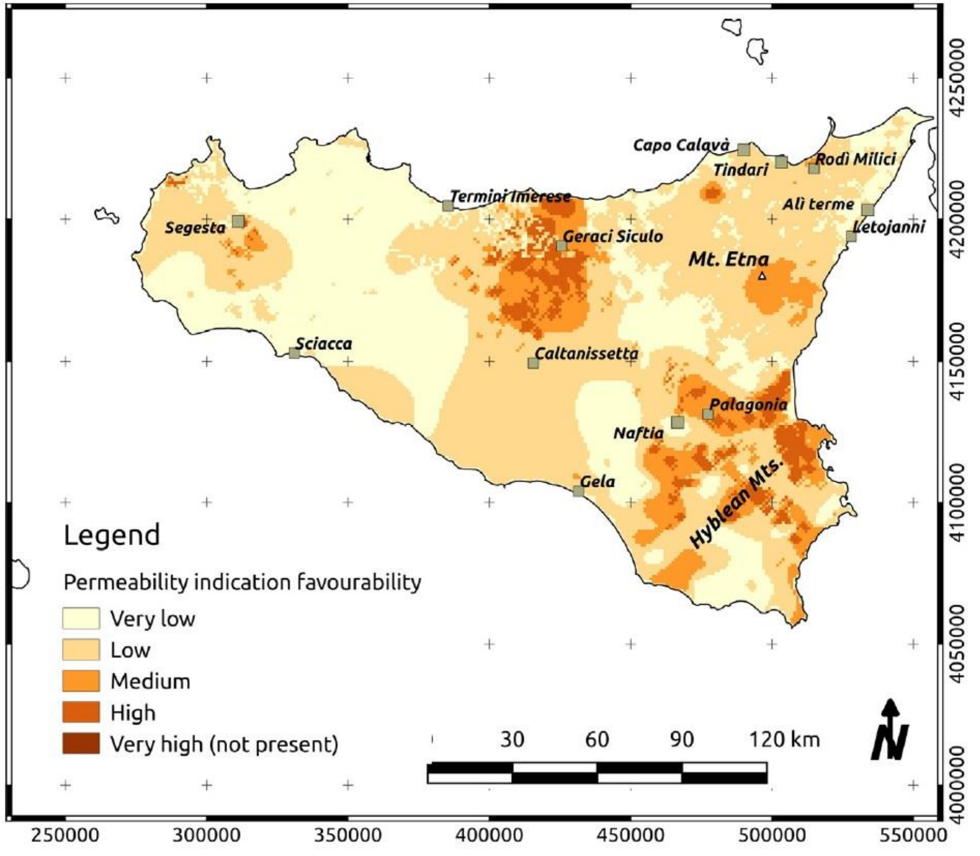
Thermal signature



- The effective reservoir favorability is estimated on the base of an exponential function describing the average-completed well costs as a function of depth.
- The thermal signature favorability is estimated by combining the temperature at the top of the potential reservoir and the depth of this top and taking into account the unfavourable effect of seeping meteoric water recharge where carbonates crop out or have a thin sedimentary cover (<500 m).

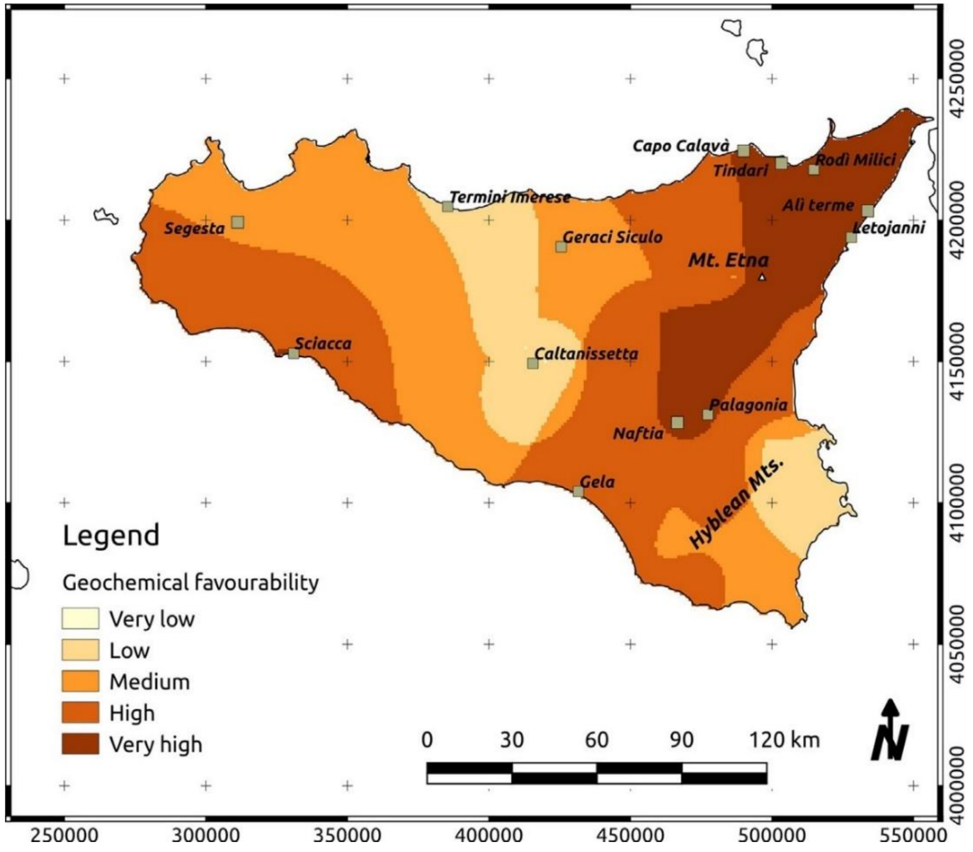
Geothermal Systems: Sicily

Permeability



Trumpy et al., 2015, Geothermics, 56

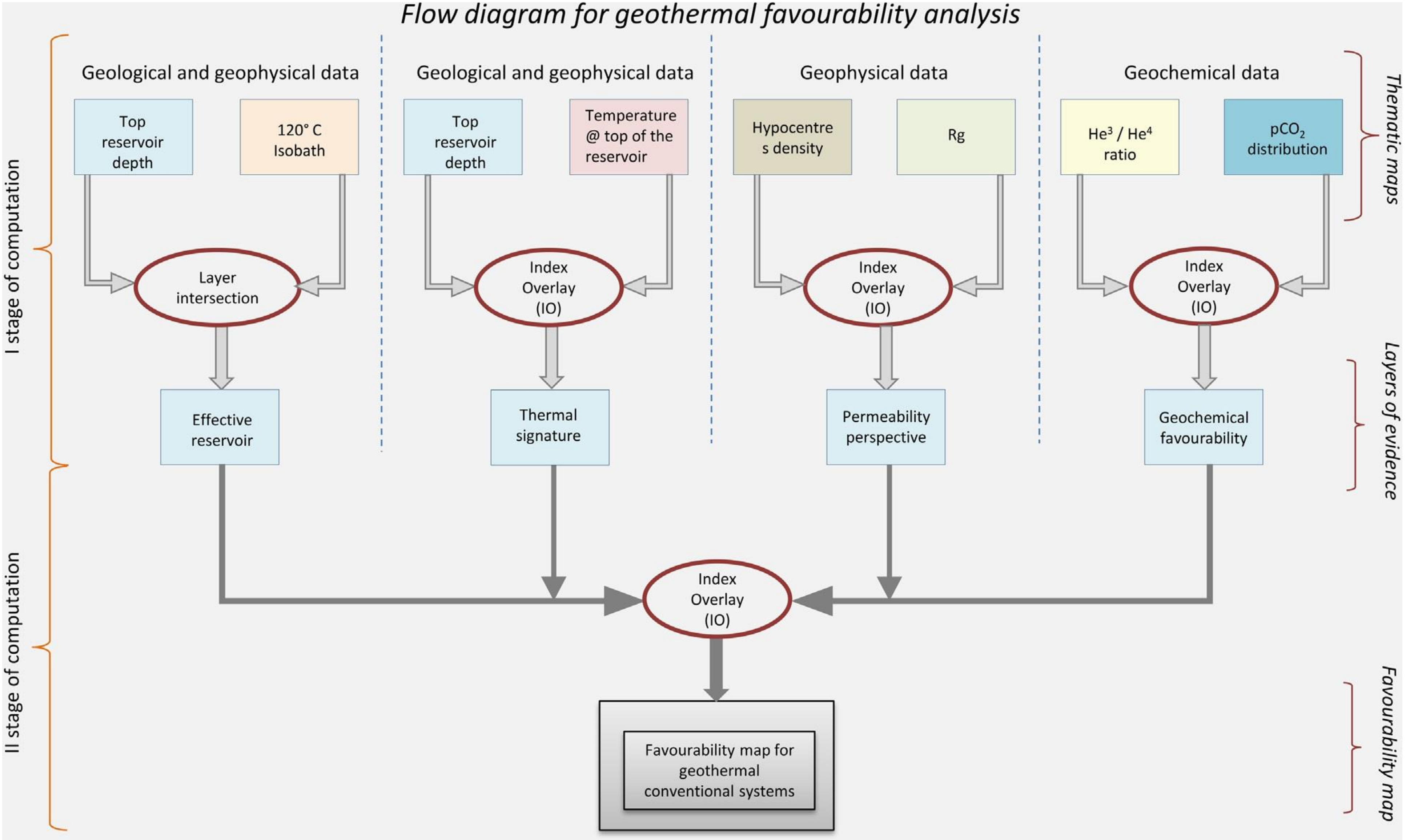
Geochemical favourability



- **The permeability indication favorability is estimated on the base of : (i) the earthquakes density analysis and (ii) the geothermal gradient ratio analysis.**
- **The geochemical favourability is estimated on theb ase of the $^3He/^4He$ ratios and the pCO_2:**

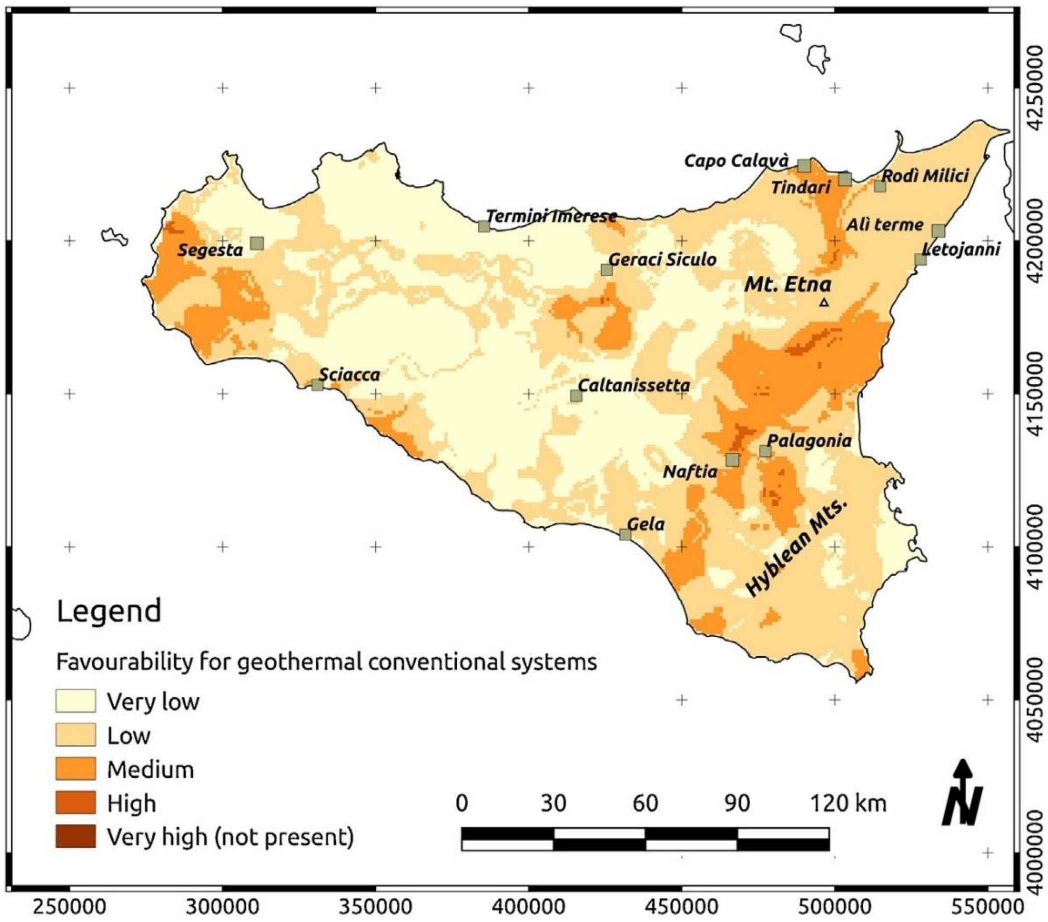
➢ Ratios of $^3He/^4He$ less than 0.05 are typically crustal, whereas ratios of more than 0.2 suggest an increase in primordial ^3He from the mantle. The maximum values of pCO_2 are located: (i) in eastern Sicily, (ii) around the Etna Volcano and the Hyblean Quaternary volcanics and (iii) at Sciacca, while the minimum values are centred in the Caltanissetta basin, the Hybean Plateau and the thermal springs in the N–NW sector of Sicily.

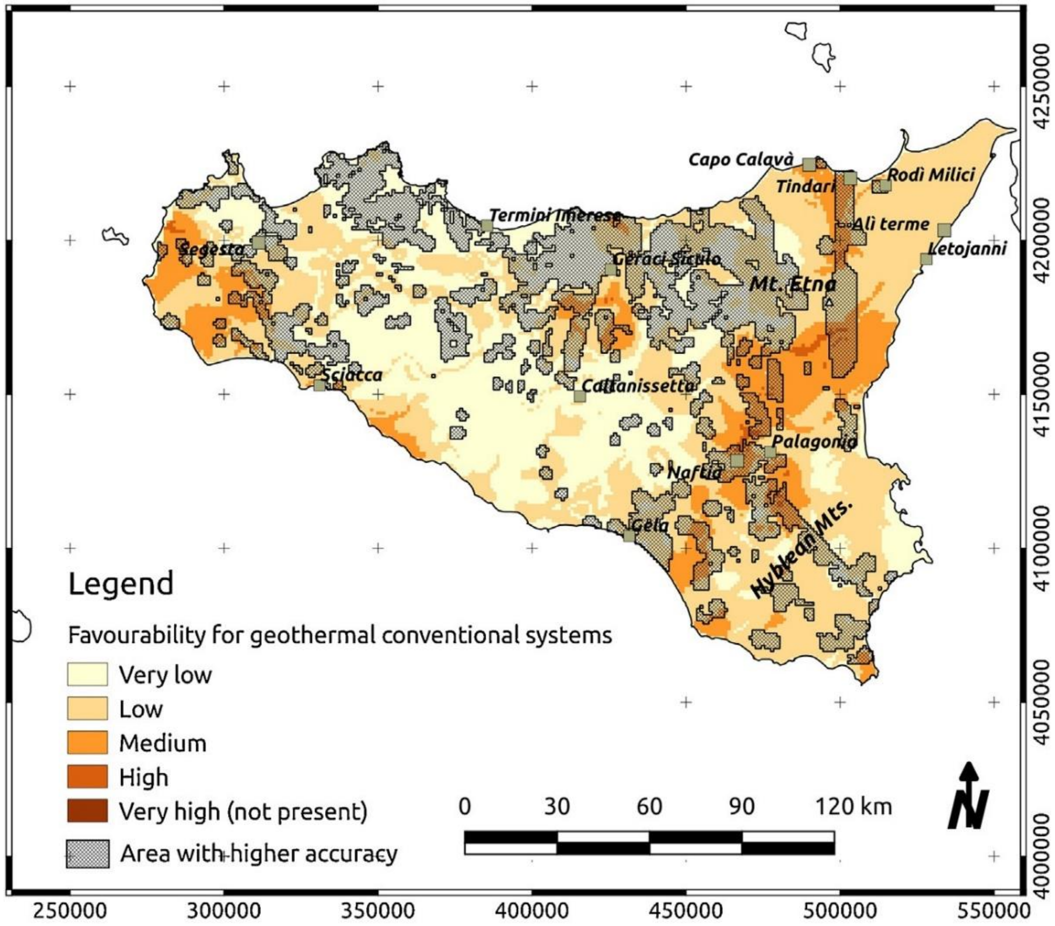
Geothermal Systems Cases of Study: Sicily



- Most geothermal wells exploit resources shallower than 3500 m, drilling depth higher than 4000–4500 m is considered critical in terms of cost for commercial purposes

Geothermal Systems: Sicily



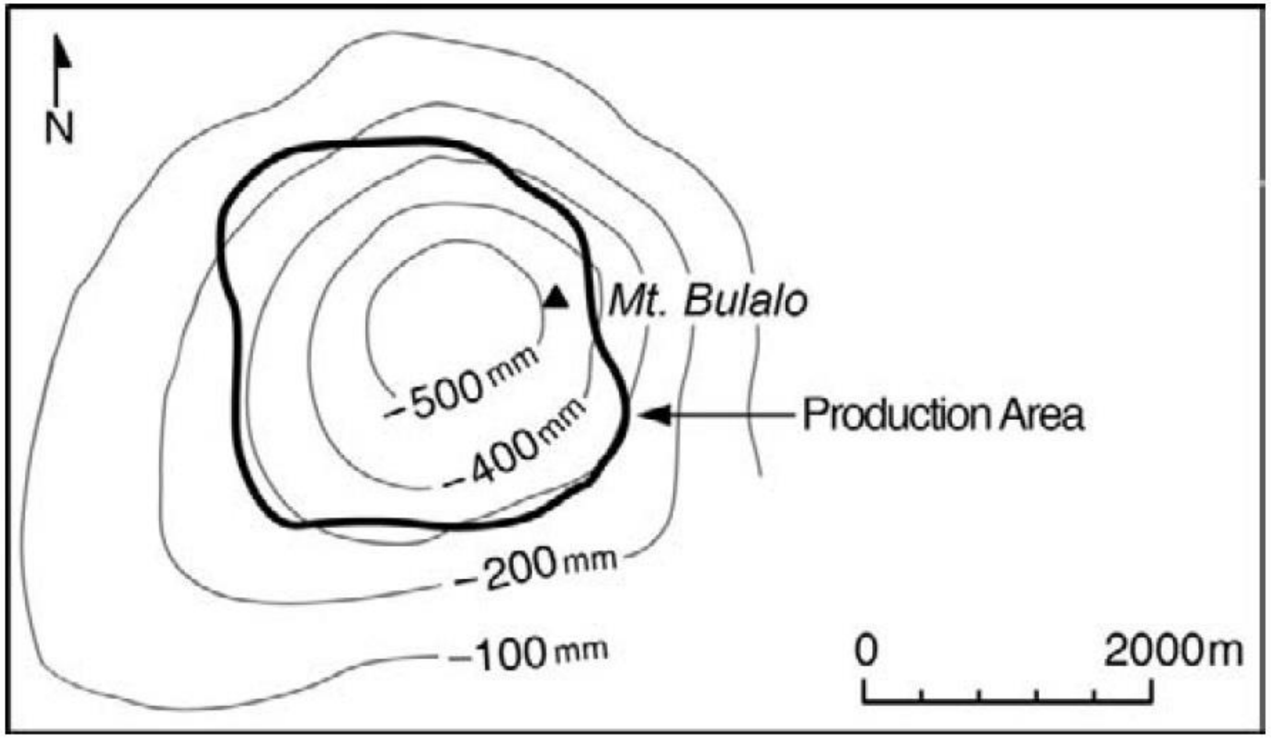


Trumpy et al., 2015, Geothermics, 56

- Low favourability values are located mainly where the regional reservoir geological units crop out and in the Caltanissetta basin where the reservoir is very deep and there is a low geothermal gradient in the sedimentary cover.
- Medium and highly favourable areas in eastern Sicily, from Mt. Etna towards the town of Gela, and in the westernmost area of Sicily.
- The accuracy map is obtained on the base of: (i) geological data, (ii) geochemical data, (iii) seismological data, and (iv) thermal data.

Consequences of Geothermal Exploitation: Subsidence

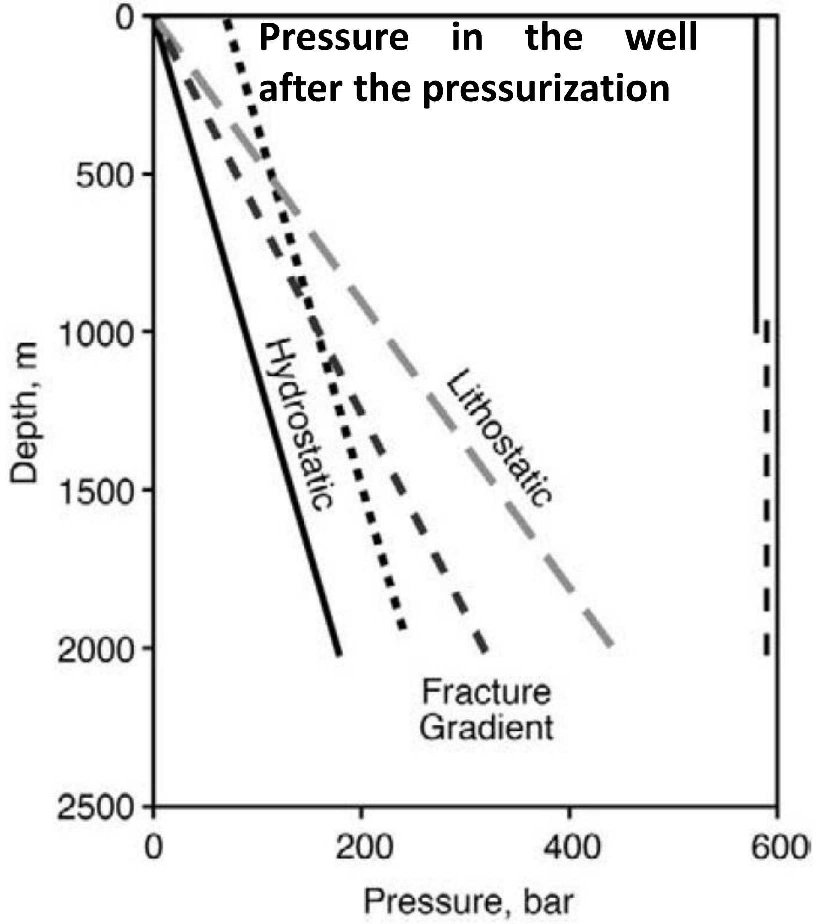
- The underground changes in *P* and *T* cause contraction (or expansion) of the rock, and this in turn causes a change in surface elevation.
- If the elastic properties of the rock are similar throughout and above the reservoir, the contraction is roughly proportional to the pressure (or temperature) change.
- Where injection causes a (local) rise in *P*, there may be inflation around the area of pressure increase.



Elevation changes 1979-1999 at Mak-Ban

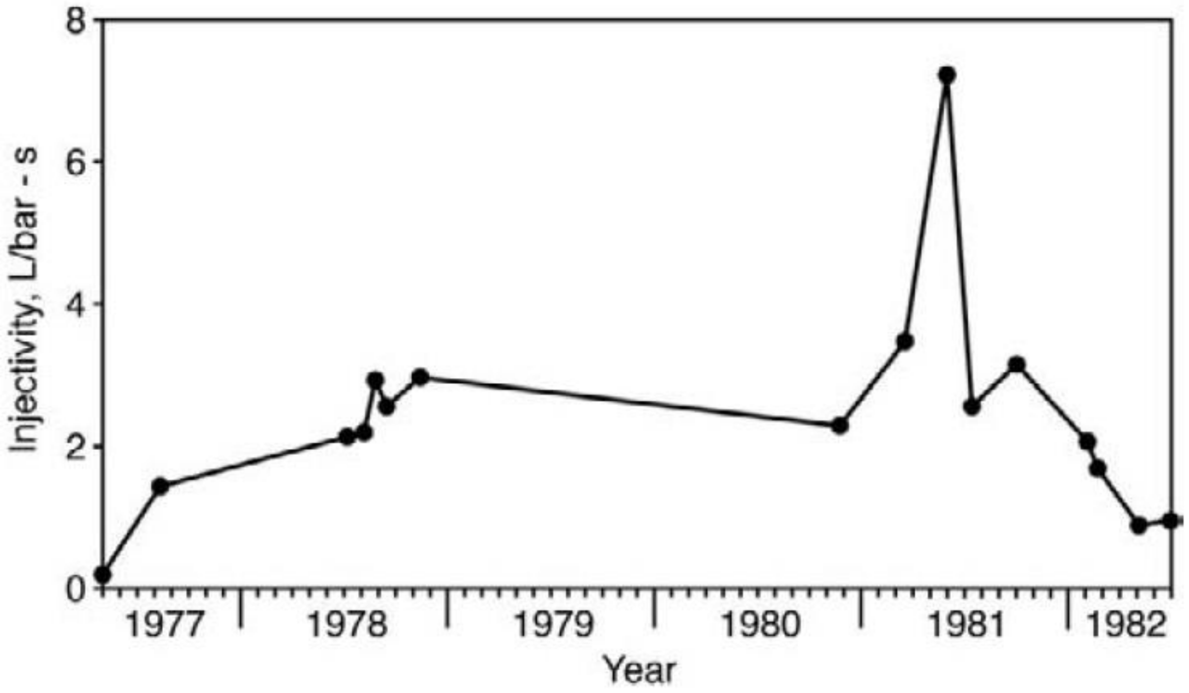
Well stimulation to enhance a geothermal reservoir

- Hydrofracking (pumping water or more viscous fluid) is a standard technique in petroleum reservoirs that is used to stimulate a well, to improve the flow of a well drilled into a formation, where the existing permeability is too poor to support sufficient flow.
- Under pressurization a fracture will form when downhole pressure exceeds the fracture gradient. It will tend to form near the top of the open interval, and it will be oriented normally to the direction of the least principal stress, which is usually in horizontal direction.
- If more pressure is applied, the fracture will grow upward, since there is more pressure excess, relative to the fracture gradient, at the top of the fracture than at the bottom.



Well stimulation to enhance a geothermal reservoir

- The simplest form of stimulation in high-temperature formations is achieved by pumping cold water into a well, usually at pressures too low to cause hydraulic fracturing.
- The mechanism is a result of thermal contraction of the rock, causing fractures to open. The effect is reversible, at least in part, since injectivity decreases if the well is allowed to warm up and then increases again under new injection.
- Stimulation of a geothermal well by acid occurs when acid is injected into the formation and dissolves some rock or material filling the fractures.



Injectivity history of well 4R1, Tongonan.

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