

Geothermics

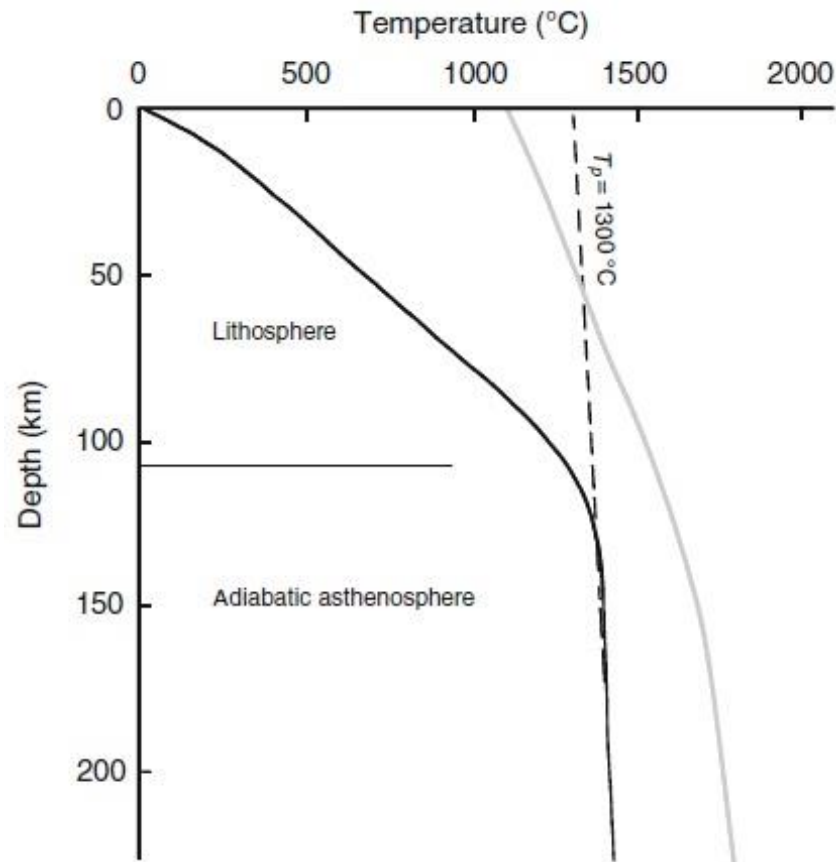
Course Outline:

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

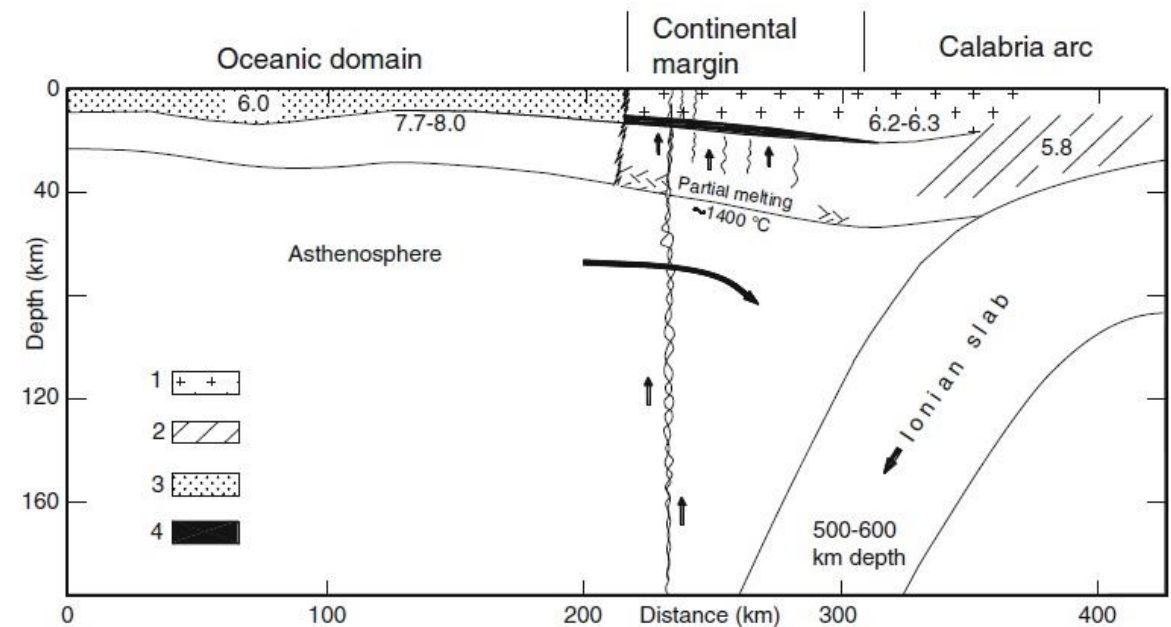
Temperature and magmatic processes

Melting generation is a consequence of:

- Decompression occurring during upper mantle rising (at midoceanic ridges).
- Water enrichment of the upper mantle due to dehydration of the subducting plate.
- Chemical elements (trace elements) enrichment/depletion of the upper mantle during metasomatic processes.



Garnet Peridotite Solidus



1–continental crust, 2–low velocity layers, 3–oceanic crust, 4–crustal underplating

Temperature, melting, and density

T of main volcanic rocks during eruptionss are: 700–900°C for rhyolite, 800–1100°C for dacite, 950–1200°C for andesite and 1000–1200°C for basalt.

According to lab experiments, for an anhydrous garnet peridotite (in absence of a fluid phase H_2O and CO_2) T_S and T_L depend on P (increasing with the increase of the depth, z):

$$T_S(z) = 3.0 z + 1100 \quad \text{solidus}$$

$$T_S(z) < T(z) < T_L(z)$$

$$T_L(z) = 1.5 z + 1850 \quad \text{liquidus}$$

$$x_f = [T(z) - T_S(z)] / [(T_L(z) - T_S(z))]$$

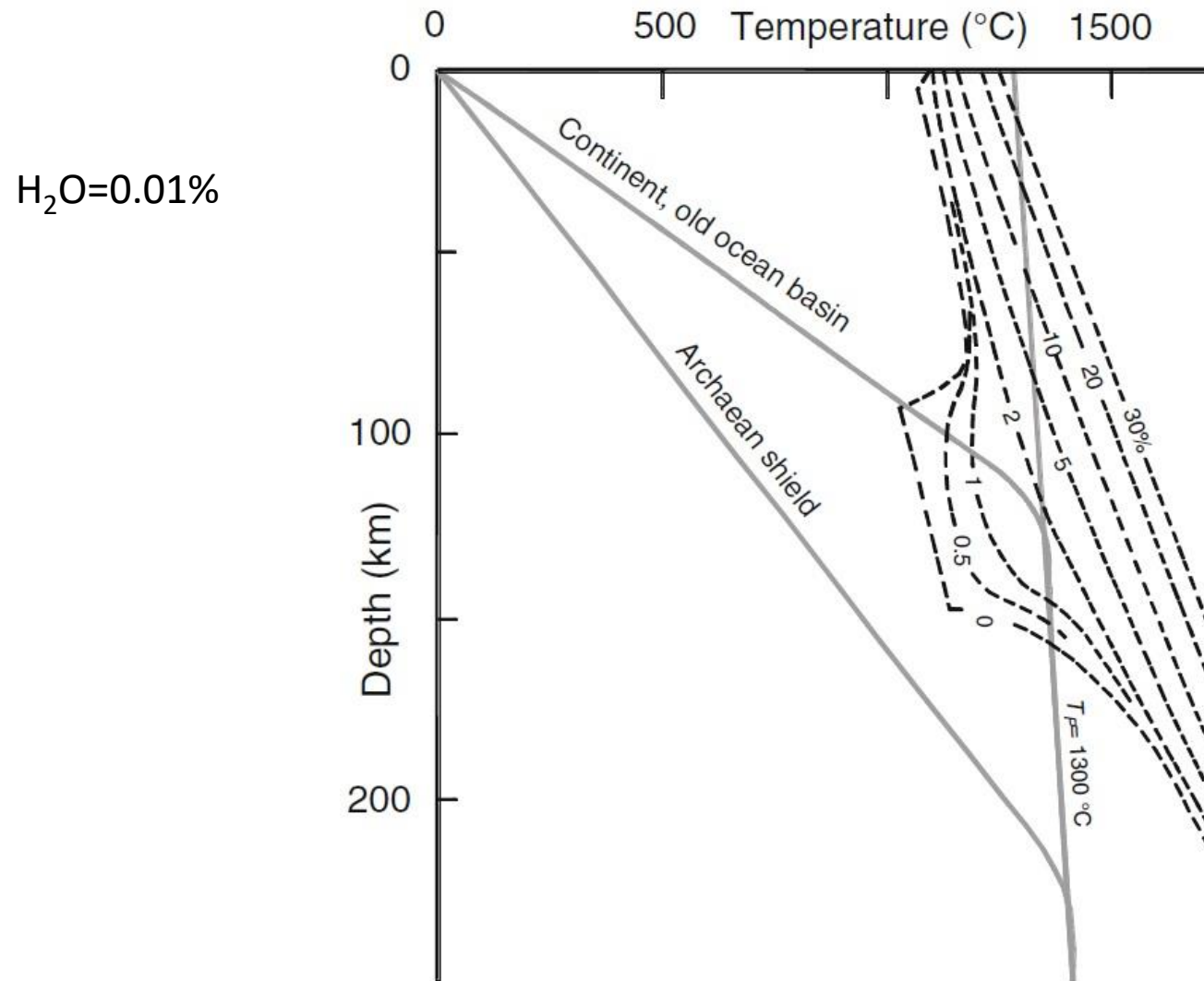
x_f = percentage of melt at a given depth z (in km) and temperature $T(z)$ in °C

Density of a magma varies from 2.2 g/cm³ (rhyolitic composition) to 2.8 g/cm³ (basaltic composition)

$$\rho = \sum x_i M_i / \sum x_i V_i$$

x_i , M_i , and V_i = molar fraction, molecular weight, and molar volume of the various chemical species.

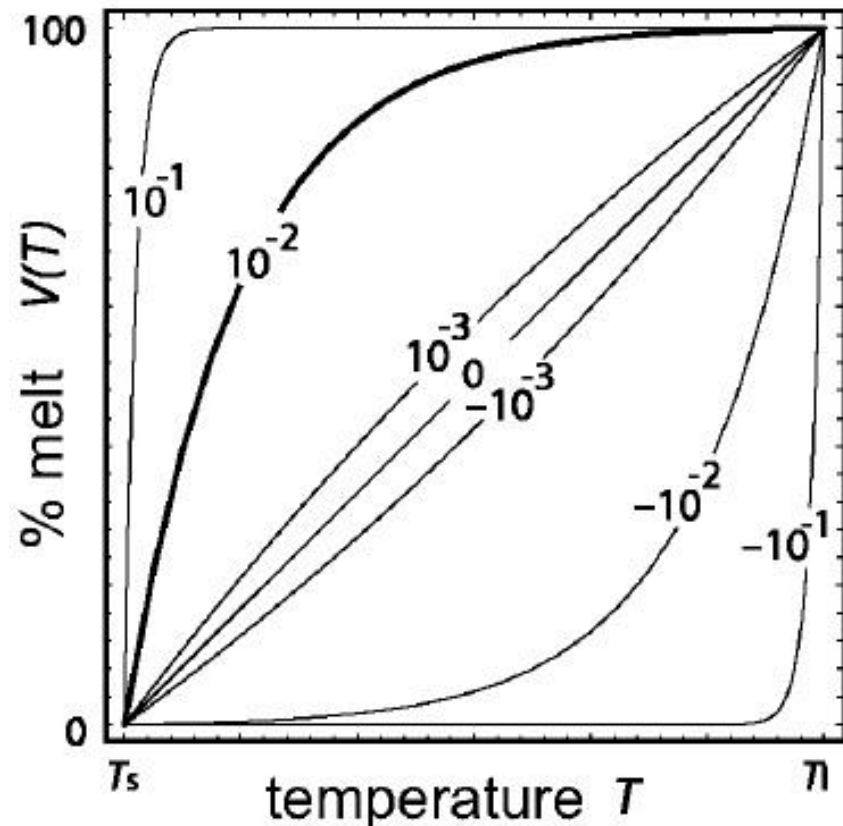
Dependence of the partial melt fraction on P and T



The solidus curve shows a negative slope between 70 and 100 km, where the amphibole changes into pyroxene, olivine and garnet and releases water.

Melt volume and Temperature

Relationship between melt volume and temperature in the melting interval:



Stuwe, 2002

$$V(T) = \frac{e^{aT} - e^{aT_s}}{e^{aT_l} - e^{aT_s}} \quad \frac{dV}{dT} = \left(\frac{a}{e^{aT_l} - e^{aT_s}} \right) e^{aT}$$

- If a has a large positive value, then most of the rock melts near the solidus.
- If a has large negative values, then most of the melting occurs near the liquidus.
- For values of $a \rightarrow$ zero, melting becomes linear in temperature: $V(T) = T/(T_l - T_s)$

Realistic values for a are positive and possibly around 0.01

T_s =Solidus temperature

T_l =Liquidus temperature

Magma Viscosity

Viscosity of a magma depends on its chemical composition (ionic or molecular cohesion):

- Basic melts have lower viscosity ($10-10^4$ Pa s) than acid melts ($10^{11}-10^{12}$ Pa s), since they have higher ratio between oxygen atoms (not forming structural bridges) and silicon.
- Peralkaline magma have low viscosity, since metal cations ($\text{Na}^+, \text{K}^+, \text{Ca}^+, \text{Mg}^+$) decrease the degree of melt polymerization (interrupt Si-O-Si-O chain).

Newton's relation

$$\sigma = \sigma_0 + \eta \left(\frac{d\varepsilon}{dt} \right)^n$$

σ_0 = shear stress required for the flow onset.

In Newtonian fluids $\sigma_0=0$ and $n = 1$.

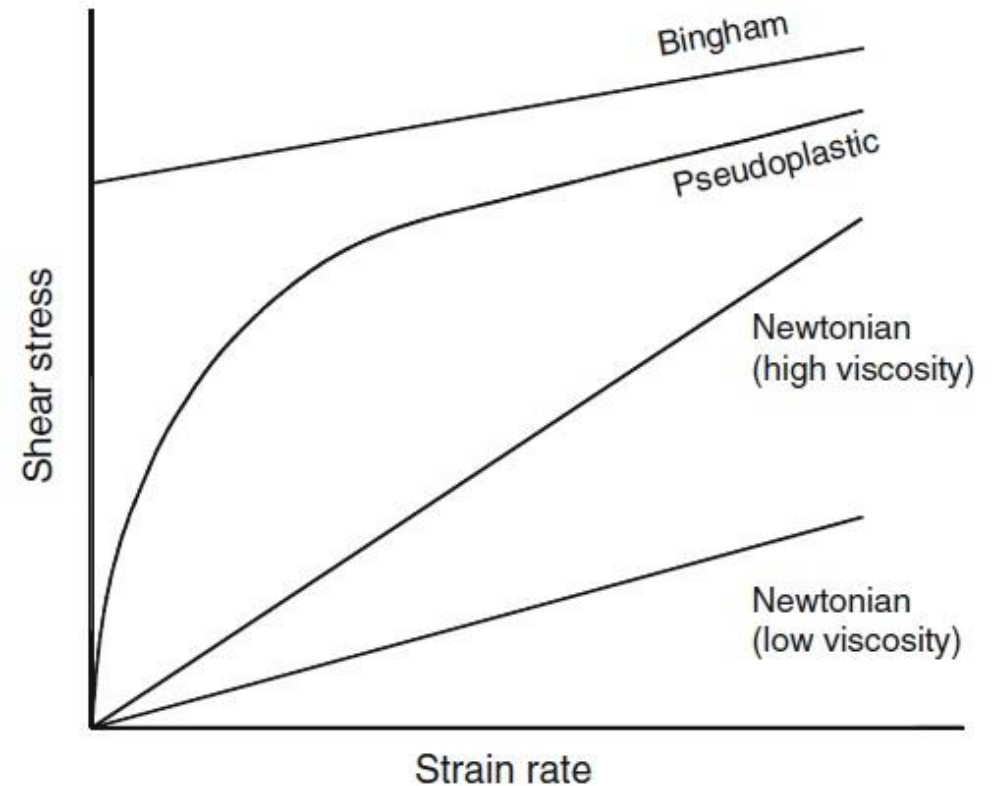
For a pseudoplastic magma (Non-Newtonian fluids) $\sigma_0 = 0$ and $n < 1$.

For a Bingham fluid σ_0 has a finite value and $n = 1$

For Non-Newtonian fluids:

$$\tau_s^n = A_{\text{eff}} \dot{\gamma} \quad \eta_{\text{eff}} = \frac{\tau_s}{\dot{\gamma}} = A_{\text{eff}}^{1/n} \times \dot{\gamma}^{(1/n)-1}$$

η =effective viscosity



Many rocks deform roughly 8 times as rapid if the applied stress is doubled

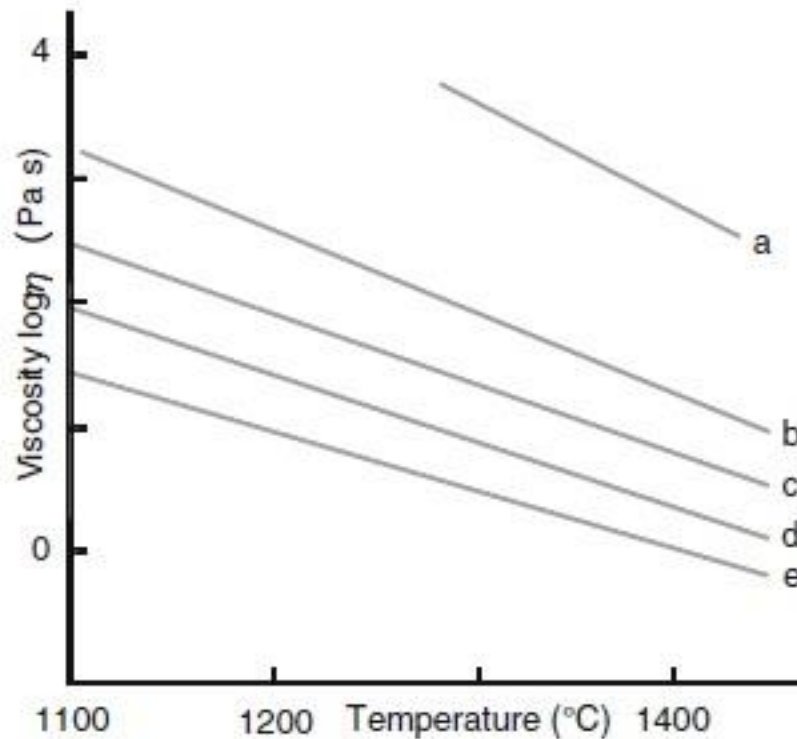
Magma Viscosity (P,T)

Dynamic viscosity variations with temperature and pressure p :

$$\eta = A_{\eta} \exp\left(\frac{E^* + p V^*}{RT}\right)$$

V^* = activation volume, R = gas constant and A_{η} constant material.

The activation energy E^* is mainly dependent on the ratio $(\text{Si} + \text{Al})/\text{O}$ of the melt.



a-pantellerite; b-basic andesite; c-oceanic island tholeiite; d-olivine basalt; e-olivine melanephelinite.

- At high pressure structural variations in silicate melts decrease the viscosity, due to the role of Al^{3+} (octahedral coordination is strongly favored at high pressure).

Magma Viscosity (gas, bubbles, and crystals content)

- Among gases, H₂O and CO₂ can be present in a significant concentration in a magma (magmas are less rich in HCl, HF, SO₂, SO₃, S, N₂).
- Solubility of CO₂ increases in presence of H₂O and causes an increase of polymerization.
- Effects of bubbles on magma viscosity depends on the degree of vesiculation, the size and distribution of bubbles, and the melt viscosity (influence viscosity of acidic magma).
- Suspended crystals in magma increases the viscosity, the critical threshold is ~55 %: below effective potential of diffusion in the form of lava, above potential to become intrusive bodies.

From lab experiments:
$$\eta = \eta_L (1 - 1.67 x_s)^{-2.5}$$

x_s = volume fraction of suspended solid and η_L = viscosity of the liquid

Lava Flows

Transition between laminar and turbulent flow can be estimated from the **Reynolds number (R_e)**

$$R_e = \frac{\rho v D}{\eta}$$

where v is the mean velocity, ρ the density and D is the diameter of a circular conduit

Transition between laminar and turbulent flow occurs at values of R_e between 500 and 2000

For Bingham fluids:

$$R_e \geq 1000 \quad B = \frac{\tau_o D}{\eta v} \quad \text{Hampton Number: } Re/B = \frac{\rho v^2}{\tau_o} \geq 1000$$

with τ_o the mechanical strength of the material

Only lava with low viscosity can have a turbulent flow (when the ground slope increases the already high velocity of the lava).

$$\eta = \frac{g \rho h^2 \sin i}{n v}$$

where i is the ground inclination, h and v is the thickness and velocity of the lava flow, g is the gravity acceleration, ρ is the density, and $n = 3$ (for thick lava flows) up to 4 (for thin lava flows).

Magma Upwelling

Magma ascension occurs because of:

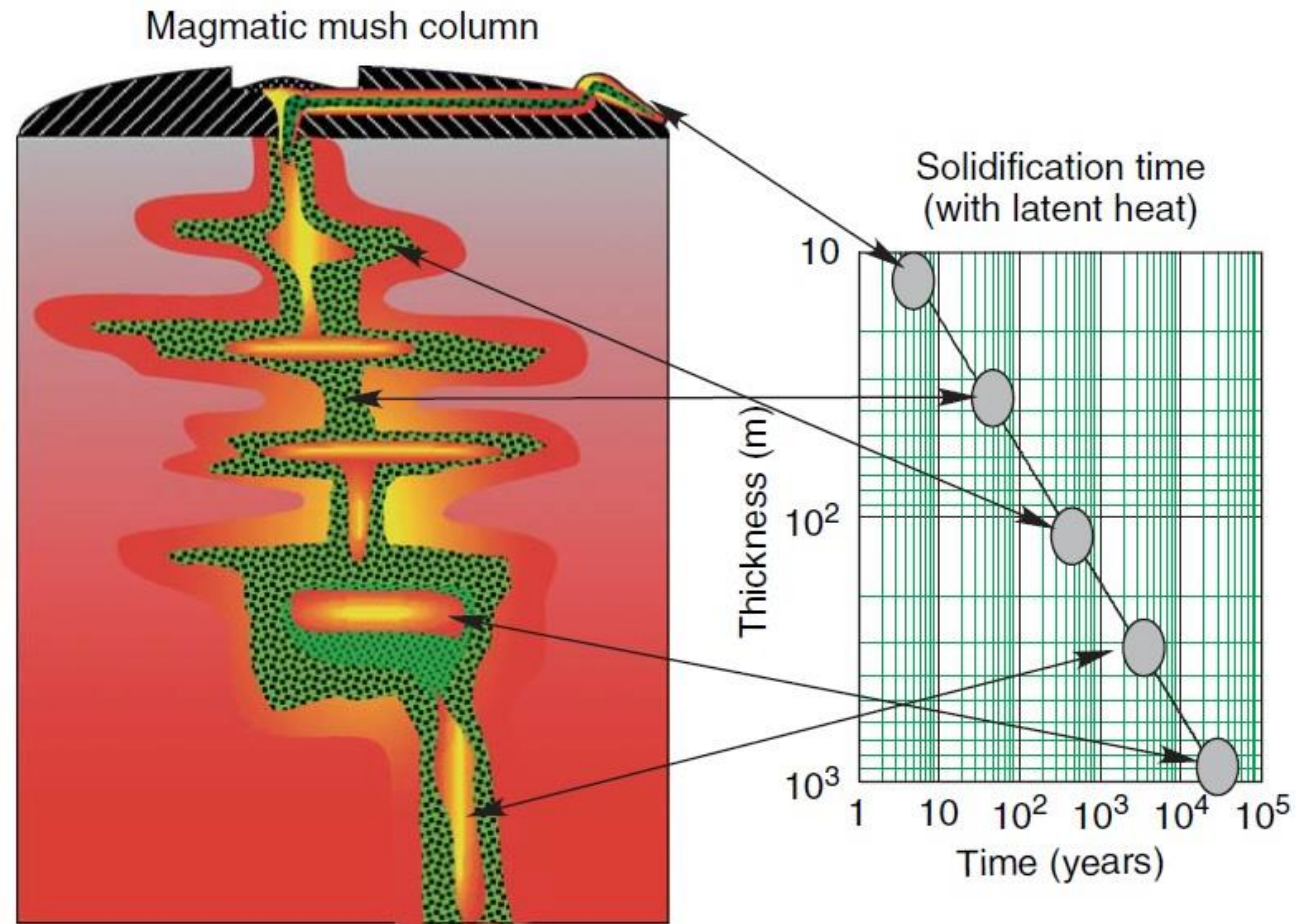
- Difference in density (due to T) with the hosting rocks (diapirs).
- Migration through fractures generated by magma pressure.

- Fractures move upwards and reach a critical length depending on (1) the elastic properties of the medium (2) fracture size (3) density contrast between the magma and the medium.
- Fractures formation is favored by gases (besides high temperature and water), which would corroder the rocks, reducing the energy required for fracturing the medium.

The rate of propagation of dykes is controlled by the rate of the fracturing at the tip and by the flow rate of magma inside the dyke:

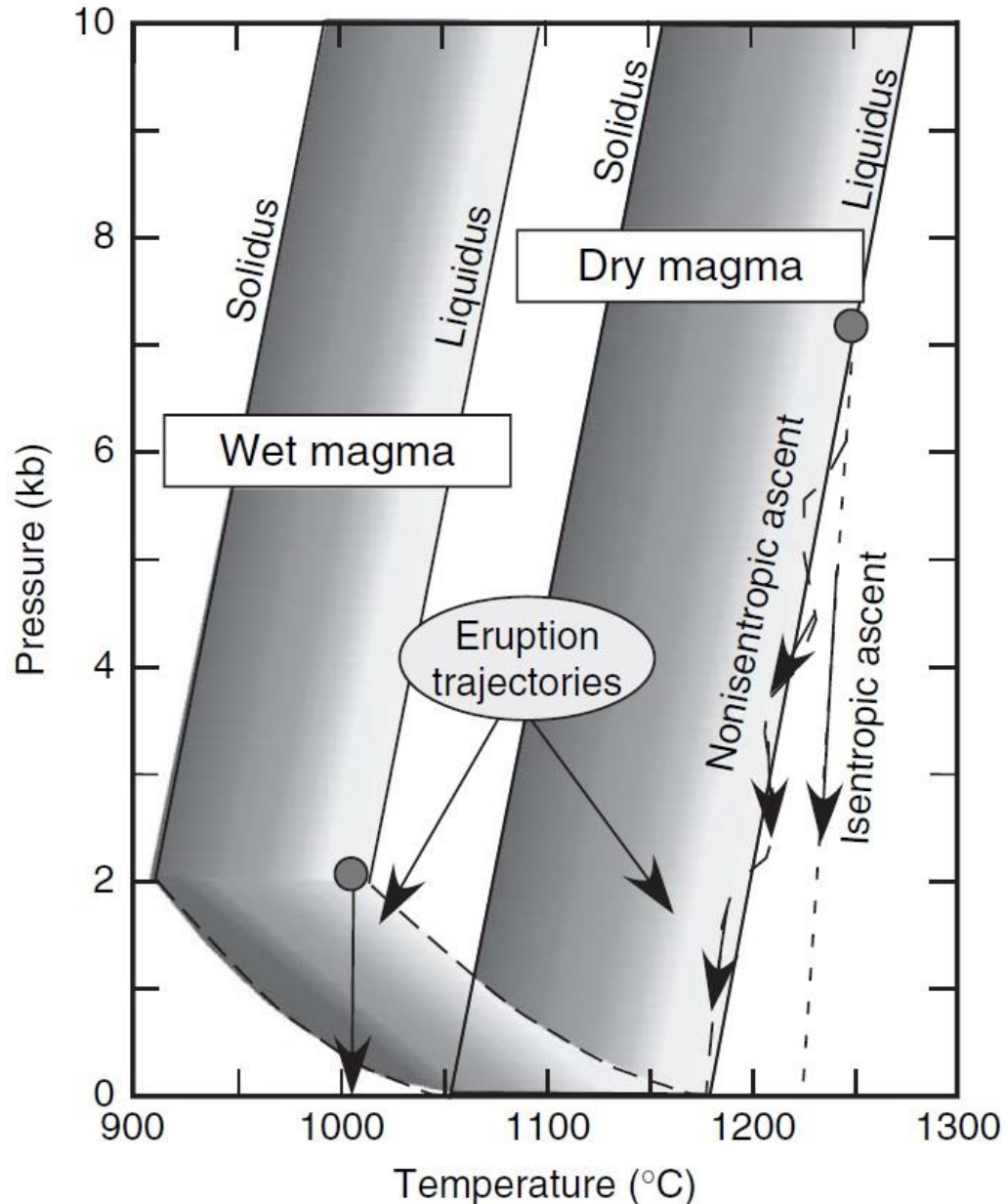
- (1) When high energy is needed to fracture the host rock and magma viscosity is low, the rate of propagation is controlled by the rate of fracturing (fracture controlled regime).
- (2) When the energy needed to fracture the host rock is low, propagation is controlled by the magma flow rate (magma-controlled regime).

Magmatic Systems



- The thermal timescales for solidification vary throughout the system due to the local thermal regime and the length scale of the system.

Phase equilibria for a wet and dry magma



- The solubility of volatiles is directly proportional to the confining pressure and there is also a point at which no more water can be dissolved in the magma.
- At $P > 2$ kb (200 Mpa), corresponding to the saturation point, the wet magma is undersaturated with water and the phase diagram is similar to that of the dry magma (just at lower T).

Volcanic Eruptions

- The magma starts to rise when the pressure P within the magma becomes larger than the lithostatic pressure P_0 ($P > P_0$).
- Lower pressure are required for the basaltic magma, due to their lower density.
- Higher temperature and a decrease of volume of the magma leads to an increase of pressure compared to the lithostatic value (increase in excess pressure):

$$\alpha = \frac{\left(\frac{\partial V}{\partial T}\right)_P}{V_0} \quad \beta = -\frac{\left(\frac{\partial V}{\partial P}\right)_T}{V_0} \quad \frac{dV}{V_0} = -\beta dP + \alpha dT \quad P = P_0 + \frac{\alpha}{\beta}(T - T_0) - \frac{1}{\beta} \frac{\Delta V}{V_0}$$

$$P_T = \frac{\alpha}{\beta}(T_m - T_0) \quad P_c = -\frac{1}{\beta} \frac{\Delta V}{V_0} \quad P_c = \frac{1}{\beta} \frac{\Delta \sigma}{\sigma}$$

P_T and P_c are the additional pressure due to the excess temperature and decrease of volume during magma formation (a magma has a lower density than when is solid and thus needs more space).

$$P = P_0 + P_T + P_c \quad P - P_0 = P_T + P_c \quad P - P_0 > 0$$

Unloading pressure ($\Delta V/V_0$) depends on the volume of the magma chamber V_0 (ΔV must increase for large magma chamber).

Density variations due to melting

Changes in densities of some typical magmatic rocks upon melting

| Rock | Density of molten rock (in kg/m ³) | Density of solid rock (in kg/m ³) | Relative change of density on fusion ($\Delta\sigma/\sigma$) |
|------------------------|--|---|--|
| Rhyolites ^a | 2,170 | 2,280 | −0.0482 |
| Andesite ^a | 2,410 | 2,590 | −0.0695 |
| Tholeiite basalt | 2,600 | 2,760 | −0.0580 |
| Alkali olivine basalt | 2,680 | 2,830 | −0.0530 |
| Diabase ^a | 2,640 | 2,960 | −0.1081 |
| Basalt, gabbro | 2,600 | 2,900 | −0.1034 |
| Rhyolite, granite | 2,300–2,400 | 2,600 | −(0.0769–0.1154) |
| Diabase ^a | 2,603 | 2,890 | −0.0993 |
| Diabase ^a | 2,640 | 2,880 | −0.0833 |
| Rhyolite ^a | 2,310 | 2,450 | −0.0571 |
| Andesite ^a | 2,580 | 2,830 | −0.0883 |
| MORB | 2,650 | 2,950 | −0.1017 |
| Mars basalts | 2,720 | 3,130 | −0.1310 |
| Komatiites | 2,730 | 3,090 | −0.1165 |
| Basalt | 2,600 | 2,800 | −0.0714 |
| Silicic melt | 2,300 | 2,600 | −0.1154 |

- The process of magma formation requires a change in volume of about 4.8–13.1 % (the density of magmatic rocks is 10–20 % greater than their corresponding melts).

Further values of density variations: -0.2012 for both diopside and for enstatite, -0.0462 for anorthite, -0.0779 for forsterite and -0.1138 for fayalite.

Thermal expansion coefficient (α) and compressibility coefficient (β)

Thermal expansion coefficient (α) and compressibility coefficient (β) values for some typical magmatic melts

| Molten rock | Thermal expansion coefficient (α), 10 ⁻⁵ /K | Compressibility coefficient (β), 1/GPa | References |
|--|---|--|------------------------------|
| Average silicate melt | 3 | 0.07 | McBirney (2000) |
| Average silicate melt | 3 | 0.07 | Best (2002) |
| Kilauea gas poor magma | – | 0.06–0.10 | Rivalta and Segall (2008) |
| Rhyolite ^a | 7.1 | – | Guillot and Sator (2007) |
| Andesite ^a | 9.8 | – | Guillot and Sator (2007) |
| MORB ^a | 11.1 | – | Guillot and Sator (2007) |
| Mars basalts ^a | 12.3 | – | Guillot and Sator (2007) |
| Komatiites ^a | 13.1 | – | Guillot and Sator (2007) |
| Rhyolite | 10.1 | – | Nelson and Carmichael (1979) |
| Rhyolite | 3.8 | – | Lange and Carmichael (1987) |
| Rhyolite | 5.3 | – | Ghiorso (2004) |
| Andesite | 8.1 | – | Nelson and Carmichael (1979) |
| Andesite | 5.5 | – | Lange and Carmichael (1987) |
| Andesite | 6.0 | – | Ghiorso (2004) |
| MORB | 10.6 | – | Nelson and Carmichael (1979) |
| MORB | 6.4 | – | Lange and Carmichael (1987) |
| MORB | 7.0 | – | Ghiorso (2004) |
| Komatiites | 4.7 | – | Courtial et al. (1997) |
| Komatiites | 9.7 | – | Ghiorso (2004) |
| Komatiites | 8.1 | – | Lange and Carmichael (1987) |
| Peridotite | 6.2 | 0.05587 | Suzuki et al. (1998) |
| Peridotite | 5.9 | 0.05649 | Suzuki et al. (1998) |
| Basalt | – | 0.1 | Rivalta and Segall (2008) |
| Mount St. Helens Magma | – | 0.28 ^b | Mastin et al. (2008) |
| Magma, Soufrière Hills Volcano, Montserrat | – | 0.758 ^c | Voight et al. (2010) |

^a At a temperature of 1,800 K and low pressure

^b Magma contains ~ 1.2 % bubbles

^c Saturated magma plus exsolved gas

$$\frac{\Delta P}{\Delta T} = \frac{\alpha}{\beta} \text{ (e.g., or the average silicate melt ratio } \alpha/\beta = 0.43 \text{ MPa/K)}$$

Average values of this ratio for basaltic and rhyolitic magmatic rocks are 1.571 and 0.943 MPa/K

- Presence of gases and water in magma significantly increases β
- Both α and β depend on the magma composition and its $T - P$ conditions.

Solidus/Liquidus Temperature for igneous rocks

Melting point/solidus temperature (in K) and liquidus temperature (in K) of most typical magmatic rocks for dry and wet conditions at pressures of 0.1 MPa and 1 GPa

| Rock | Melting point or solidus temperature of dry rock at 0.1 MPa/1 GPa | Melting point or solidus temperature of wet rock at 0.1 MPa/1 GPa | Melting point or solidus temperature of saturated rock at 0.1 MPa/1 GPa | Liquidus temperature of dry rock 0.1 MPa/1 GPa | Liquidus temperature of wet rock 0.1 MPa/1 GPa |
|------------------|---|---|---|--|--|
| Granites | 1,233/– | 1,233/953 ^a | | 1,569/– | |
| Granite | 1,223/1,333 | | 1,223/893 | 1,293/– | 1,293/983 |
| Granite | 1,153/1,333 | | 1,153/893 | | 1,293/993 |
| Andesite | 1,263/1,343 | | –/903 | –/1,513 | 1,473/1,213 |
| Tholeiite basalt | 1,353/1,413 | 1,303/913 | | 1,483/1,533 | 1,463/1,463 |
| Basalt | 1,353/1,413 | | 1,343/913 | –/1,523 | –/1,333 |
| Peridotite | 1,403/1,493 | | 1,403/1,153 | | |
| Peridotite | 1,518/1,587 | | | | |
| Rhyolite | 943–953/– | 923/– | 923–973/– | | |
| Peridotite | 1,363/1,613 ^b | –/1,353 | | | –/1,523 |
| Peridotite | 1,400/1,700 ^b | 1,340/1,270 ^b | | | |
| Basalts | 1,525 ^c /– | | | | |
| MORB | 1,516–1,624 ^d /– | | | | |
| OIB | 1,559–1,645 ^d /– | | | | |
| Picrite | –/1,798 | | | | |
| Komatiite | 2,063/– | | | | |
| Tonalite | 1,173/– | | 1,153/903 | | 1,293/993 |
| Peridotite | 1,405/1,535 | | | | |
| Quartz eclogite | –/1,588 ^b | | | –/1,773–1,798 ^b | |

^a At 200 MPa

^b At ~3 GPa

^c Eruption temperature of most basalts (Schubert et al. 2001)

^d The parental liquids (Falloon et al. 2007)

- At a depth of about 100 km, peridotite begins to melt at ~1,073 K (with an excess of water), while in the absence of water at ~1,773 K.
- T_s of a Granitic melt drops from ~1,233 K with no water to 953 K with a water content increase to ~6.0 wt%.
- The solubility of H₂O in a granitic melt increases from 0 to ~ 6.0 wt% with an increase of pressure from 0 to 200 MPa.
- In most cases for water-rich magmas these parameters have a negative gradient of $\Delta T/\Delta P < 0$ within a P range of 0–1 GPa. The gradient is usually negative up to a pressure of about 2 GPa and stabilizes at higher values of P and can also starts to slightly increase.

Conditions in forming and rising basaltic magma

P (0-4 GPa), T (1600-1900 K), α ($8.8 \times 10^{-5} \text{ K}^{-1}$), β (0.056 GPa^{-1})

Thermodynamic conditions within forming and rising magma of basaltic composition

| Temperature of magma (K) | Lithostatic pressure (P_0) (GPa) | Normal temperature (T_0) (K) | P_T (GPa) | P_C (GPa) | Total P (GPa) |
|--------------------------|--------------------------------------|----------------------------------|-------------|--------------|-----------------|
| 1,600 | 4 | 2,818 | -1.911 | 1.00-2.16 | 3.09-4.25 |
| | 3 | 2,182 | -0.913 | 0.24-1.42 | 2.33-3.51 |
| | 2 | 1,546 | 0.085 | -0.48-0.64 | 1.61-2.63 |
| | 1 | 909 | 1.084 | -(1.24-0.12) | 1.16-1.96 |
| | 0.5 | 591 | 1.583 | -(1.62-0.88) | 0.46-1.20 |
| | 0 | 273 | 2.082 | -(2.00-1.26) | 0.08-0.82 |
| 1,700 | 4 | 2,818 | -1.754 | 1.00-2.16 | 3.25-4.31 |
| | 3 | 2,182 | -0.752 | 0.24-1.42 | 2.49-3.67 |
| | 2 | 1,546 | 0.242 | -0.48-0.64 | 1.76-2.88 |
| | 1 | 909 | 1.241 | -(1.24-0.12) | 1.00-2.12 |
| | 0.5 | 591 | 1.740 | -(1.62-0.88) | 0.62-1.36 |
| | 0 | 273 | 2.238 | -(2.00-1.26) | 0.24-0.98 |
| 1,800 | 4 | 2,818 | -1.597 | 1.00-2.16 | 3.40-4.56 |
| | 3 | 2,182 | -0.599 | 0.24-1.42 | 2.64-3.82 |
| | 2 | 1,546 | 0.399 | -0.48-0.64 | 1.92-3.04 |
| | 1 | 909 | 1.398 | -(1.24-0.12) | 1.16-2.28 |
| | 0.5 | 591 | 1.897 | -(1.62-0.88) | 0.78-1.52 |
| | 0 | 273 | 2.396 | -(2.00-1.26) | 0.40-1.14 |
| 1,900 | 4 | 2,818 | -1.440 | 1.00-2.16 | 3.56-4.72 |
| | 3 | 2,182 | -0.442 | 0.24-1.42 | 2.80-3.98 |
| | 2 | 1,546 | 0.555 | -0.48-0.64 | 2.08-3.20 |
| | 1 | 909 | 1.555 | -(1.24-0.12) | 1.32-2.44 |
| | 0.5 | 591 | 2.054 | -(1.62-0.88) | 0.93-1.67 |
| | 0 | 273 | 2.553 | -(2.00-1.26) | 0.55-1.29 |

$$\Delta P_T = \frac{\alpha}{\beta} (T_{02} - T_{01}) \quad P_c = \frac{1}{\beta} \frac{\Delta \sigma}{\sigma}$$

T_{02} and T_{01} = normal T of magma at two different depths

- Total P does not consider the unloading of pressure caused by the increase of magma volume as it rises through the conduit ($\Delta V_c = 0.179$ GPa and 0.143 GPa per 1 % of increased volume for basaltic and rhyolitic magmas, respectively) and the erupted magma volume (ΔV_e equal to the real volume of a magmatic eruption by a volcano).
- P_c is positive at large depth (> 2 GPa), where the melt cannot expand.
- P_T increases with decreases depth, since T_0 of magma lowers with decreases depth (at pressure > 2 GPa the effect of β overcomes that of α).
- At $T \sim 1900$ K and $P = 2$ GPa the total pressure exceeds the lithostatic pressure of ~ 0.08 GPa, but it is not sufficient for rising (to overcome the friction of magma or its small expansion).
- In cases of rock with a very large difference of density between its solid and molten states ($\sigma \sim 0.1212$), the excess in pressure would be high enough for the magma to begin rising from depths with a pressure of 4 GPa or lower.
- For volcanoes as high as 2, 3, 4, and 5 km, the minimal additional P for the basaltic magma rising to the surface would be 0.051, 0.076, 0.101, and 0.127 GPa, respectively.

Conditions in forming and rising rhyolitic magma

P (0-1 GPa), T (1000-1500 K), α ($6.6 \times 10^{-5} \text{ K}^{-1}$), β (0.07 GPa^{-1})

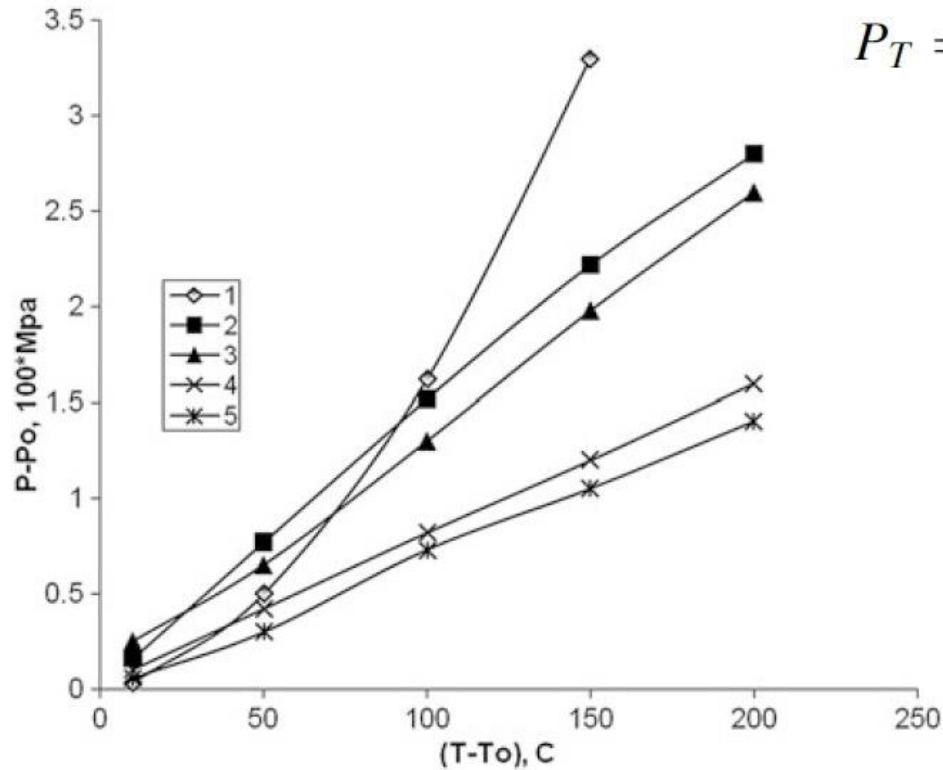
Thermodynamic conditions within forming and rising magma of rhyolitic composition

| Temperature of magma (K) | Lithostatic pressure (P_0) (GPa) | Normal temperature (T_0) (K) | P_T (GPa) | P_C (GPa) | Total P (GPa) |
|--------------------------|--------------------------------------|----------------------------------|-------------|-------------|-----------------|
| 1,500 | 1 | 1,333 | 0.16 | 0.72–1.86 | 1.88–3.02 |
| | 0.7 | 1,012 | 0.46 | 0.53–1.67 | 1.69–2.73 |
| | 0.5 | 803 | 0.66 | 0.40–1.54 | 1.56–2.70 |
| | 0.3 | 591 | 0.86 | 0.27–1.41 | 1.43–2.57 |
| | 0 | 273 | 1.16 | 0.08–1.22 | 1.24–2.38 |
| 1,200 | 1 | 1,333 | –0.13 | 0.72–1.86 | 1.59–2.73 |
| | 0.7 | 1,012 | 0.18 | 0.53–1.67 | 1.41–2.55 |
| | 0.5 | 803 | 0.37 | 0.40–1.54 | 1.27–2.41 |
| | 0.3 | 591 | 0.57 | 0.27–1.41 | 1.07–2.28 |
| | 0 | 273 | 0.87 | 0.08–1.22 | 0.95–2.09 |
| 1,000 | 1 | 1,333 | –0.31 | 0.72–1.86 | 1.41–2.55 |
| | 0.7 | 1,012 | –0.01 | 0.53–1.67 | 1.22–2.36 |
| | 0.5 | 803 | 0.19 | 0.40–1.54 | 1.09–2.21 |
| | 0.3 | 591 | 0.39 | 0.27–1.41 | 0.96–2.10 |
| | 0 | 273 | 0.69 | 0.08–1.22 | 0.77–1.91 |

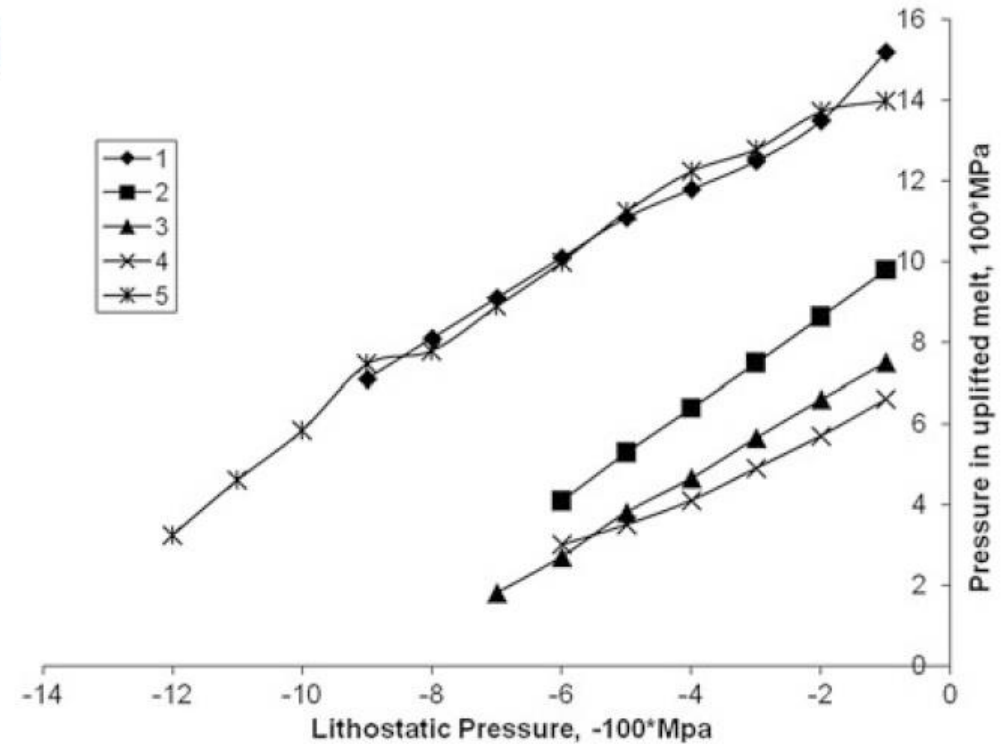
- P_c is positive at large depth, where the melt cannot expand
- P_T increases with decreases depth, since T_0 of magma lowers with decreases depth.
- For volcanoes as high as 2, 3, 4, and 5 km, the minimal additional pressure for them to reach the craters would be 0.043, 0.064, 0.085, and 0.107 GPa, respectively.
- The excess P is much higher than within basaltic magmas (rising magma is favored), making rhyolitic magmas more destructive (caldera formation).

Excess of Pressure due to Temperature

$$P_T = \frac{\alpha}{\beta} (T - T_0)$$



1 wet clay, 2 limestone, 3 granite, 4 sandstone, 5 basalt

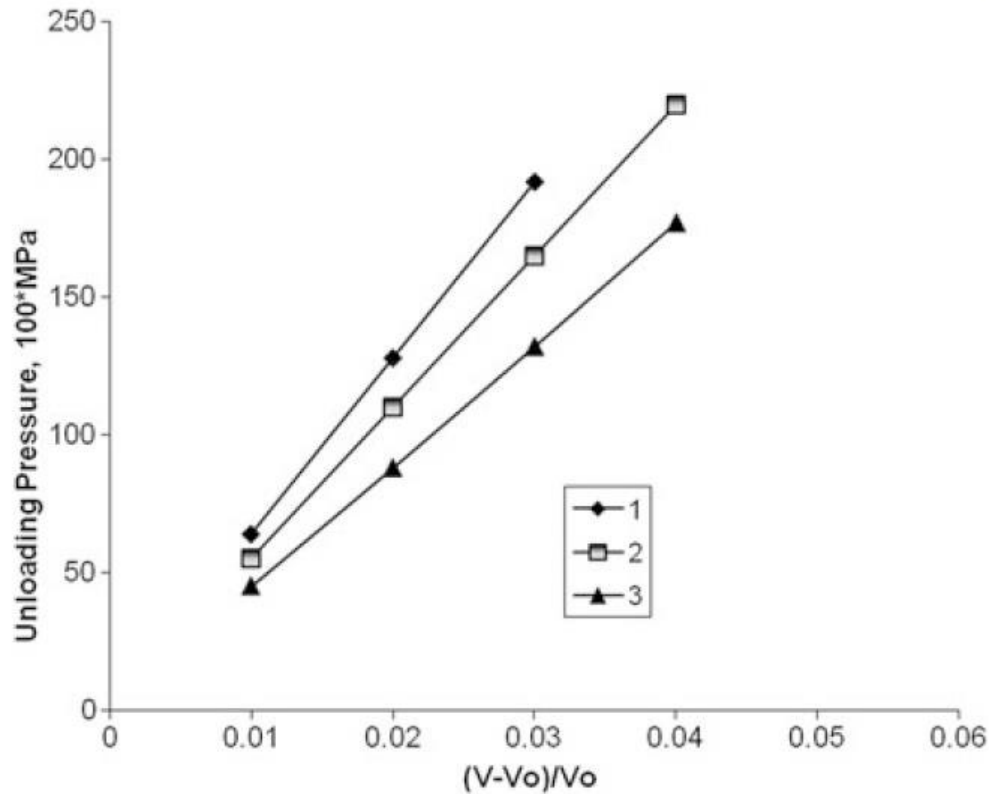


1 gabbro, 2 granite, 3 basalt, 4 diorite, 5 diabase

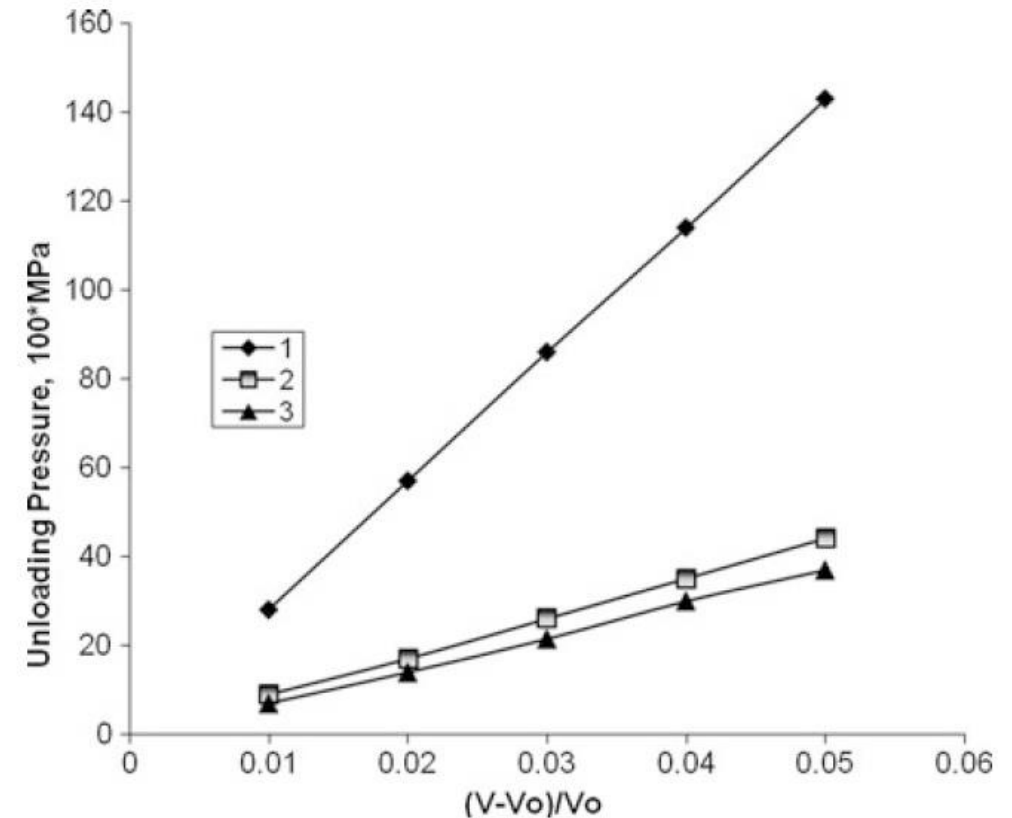
- Temperatures in excess of 100 K generate greater P increases in wet clays than in other rocks because of the large value of α_{water} (other rocks have larger viscosity and lower chances to expand).
- Heating of wet clays, salt and/or porous water high pressure results in the formation of mud volcanoes, geysers, fumaroles, etc.
- Additional increase in P caused by rising magma (due to the decrease of T_0 with the decrease in depth) is the largest in gabbro-diabase compositions and lowest in basalt-diorite melts.

Unloading/Loading of Pressure

$$P_U = \frac{1}{\beta} \frac{\Delta V}{V_0}$$



1 limestone, 2 granite, and 3 basalt



1 sandstone, 2 basaltic melt, and 3 wet clay

- Wet clay and basaltic melt in (mud) volcano eruptions have a relatively slow pressure unloading with increasing volume (they retain high excess pressure during eruption for a longer period).
- In compressional environment wet clay and basaltic melt acquire pressure much slower than granite.

Mud Volcanoes

It is a tectonic structure formed during the formation and eruption of significant amounts of water-rich sedimentary rocks (mostly clays and/or silts)



Mud Volcanoes

Conditions for mud volcano formation:

- A thick sedimentary cover (several kilometers) and plastic clayey components with an anomalously high formation of pore pressure (greater than the lithostatic pressure) and the presence of thermal water.
- Many mud volcanoes form in regions of volcanic activity (e.g., Pacific Ring of Fire) and are associated to high T within the crust (which gives excess T under the main plastic clay-rich layer) and regions characterized by serpentinization and salt layers.
- Temperature in a mud volcano chamber is often $> 300\text{ }^{\circ}\text{C}$ and causes a large increase of P : we should consider that the clay waterproof cap layer has a very small heat conductivity coefficient and high specific heat capacity, which can lead to the accumulation of huge amounts of heat energy below it.
- Clay material can reach the surface, due only to their low density, if the presence of pressure in the chamber is greater than the lithostatic pressure exerted by the weight of the column of clay rocks from the chamber to the surface. This would happen only if volcanoes chamber are 4-5 times greater than its height.
- Mud volcanoes require the presence of huge compressive tectonic forces, which increase fluid pressure, and faults below the thick plastic clay-rich layer, which can act as a pathway for upward migration of overheated fluids. Then, they are often associated to large earthquakes, which favour faults formation.

Mud Volcanoes

Thermo-physical properties of clay:

- The high porosity (~45 %) and very low permeability prevents the water from leaving the clay.
- The high water content decreases the clay viscosity to values of 10^6 Pas and favours the transfer of the pressure applied to the clay in all directions.
- The clays have a very low low heat conduction coefficient and a very high thermal expansion coefficient (α_c) and specific heat capacity. Then, they accumulate large amounts of heat energy, which remains trapped due to their low permeability.

$$a_c \approx K_p \cdot a_w$$

K_p = porosity coefficient and α_w = thermal expansion coefficient of the water (two orders of magnitude higher than for others rocks).

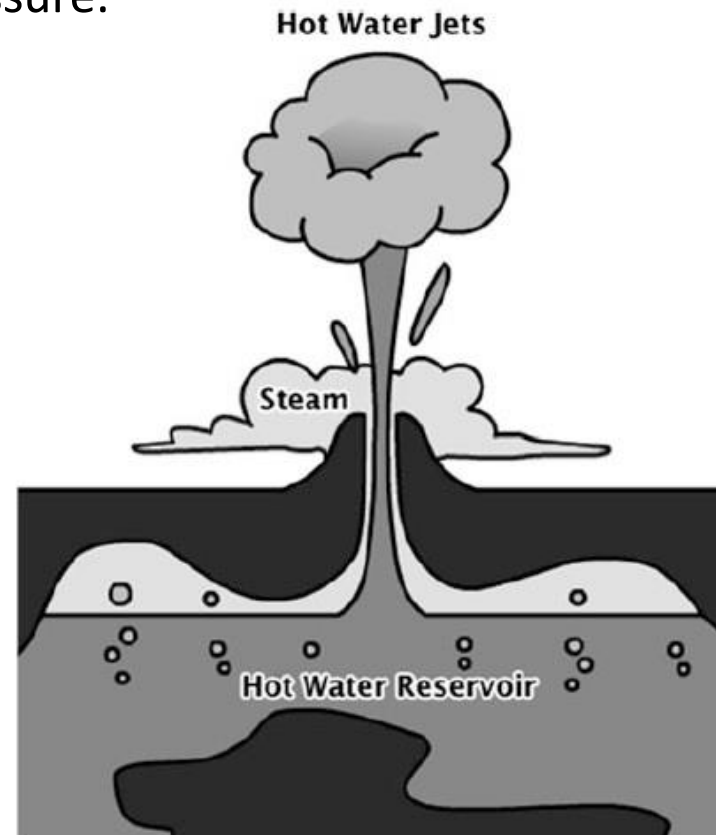
- This increase of volume and pressure is transferred through the plastic clay to all parts of the layers, exerting great pressure on the overburden, which can overcome lithostatic pressure and cause the rise of a diapir, with a mud volcano formation.

Magma volatiles

- The solubility of the volcanic gases in a magma increases with P and decreases with T , while their concentration depends mostly on the rock composition through which the magma rises.
- The solubility of water in magmas can reach 9–12 wt%, while that of CO_2 reaches values of 0.5–4.0 wt%, but drops to zero at 0.1 MPa.
- Saturated magma continues the decomposition of rocks and minerals during its rises and stores the gases within the porous space formed in it: during the decompression of magma with an initial dissolved gas content of 5–7 wt%, the bubbles could occupy up to 99 % of its volume.
- Magmatic volatiles are composed of mostly H_2O , CO_2 , and SO_2
- In rhyolites containing between 69 and 78 wt% Si, water is the most abundant (3-7 wt%) and follow CO_2 up to 1,000 ppm, chlorine 600 to 2,700 ppm, fluorine 400 to 1,500 ppm, and sulfur ~200 ppm.
- The low concentration of SO_2 in the rhyolitic magma is due to the fact that SO_2 within the crust is a product of the oxidation of elemental sulfur released during the decomposition of sulfides at high temperatures (when they come in contact with basaltic magmas).

Thermal waters, Hot springs, Geysers, Fumaroles

- Thermal water is related to the circulation of groundwater in the vicinity of sources of heat within the crust.
- A hot spring is a natural discharge of groundwater with a temperature of about 6.5 C or more above the mean air temperature.
- Reservoir temperatures for hot-water systems are possibly divided into low temperature (<363 K), intermediate temperature (from 363 to 423 K), and high temperature (from 423 to 513 K).
- A geyser is a hot spring with a mixture of steam and water. The boiling point of water at the discharge area is mostly dependent on the atmospheric pressure.



Temperature indicators from groundwater geochemistry

The water solubility of many geochemical compounds (e.g., silica) increases with T and their amount can be used to estimate T :

$$T = [1315 / (5.205 - \log_{10} \text{SiO}_2)] \pm 0.5^\circ\text{C} \quad (1) \quad T \text{ range of precision: } 125^\circ\text{-}250^\circ\text{C} \quad \text{SiO}_2 \text{ (ppm by weight)} \quad T(\text{K})$$

Swanberg and Morgan (1979)

$$T = [1533.5 / (5.768 - \log_{10} \text{SiO}_2)] \pm 2.0 \quad (2)$$

$$T = [1015.1 / (4.655 - \log_{10} \text{SiO}_2)] \pm 2.0 \quad (3)$$

For sodium (Na), potassium (K), and calcium (Ca):

$$T = [855.6 / (0.8573 + \log_{10}(\text{Na}/\text{K}))] \pm 2.0$$

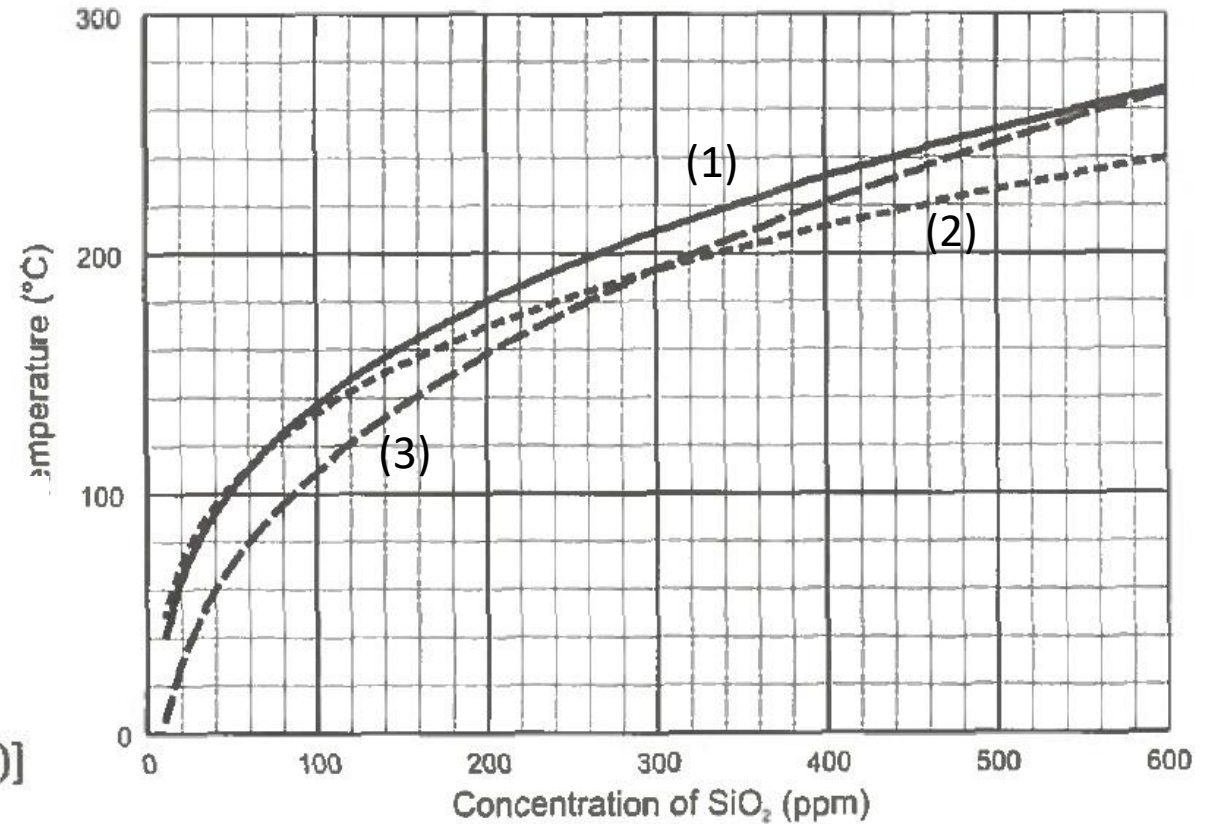
$$T = [846 / (0.5964 + \log_{10}(M_{\text{Na}}/M_{\text{K}}))]$$

$$T = [1647.3 / (2.24 + \log_{10}(M_{\text{Na}}/M_{\text{K}}) + \beta \log_{10}(\sqrt{M_{\text{Ca}}}/M_{\text{Na}}))]$$

M_X = molar concentration of element X (i.e. moles per litre)

$\beta = 4/3$ for $(\sqrt{M_{\text{Ca}}}/M_{\text{Na}}) > 1$ and $T < 373.15 \text{ K} (100^\circ\text{C})$

$\beta = 1/3$ for $(\sqrt{M_{\text{Ca}}}/M_{\text{Na}}) < 1$ or $T > 373.15 \text{ K} (100^\circ\text{C})$

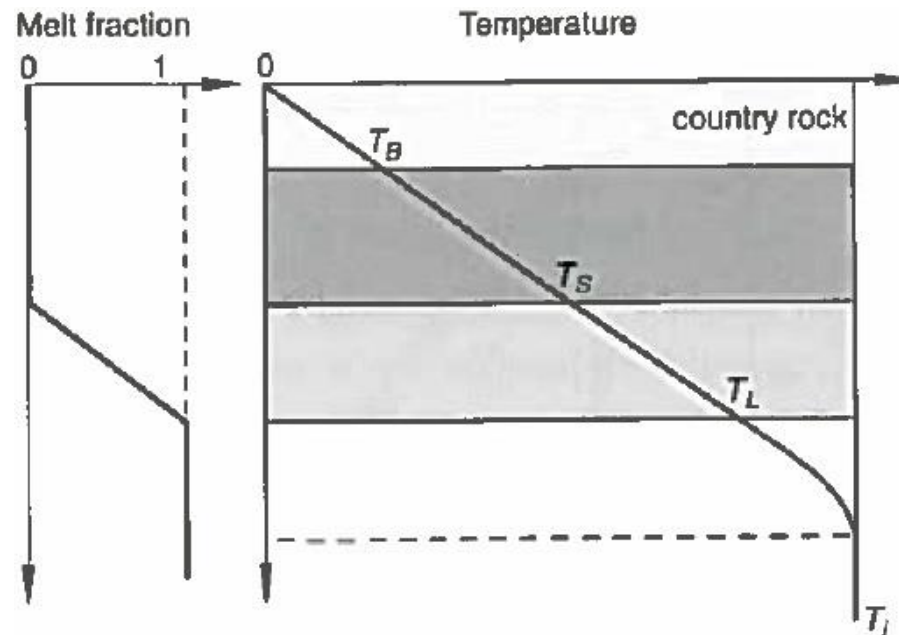


Magma cooling

- Cooling is the driving mechanism for crystallization and differentiation. It determines how, where, and how quickly crystals nucleate and grow.
- Most crystals are generated in the cold thermal boundary layers that develop at the margins of a magma body, where there are strong gradients of temperature, crystal content, and viscosity.

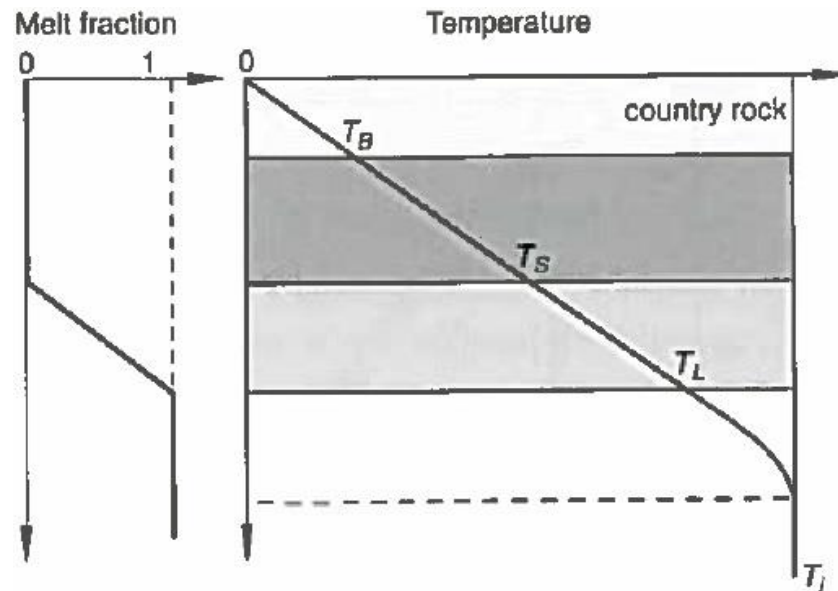
The thermal structure at the margin of a magma body is split in three different parts:

- 1) The part below the solidus is fully solid and may include a chill phase.
- 2) The part between the solidus and liquidus (magma is partially crystallized).
- 3) The part below the liquidus made of superheated magma (its existence depends on the initial conditions).



Thermal structure of magma bodies

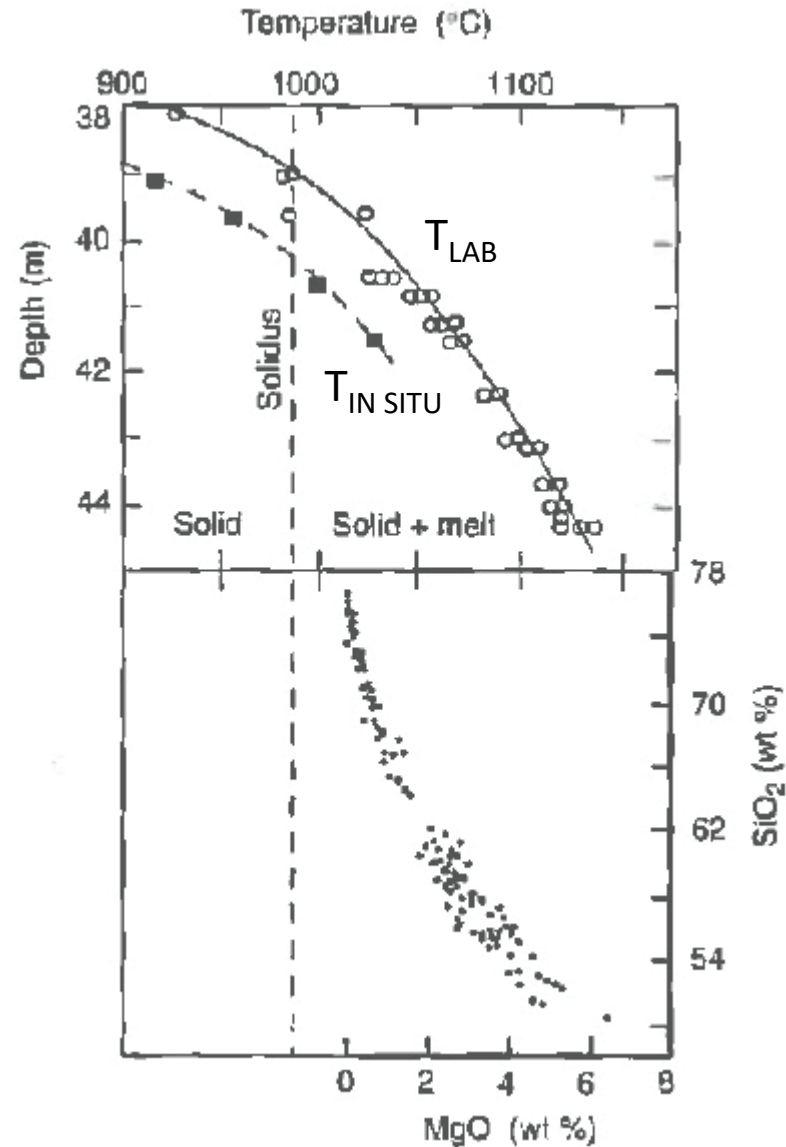
- The partially crystallized zone, may be treated as an equivalent liquid with properties depending on the crystal content and composition of interstitial liquid.
- This layer is separated by a rheological transition occurring for a solid fraction of ~55%: above this threshold, crystals are interconnected and immobile (cannot participate to convection), below it, crystals are suspended within a continuous liquid phase.
- Depending on the initial melt composition, the interstitial liquid may be less dense or denser than the uncrystallized magma.
- In a magma reservoir convection occurs, since cooling generates a negative density contrast, potentially unstable despite the magma composition ($\Delta\rho = -\rho_i\alpha\Delta T$). For $\alpha=3\times 10^{-5} \text{ K}^{-1}$, $\kappa=10^{-6}\text{m}^2\text{s}^{-1}$, $\Delta T=10 \text{ K}$, $\eta_m=10 \text{ Pa s}$, the upper layer is at least 10 cm and the magma layer 40 cm, to get $N_U=2$.



Partially crystallized magma layer

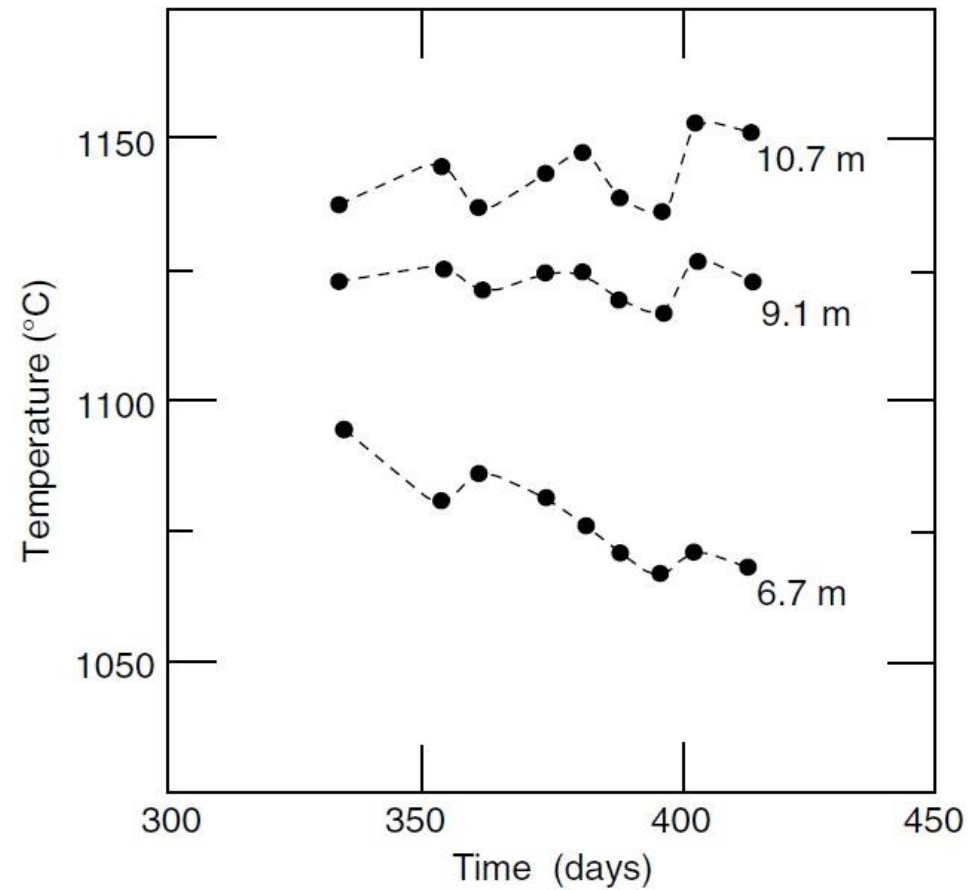
Kilauea Iki lava lake

The difference in temperature between the lab experiments and the *in situ* partially crystallized lava is < 15 K (crystals and coexisting melt are in local thermodynamic equilibrium).



Convection in magmas

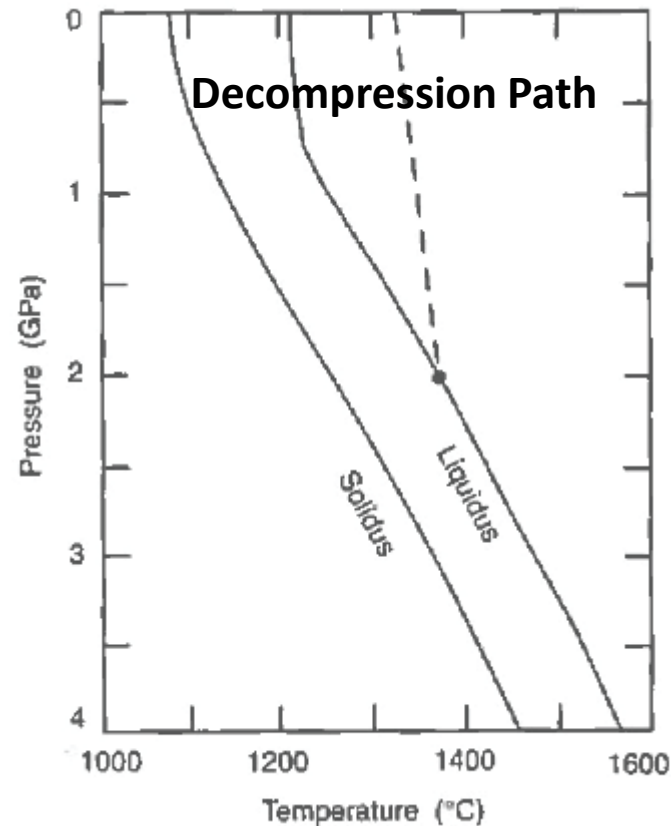
Time variation of T in the Makaopuhi lava lake, Kilauea volcano, Hawaii, at three different depths



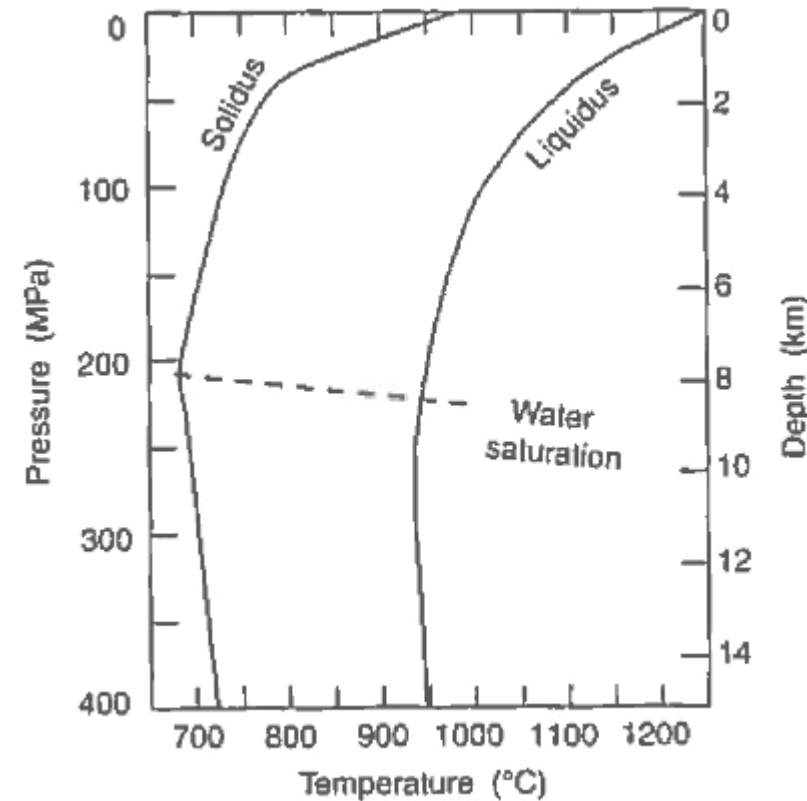
- In magma reservoirs, thermal convection does not modify temperature profiles greatly in comparison with those for purely conductive cooling.

Slopes of solidus-liquidus curves

Anhydrous tholeiite basalt



Mount St Helens dacite with 6% water



- At $P > 800\text{MPa}$ ($>25\text{ km}$) the slope of the liquidus is 3K km^{-1}
- At $P < 800\text{MPa}$ ($>25\text{ km}$) the slope of the liquidus is $1\text{-}2\text{ K km}^{-1}$ (in association with a change of mineral phase)
- Below the depth of saturation liquidus $T <$ for $P <$ (similar to dry magma)
- Above the depth of saturation liquidus $T >$ for $P <$ (amount of volatiles decrease)
- At large depth the magma is not water saturated and becomes superheated as it rises, as soon as it becomes water-saturated its superheat decreases.
- The final amount of superheat upon emplacement depends on the amount of heat lost to cold surrounding rocks during ascent and the initial water content (difference between erupted lavas and magmas intruding in the crust).

Magma cooling during ascent

Melt that rises in a dyke of length a and width d loses heat to the walls, so that:

$$\rho C_p w a d \frac{dT}{dz} \approx 2qa, \quad q = \lambda \Delta T / \delta \quad \frac{dT}{dz} \approx \frac{\kappa}{dw} \frac{\Delta T}{\delta}$$

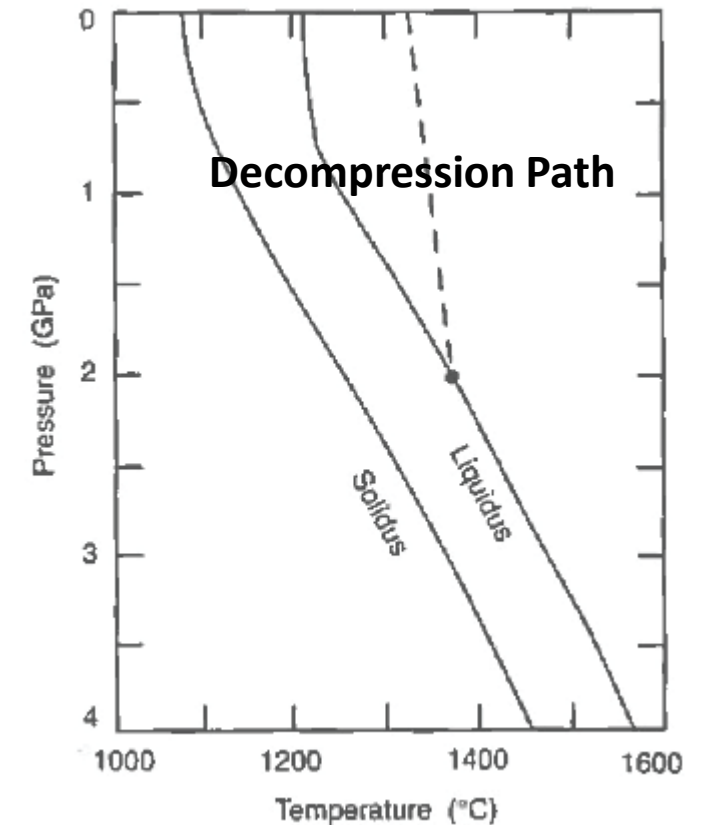
w =average magma velocity (e.g., 1 m s^{-1}), λ =thermal conductivity, ΔT =temperature difference between the wall and the magma, δ =width of the thermal boundary layer that develops, depending on the duration of the flow t . δ is about $2(\kappa t)^{1/2}$.

For $\kappa=10^{-6}\text{ m}^2\text{s}^{-1}$ and $t=1\text{ month}$ $\delta=1\text{ m}$

For $w=1\text{ ms}^{-1}$, $\Delta T=1000\text{ K}$, $d=2\text{ m}$, $dT/dz \sim 1\text{ K km}^{-1} < 1\text{--}3\text{ K km}^{-1}$ (liquidus slope)

- If the intrusion event last longer (several months or years), the amount of cooling is smaller (no formation of chilled margins of the dykes).
- Volatile undersaturated magma is likely to become and remain superheated during ascent, while hydrous saturated magma have lost their superheat and start to crystallize.

Anhydrous tholeiite basalt



Latent heat of fusion

- Latent heat of fusion: the amount of energy that must be supplied to melt a substance
- Transformation from one phase to another of a single chemical component is accompanied by the release or absorption of heat ($L=T\Delta S$), e.g., transformation (melting) of ice in water requires that heat has to be provided to the system (heat absorption), since the water has a higher density than that of the ice.

$$\frac{L}{T} = \Delta S = \Delta V \frac{dP}{dT}$$

ΔS =Entropy

*Latent heat of fusion for some pure silicate phases,
from Spera (2000)*

| Formula | T_L (K)† | L (kJ kg ⁻¹) |
|--|------------|----------------------------|
| Water | 273 | 333 |
| SiO ₂ (quartz) | 1700 | 157 |
| CaMgSi ₂ O ₆ | 1665 | 636 |
| Ca ₂ MgSi ₂ O ₇ | 1727 | 454 |
| Fe ₂ SiO ₄ | 1490 | 438 |
| Mg ₂ SiO ₄ | 2174 | 1010 |
| NaAlSi ₃ O ₈ | 1393 | 246 |
| Na ₂ Si ₂ O ₅ | 1147 | 196 |
| CaAl ₂ SiO ₈ | 1830 | 478 |

† Melting temperature.

Latent heat release

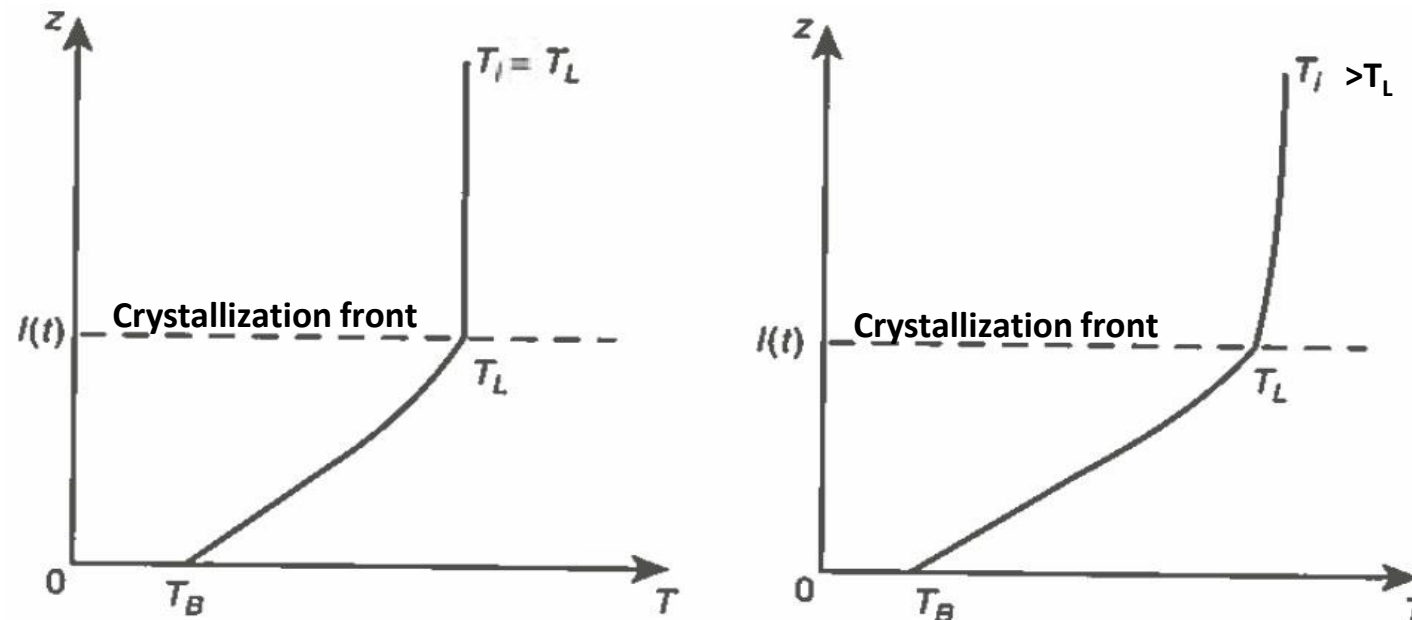
Importance of the latent heat release is provided by the Stefan number (St): it compares the amount of heat released per unit mass of liquid (L) to the amount of heat that must be extracted to cool material down at the boundary (T_B):

$$St = \frac{L}{C_p (T_L - T_B)}$$

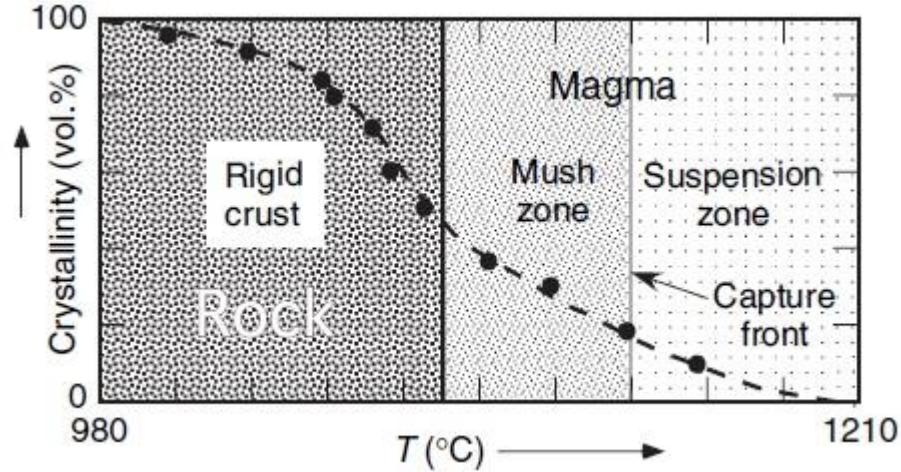
If $St \ll 1$, latent heat does not contribute significantly to the global thermal budget

For $L \sim 3 \times 10^5 \text{ J Kg}^{-1}$, $T_L - T_B \sim 800 \text{ K}$ and $C_p 10^3 \text{ J Kg}^{-1} \text{ K}^{-1}$, $St \sim 0.4$ (not negligible)

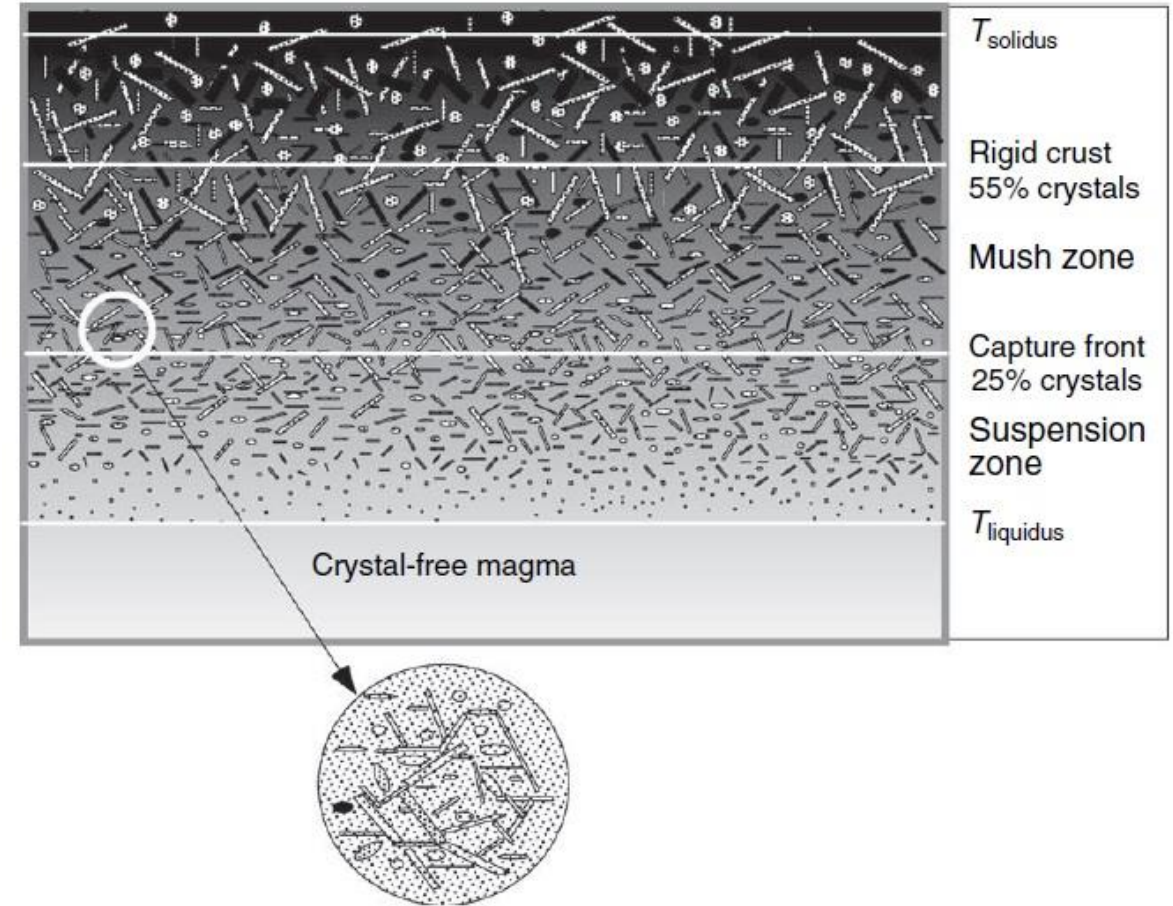
Cooling of a pure melt from a horizontal boundary



Solidification Front



- Near the liquidus, where crystallinity is low (<25 vol.%) and crystals are small, the magma is a 'suspension'.
- In the 'mush' zone the crystals are still separated from one another. This zone persists until the crystallinity reaches about 55 vol.%. .
- When crystallinity > 55 vol.% tacked together to form a solid framework of some strength. Thus, the magma becomes a rock (dilatant solid) and can no longer flow as a viscous fluid. The remaining melt, (interstitial to the crystals) can still move, but as a porous or Darcian flow.
- With continued cooling and crystallization in the magma below, gases exsolve, increasing pressure.



- A solidification front is an active zone of crystallization that forms the perimeter of all magmas.
- Solidification fronts are dynamic, since they continually move in response to the prevailing thermal regime.

Kinetic controls on crystallization

- When a magma basalt is emplaced at a depth of about 10 km at a T of $\sim 1200^\circ\text{C}$, coming in contact with the surrounding rocks at $T \sim 200^\circ\text{C}$, forms a chill zone (sometimes with glass formation) along the margins (initially at $T \sim 700^\circ\text{C}$), while crystallization occurs mainly in the inner part during undercooling.
- At liquidus, the nucleation and growth rates of crystals are zero, then they increase with increasing undercooling until they reach a maximum and then decrease to zero at large undercooling.
- Crystallization efficiency is sensitive to pre-existing nucleation sites (e.g. presence of xenocrystals) and presence of water dissolved in the melt: in hydrous basalts the amount of solid phases formed is smaller than in the dry ones, at the same cooling rate.

$$\tau_K = (IY^3)^{-1/4}$$

τ_K = Kinetic crystallization time scale, $\tau_{\max} = 3 \times 10^5 \text{ s}$

I = nucleation rates, $I_{\max} = 1 \text{ cm}^{-3} \text{ s}^{-1}$

Y = growth rates, $Y_{\max} = 10^{-7} \text{ cm s}^{-1}$

For olivine: $Y = 10^{-11} - 10^{-10} \text{ cm s}^{-1}$ and $I = 10^{-6} - 10^{-5} \text{ cm}^{-3} \text{ s}^{-1}$, $\tau_K \sim 4 \times 10^8 \text{ s}$

For plagioclase: $\tau_K \sim 10^7 \text{ s} - 10^8 \text{ s}$

If chilling occurs, cooling is faster than crystallization and conduction is the dominant heat transport and cooling time (τ_{diff}) is:

$$\tau_{\text{diff}} \approx \frac{1}{4} \frac{h^2}{\kappa}$$

h = thickness of sills/dykes

Avrami Number (Av): ratio of the cooling time to the kinetic time: $Av = \left(\frac{\tau_{\text{diff}}}{\tau_K} \right)^4$

For $Av < 1$ magma gets chilled

For an intrusion of basalt magma of 1 m $\tau_{\text{diff}} = 3 \times 10^5 \text{ s}$, $Av < 1$

- Crystals are distinguished in **phenocrysts**, representing crystallization at depth under slow, protracted cooling, and **groundmass**, representing the latest rapid phase of cooling.

Crystals escape

- When a crystal becomes deeply embedded in a solidification front, melt viscosity and hindrance from neighboring crystals increase to the point that escape is not possible.
- The region where this transition occurs marks the boundary between the suspension and mush zones and is called the 'capture front'.
- Crystals outward of the capture front are trapped in the solidification front.

In the leading part of the solidification front, the suspension zone, crystals can settle easily with a rate V_s (Stokes' law):

$$V_s = C_s \frac{\Delta \rho g a^2}{\mu}$$

C_s = constant (2/9) depending on the shape of the crystal

$\Delta \rho$ = density contrast,

g = gravity,

a = crystal radius

μ = viscosity

The position of the front, $S(t)$, is: $S(t) = 2b(Kt)^{1/2}$ The rate of advance (V_F) is: $V_F = \frac{dS(t)}{dt} = b \left[\frac{K}{t} \right]^{1/2}$

K = thermal diffusivity

b = constant that measures the effect of latent heat relative to the enthalpy of the magma $b = \left[\frac{C_P(T_m - T_w)}{5H(\pi)^{1/2}} \right]^{2/5}$

C_P is specific heat, T_m is magma initial temperature, T_w is wall rock initial temperature, and H is latent heat

Equating V_F with V_s we get the minimum size crystal that can escape from the solidification front as a function of time:

$$a(t) = \left[\frac{b \mu (K/t)^{1/2}}{C_s \Delta \rho g} \right]^{1/2}$$

Crystals escape

The measure of the crystal size must be compared with the actual size crystal growing in the suspension, which can be found by estimating how long is taken by the capture front isotherm to traverse the suspension zone as a function of time:

The position of the capture front isotherm is a large fraction (f_{CF} 0.95–0.98) of position of the leading edge of the solidification front:

$$S(t) = 2bf_{CF}(Kt)^{1/2}$$

The difference with the previous equation: $S(t) = 2b(Kt)^{1/2}$ gives the thickness of the suspension zone:

$$\Delta S(t) = (1-f_{CF})2b(Kt)^{1/2} \quad \text{Time spent traversing the suspension zone: } \Delta t(t) = \frac{\Delta S(t)}{V_F(t)} = 2(1-f_{CF})t$$

$$V_F = \frac{dS(t)}{dt} = b \left[\frac{K}{t} \right]^{1/2}$$

$$\text{General growth law for crystal growth: } a(t) = G \Delta t(t) = 2G(1-f_{CF})t \quad a(t) = \left[\frac{b}{C_s} \frac{\mu(K/t)^{1/2}}{\Delta \rho g} \right]^{1/2}$$

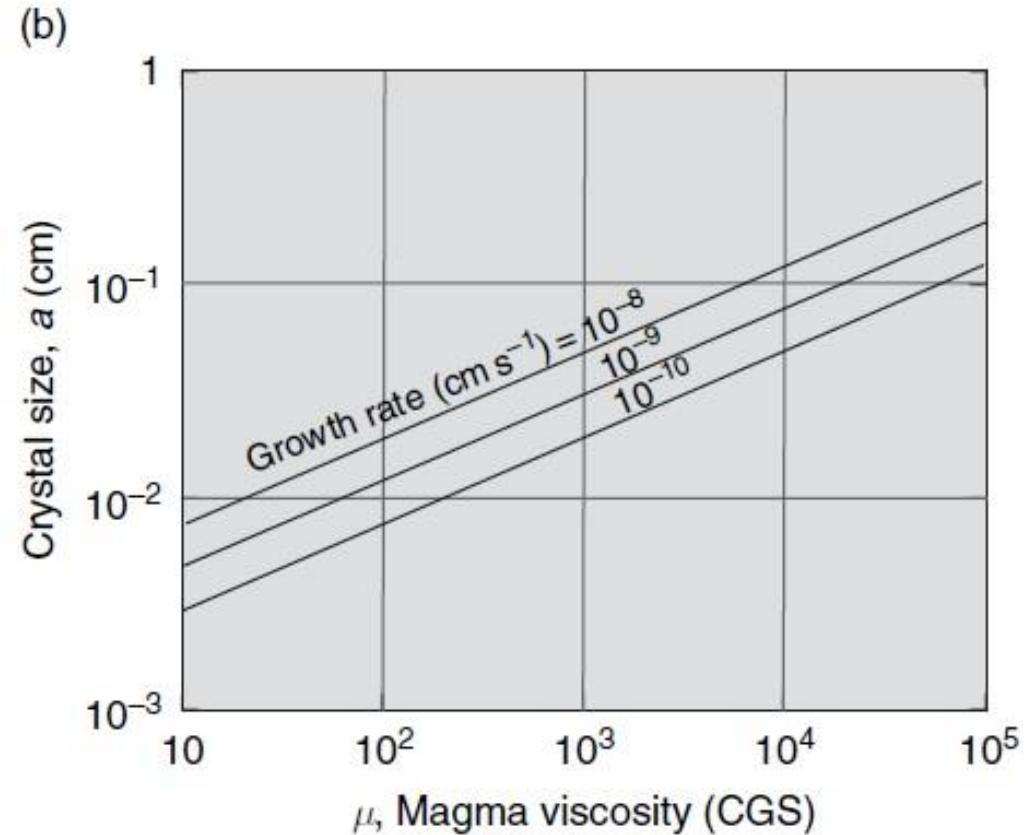
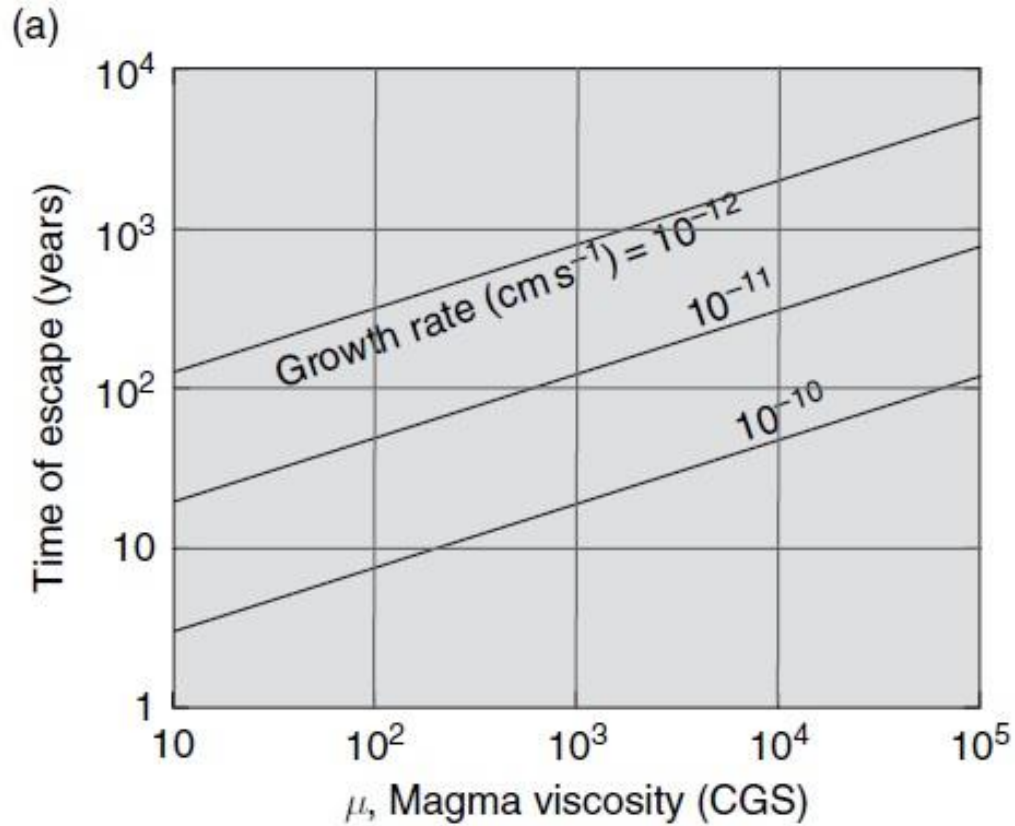
From these two equations we obtain:

$$\text{Time at which the first crystals escape from the solidification front: } t = \left[\frac{(b/C_s)\mu K^{1/2}}{(2G(1-f_{CF}))^2 \Delta \rho g} \right]^{2/5} \quad G = \text{growth rate (cms}^{-1}\text{)}$$

Size of the crystal at the time of escape:

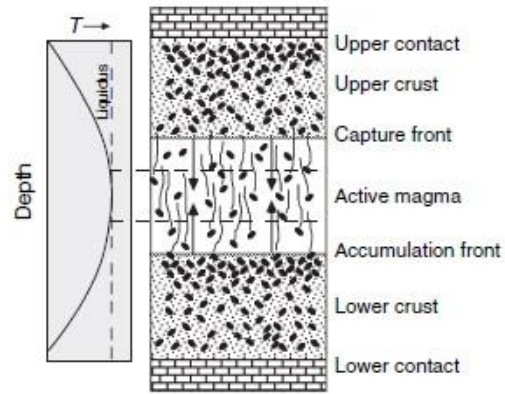
$$a(t) = \left[\frac{(b/C_s)(2G(1-f_{CF}))^{1/2} \mu K^{1/2}}{\Delta \rho g} \right]^{2/5}$$

Crystals escape



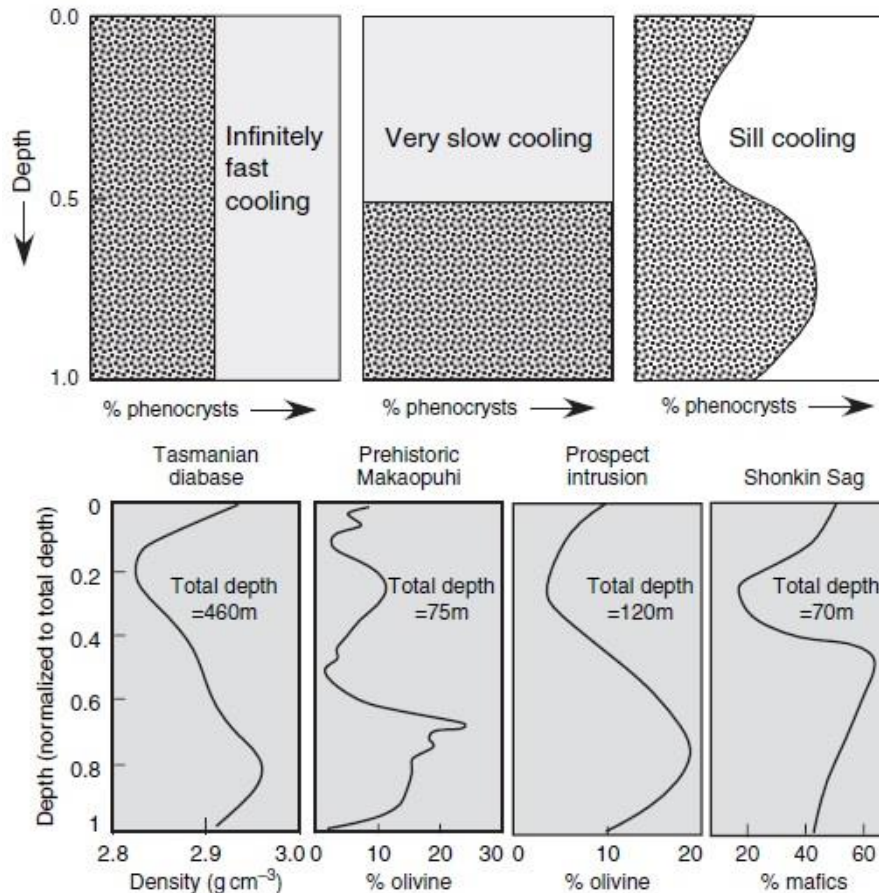
- The sooner a crystal can reach the critical size to sink faster than the rate of advancement of the capture front, the sooner it can escape (the time of escape is very sensitive to the growth rate).
- As magma viscosity increases, the time of first escape and the crystal size at escape increase.
- Crystals escaping the crystallization front enter the hottest part of the system and will begin melting and dissolving back into the magma.

Crystals distribution



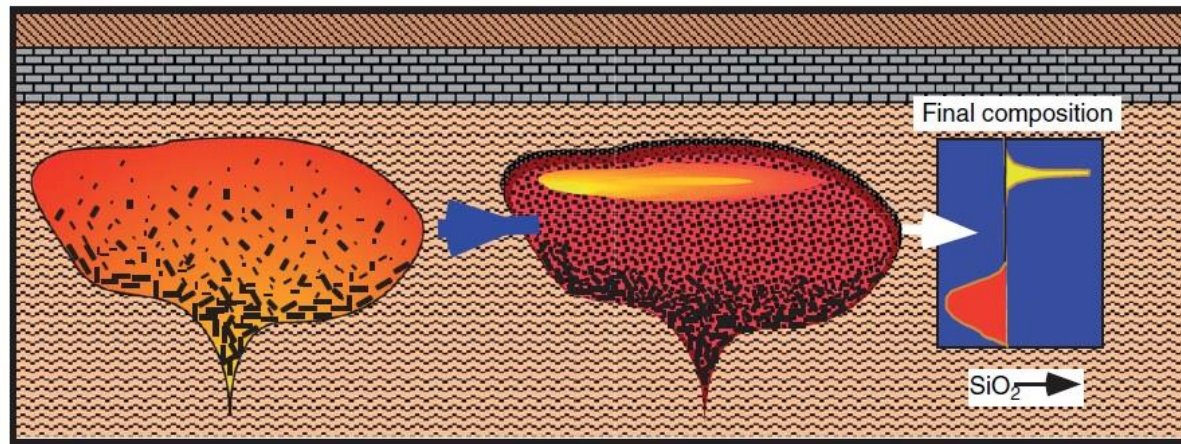
When magma carrying phenocrysts is emplaced, there is a competition between the rate of crystal settling and the rate of solidification

- If the magma could cool instantaneously all the crystals would be trapped in place before they could settle any distance at all (uniform crystals distribution from top to bottom throughout the sheet).
- If the body did not crystallize at all, all the crystals would settle to the floor and form a thick pile.
- In the most common cases, inward advancing solidification fronts from the top and bottom compete with crystal settling.
- Since the rate of solidification front advance decreases inversely with distance into the body, more and more crystals escape capture with depth and all crystals initially in the center of the body settle into the lower rising solidification front (S-shaped profile of phenocrysts abundance vertically through the body).

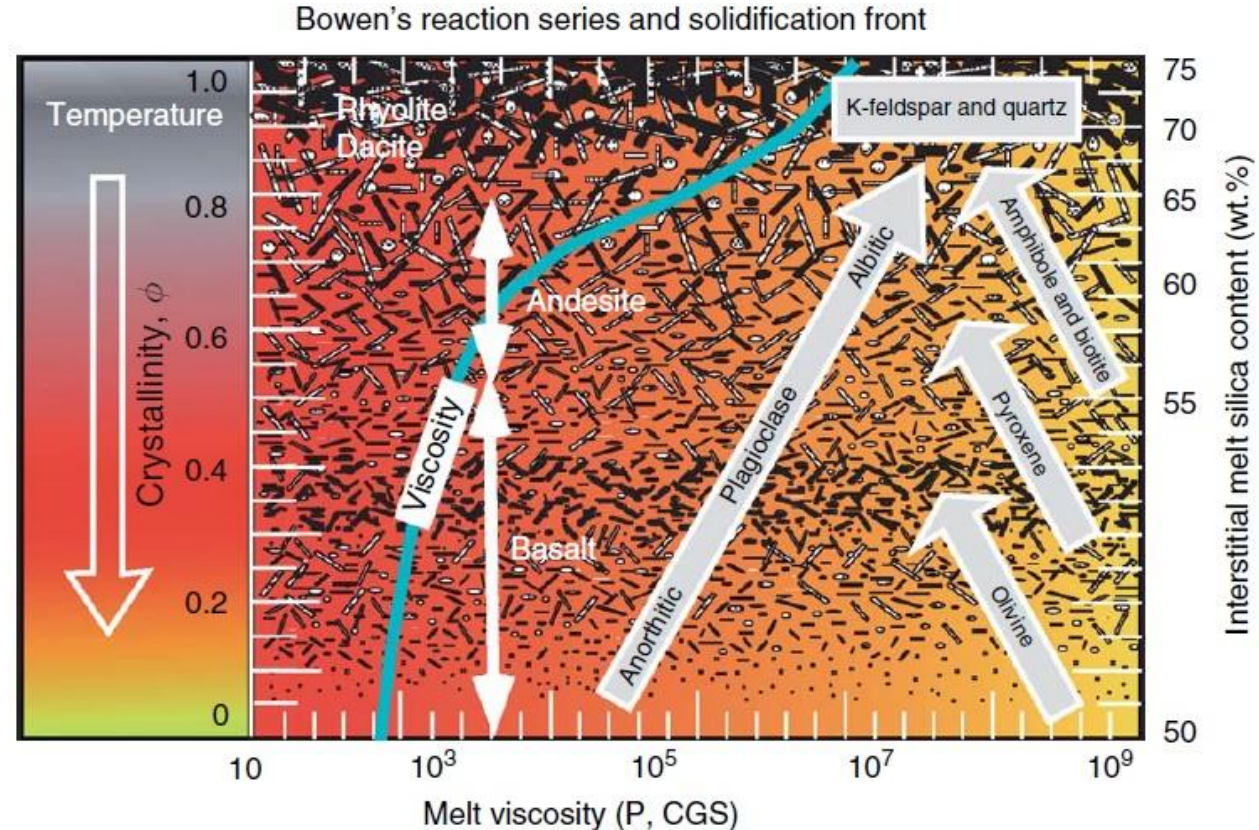


Dependence of magma composition and viscosity on crystallization

Separation of crystals from melt causes chemical differentiation



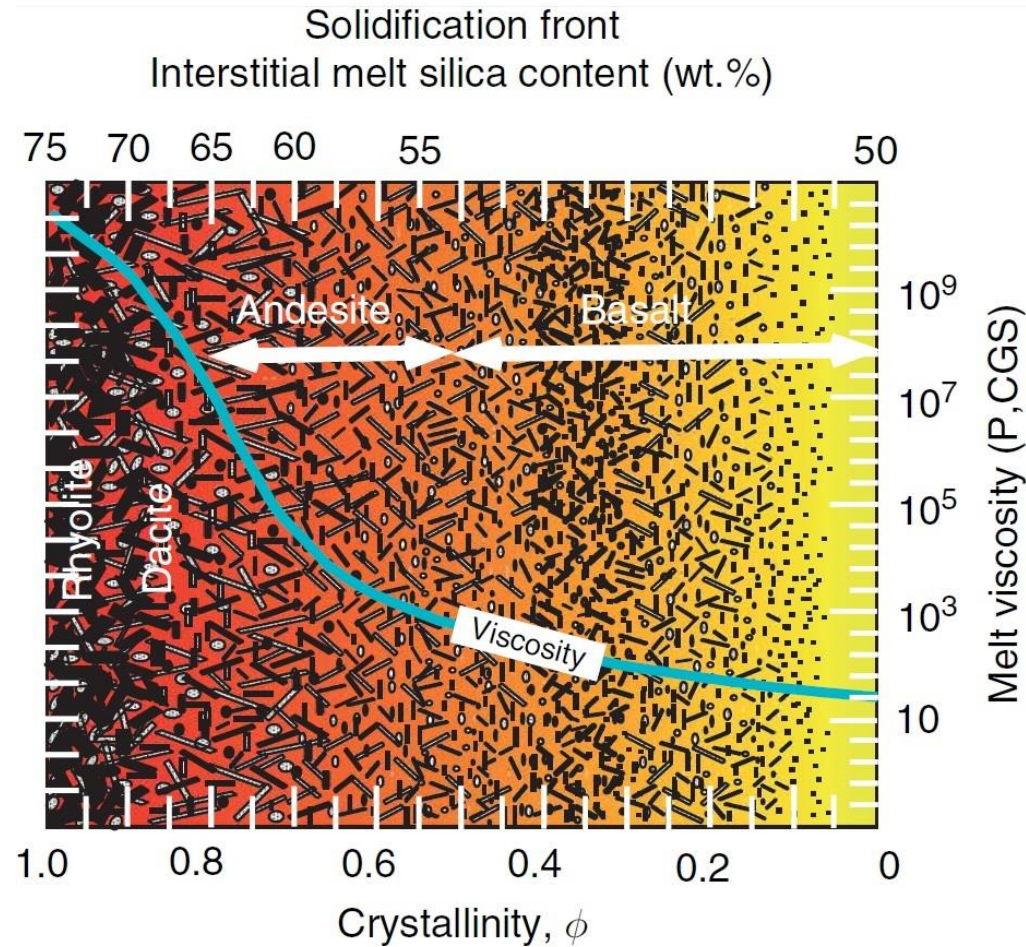
Classical magma chamber



- Crystals already present in the magma favour crystallization processes: continual growth of crystals will enhance nucleation
- Once the cooling front arrives, existing nuclei increase in size outward through the solidification front.
- The small crystals combine through aggregation and grain boundary migration into a large, optically continuous crystals (touching grain combine in response to lowering the net free energy).

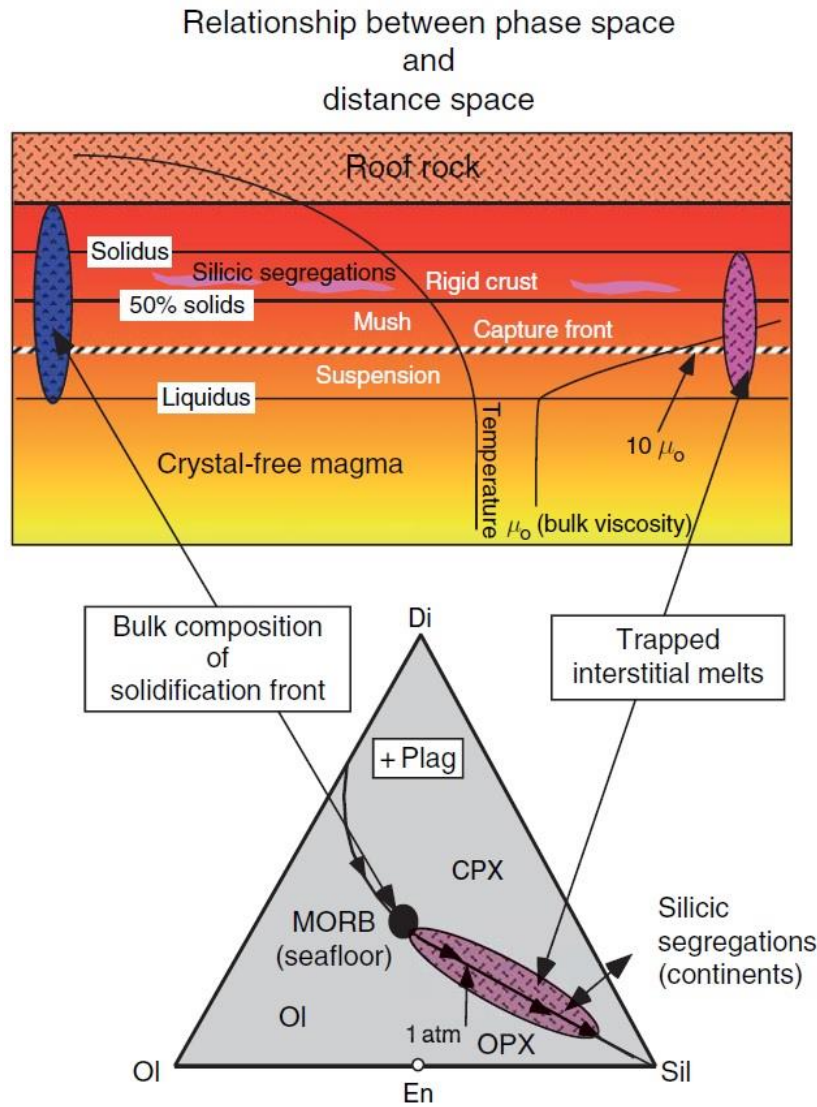
Dependence of magma composition and viscosity on crystallization

Variations in melt silica content and viscosity with crystallinity

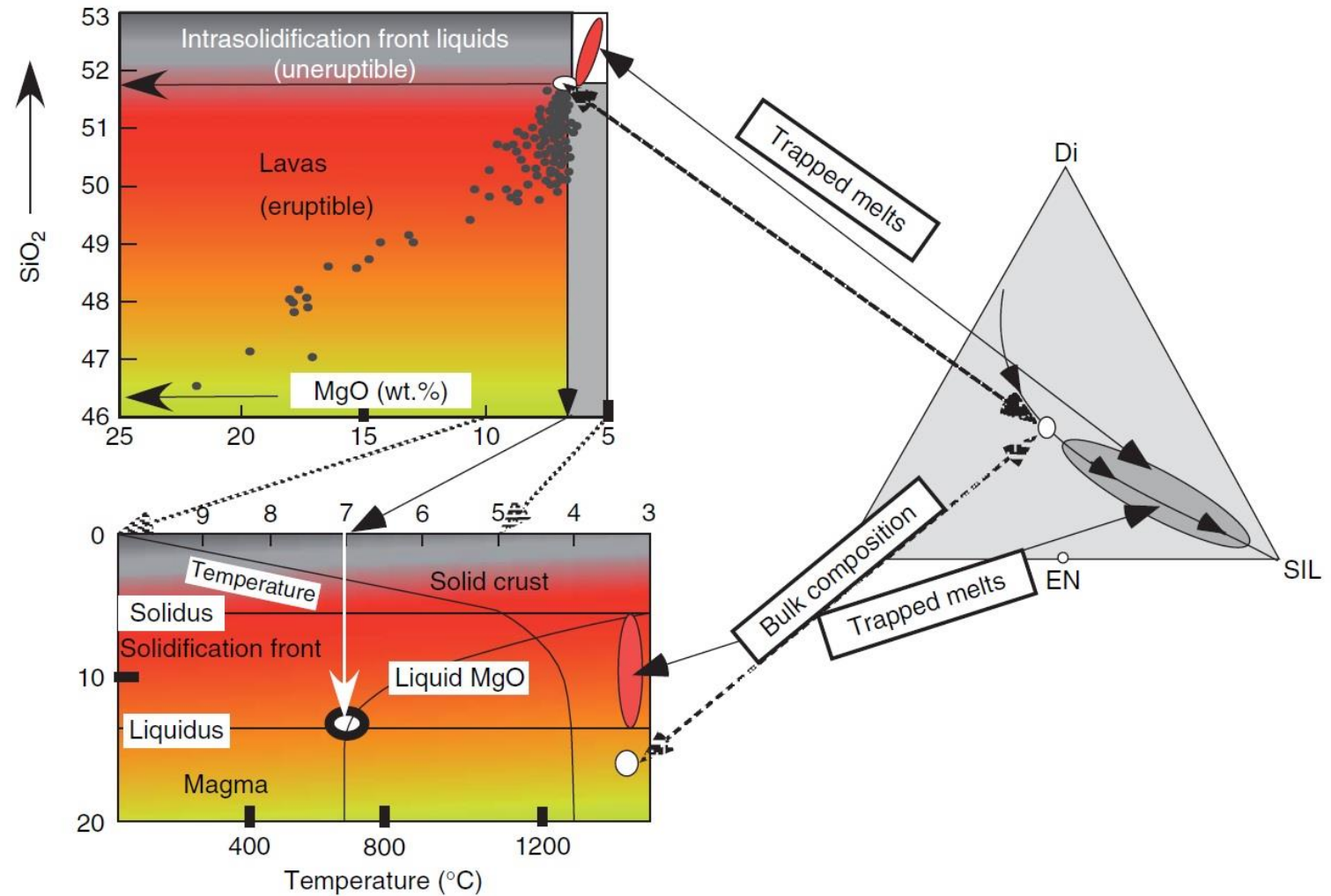


- Silica enrichment increases gradually for the first 50 vol.% of crystallization (the increase is only ~5 wt.%).
- Andesitic, dacitic, and rhyolitic melts occur at increasingly higher states of crystallinity, where the separation of melt and solids is increasingly difficult.

Dependence of magma composition on crystallization

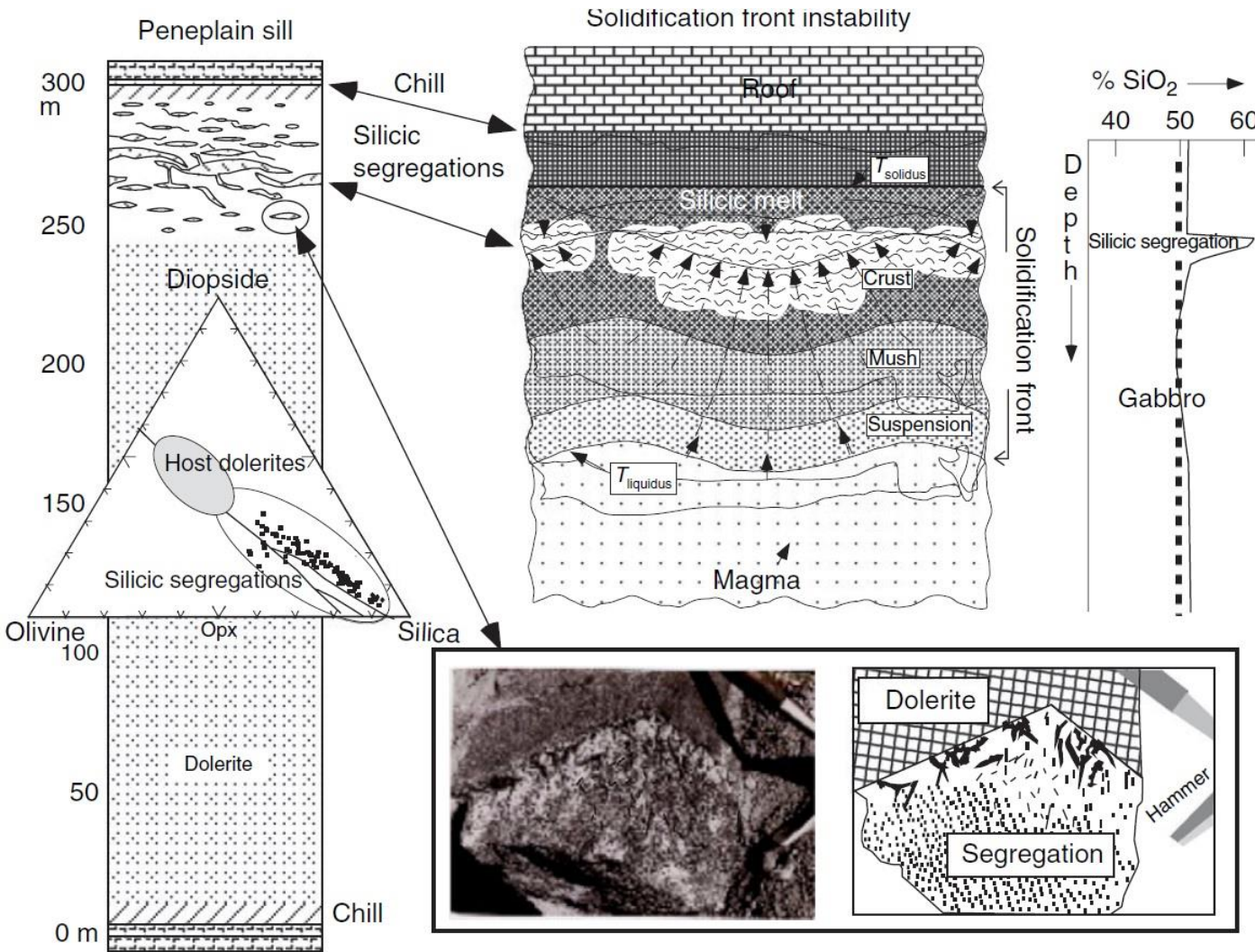


Lavas from Kilauea



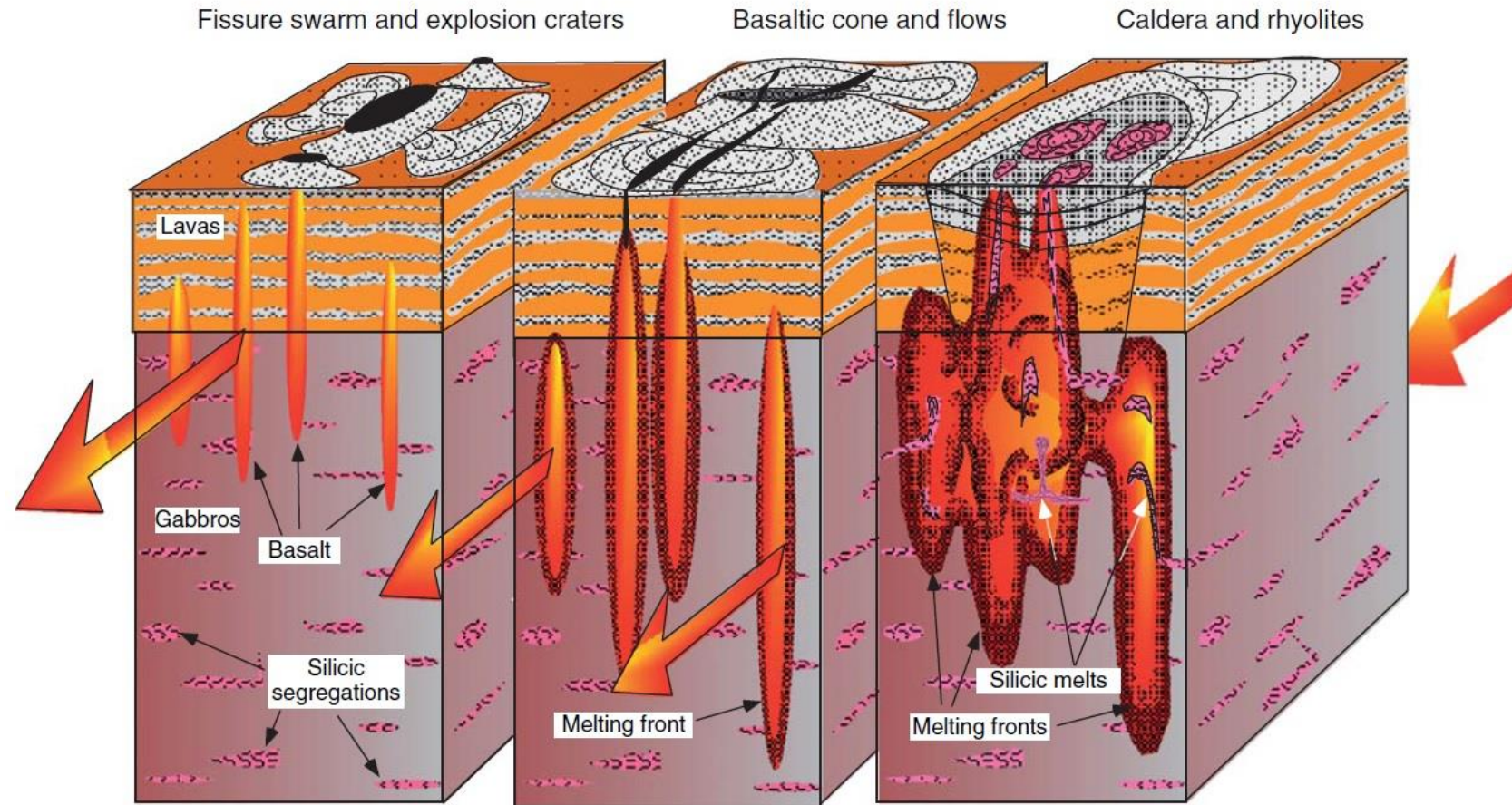
- No magma comes from within the solidification fronts (the lavas are generally always just basalts).
- Bimodal differentiation comes from solidification fronts that tear open and form lenses of silicic liquids or silicic segregations.

Differentiation Processes: Solidification front instability (SFI)



- The lenses of silicic materials (pegmatitic segregations, 60–65 wt.% SiO₂) begin thin and small near the sill roof and increase in thickness and length (usually are between 1m and 20–30m).
- A likely mechanism of formation of these lens, having a coarse texture, is by internal failure (e.g., roof failure) or tearing of the solidification front (where there is enough strength to fracture).
- With initiation of tearing, local interstitial melt is drawn into the tears. The melt arriving from lower in the solidification front is slightly hotter and as it cools it forms progressively smaller crystals downward in the segregation.
- To avoid tearing, the local strength of the solidification front must increase proportionally with weight, and strength comes from increasing crystallinity.

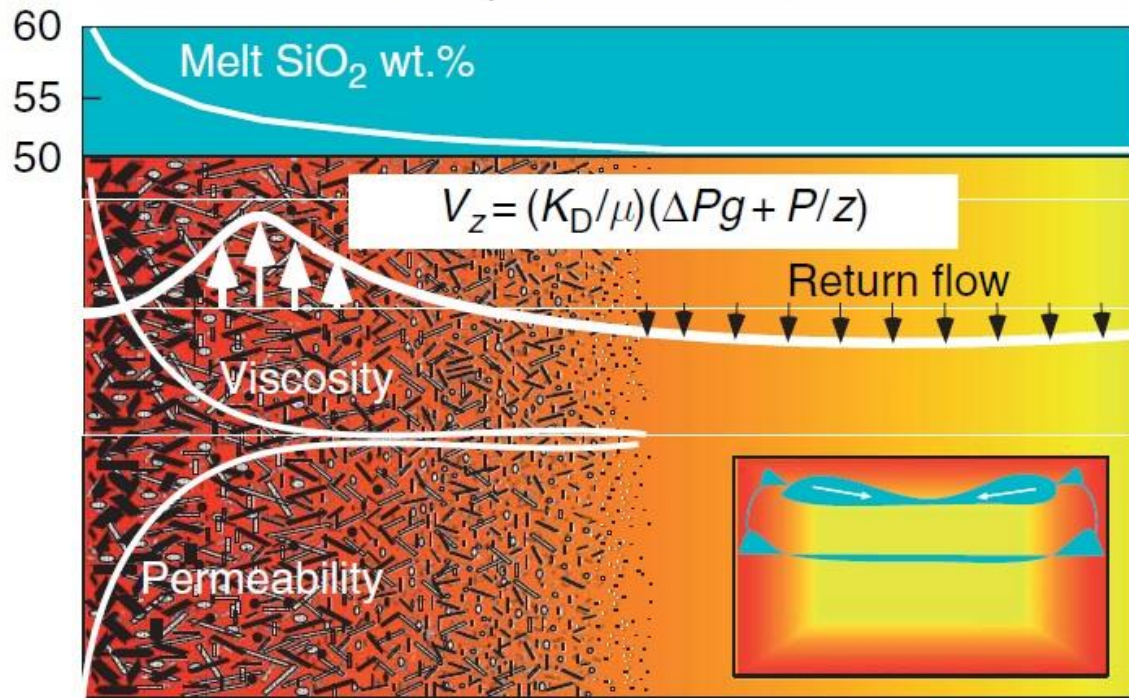
Bimodal suite of composition (basaltic and silicic endmember): Iceland volcanic rocks



- About 18% of the surface rocks of Iceland are highly silicic rhyolites.
- Fissures propagating into older crust bring hot basaltic magma into thin warm crust, which promote progressive wholesale melting via melting fronts propagating outward into the country rock.

Sidewall Uflow

Sidewall upflow of interstitial melt



$$V_m = \frac{K_D}{\mu} \left[\frac{\partial P}{\partial Z} + \Delta \rho g \right]$$

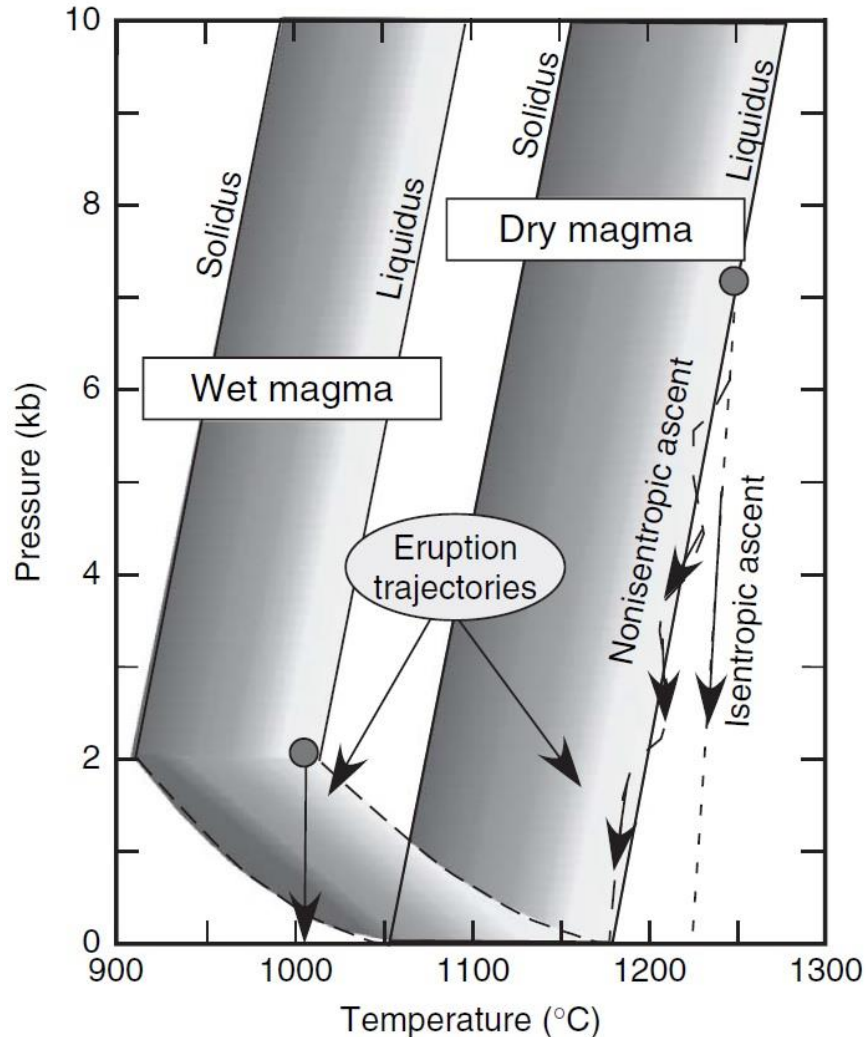
- Silicic melts can escape if the advance of the solidification front is slow enough (cooling slows with increasing depth) to allow upward flow through the porous solidification front (Darcian dynamics).

K_D/μ decreases strongly approaching the solidus T .

- There is an optimum position within the rear of the solidification front where melt is most efficient at flowing upward.

V_m = Darcy velocity, (it measures the melt flux per unit area), K_D = permeability, μ = melt viscosity, P is pressure, z = spatial coordinate, $\Delta\rho$ = density contrast, and g = gravity.

Superheat Conditions



- Thermal convection is not a common process in magma chambers.
- For a magma with a very small amount of superheat ($\sim 1^\circ\text{C}$), Ra could be large if L is sufficiently large.
- Any amount of superheat will bring on vigorous convection and rapidly bring the temperature back to the liquidus (dissipation of superheat to the wall rock).
- With the loss of superheat, thermal convection ceases, and the process repeats itself on a timescale governed by the rate of ascent.

The time (t) for convection in each layer to start and become fully established is:

$$t = \frac{L^2}{\kappa} \left[\frac{B}{Ra} \right]^{2/3}$$

$$B = 350$$

L = thickness of an individual layer

Modes of Magmatic Cooling: Superheated Magma

Rapid convective cooling The average temperature (T) in a rapidly convecting layer, strongly cooled from its margins, is:

$$T = T_L + \Delta T_o \left[1 + \frac{t}{\tau} \right]^{-1/b}$$

$$\tau = \frac{L^2}{2Cb\kappa'} Ra^{-b}$$

T_L = liquidus temperature, ΔT_o = amount of superheat, t = time

$$Nu = (C Ra^b) \quad b=1/3 \quad C=0.4 \quad (\kappa' = K_c / \rho C_p)$$

The thermal conductivity (K_c) is of the wall rock and the density (ρ) and specific heat (C_p) are of the magma.

Slower convective cooling in conductive medium

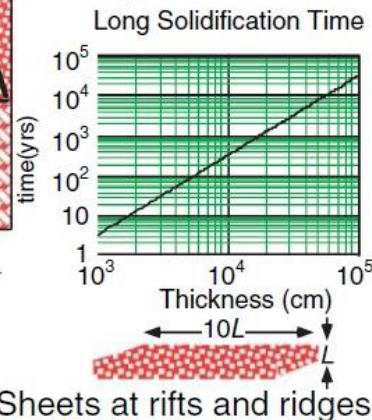
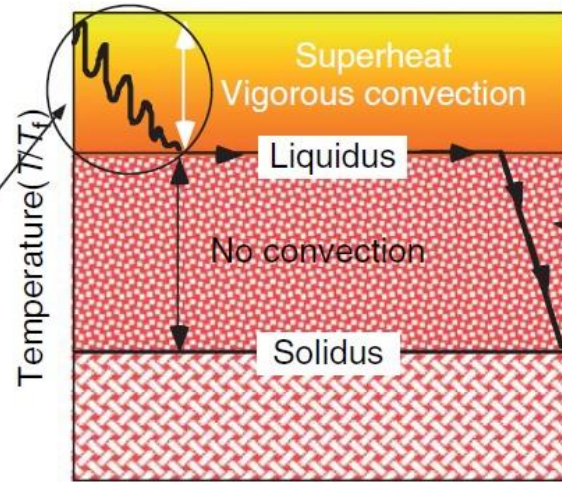
$$\frac{T - T_w}{T_m - T_w} = \exp(bx + b^2 \kappa t) \operatorname{erfc} \left(\frac{x}{\sqrt{4\kappa t}} + b\sqrt{\kappa t} \right)$$

$x=0$

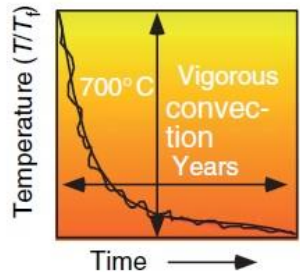
$$\frac{T - T_w}{T_m - T_w} = \exp(4F) \operatorname{erfc}(\sqrt{4F}) \quad F = \kappa t / L^2$$

$$h = 2/L \quad \text{for} \quad M = \rho' L / 2$$

Superheat (no crystals) vs no superheat (crystals)



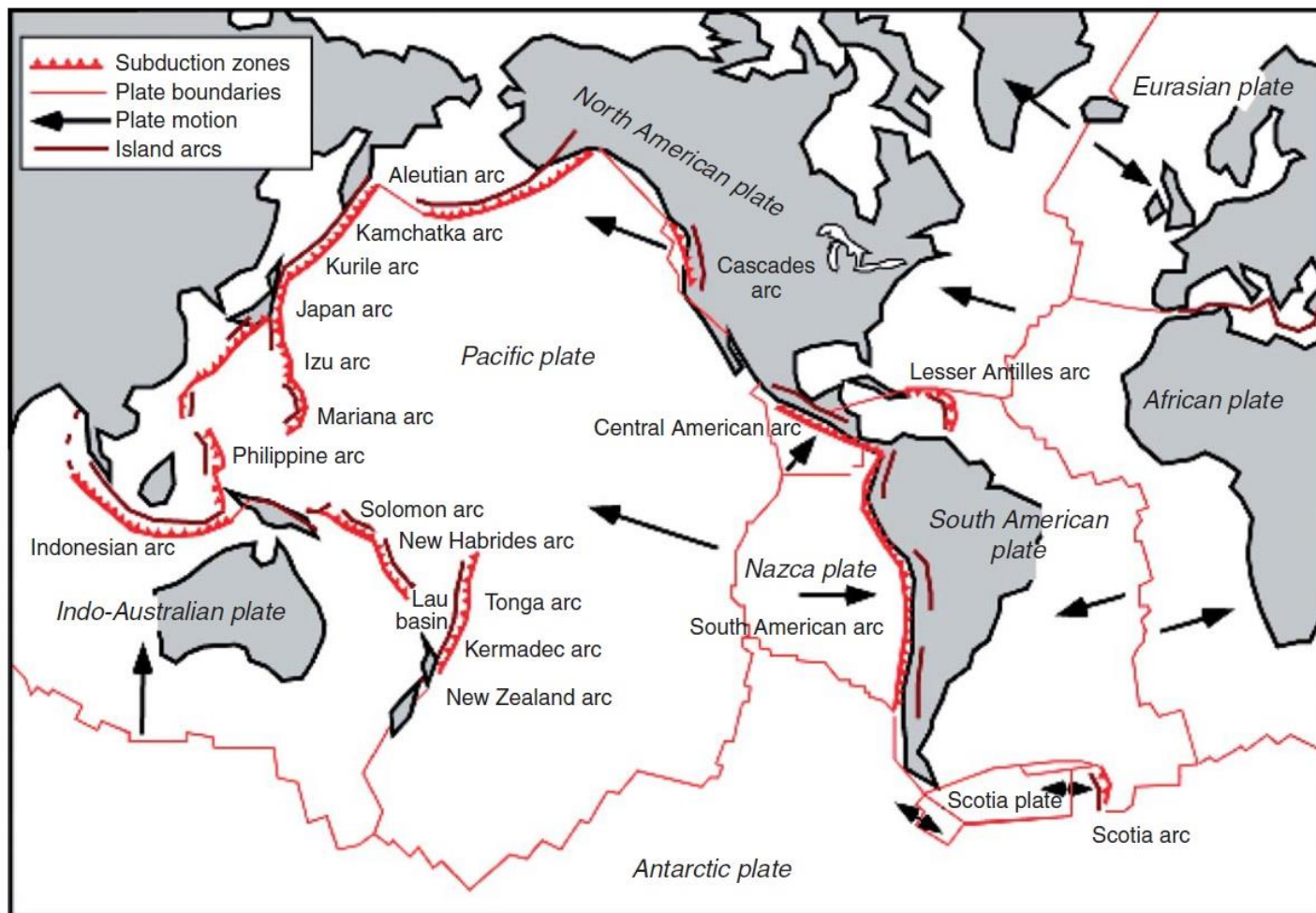
Rapid loss of superheat
No crystallization



Sudbury impact melt sheet

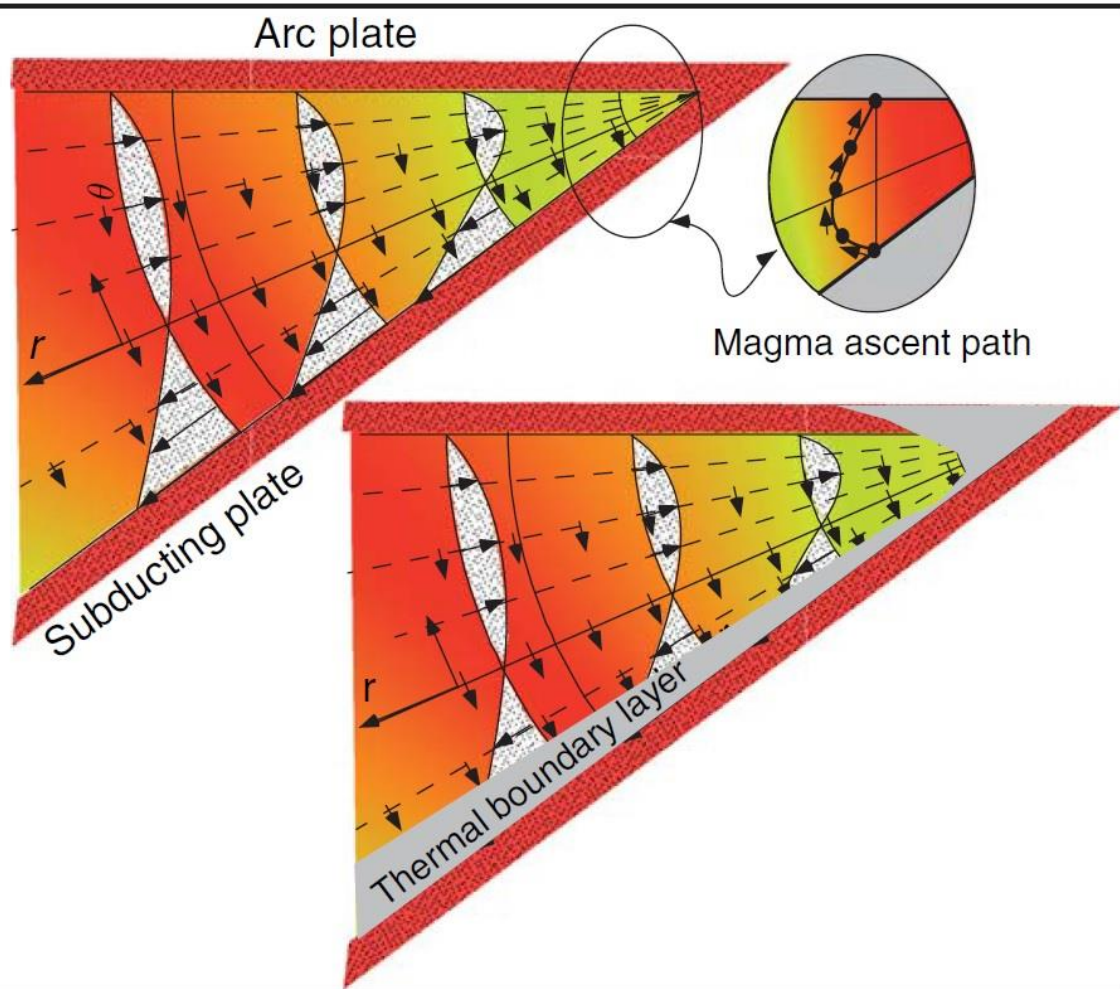
T_m = the initial magma temperature, T_w = the initial wall rock temperature, x = the vertical spatial coordinate, and $h = \rho C_p / M C_p'$, where ρ is wall rock density, C_p and C_p' = the specific heats of the wall rock and magma, and M = mass of magma (of density ρ') in contact with a unit area of wall rock.

Arc Volcanism



- The volcanic centers of the island arcs can remain fixed for long periods of time (1–10 My).
- In arcs a wedge of asthenosphere, sandwiched between the arc lithosphere and the subducting oceanic crust, provides almost any number of end members and combinations from which to get magma.
- The ages of the arcs also vary a great deal from those like the Aleutians that may be as old as 40Myr to those as young as the Scotia arc at 3 Myr.
- The detailed structure of the volcanic front of any arc often intimately reflects the structural morphology of the subducting plate (e.g., angle of subduction).
- Arc volcanic fronts are not simple continuous arcs, but are usually segmented into a collection of individual short segments forming a piecewise continuous arc.
- The dominant type of magma emitted depends on the density of the underlying crust: Arcs on continental terrains are dominantly andesitic (i.e., 60 wt.% SiO_2), whereas arcs on oceanic crust are dominated by high-alumina Basalts (HAB, 50 wt.% SiO_2).

Flows in the mantle wedge of the subduction zones

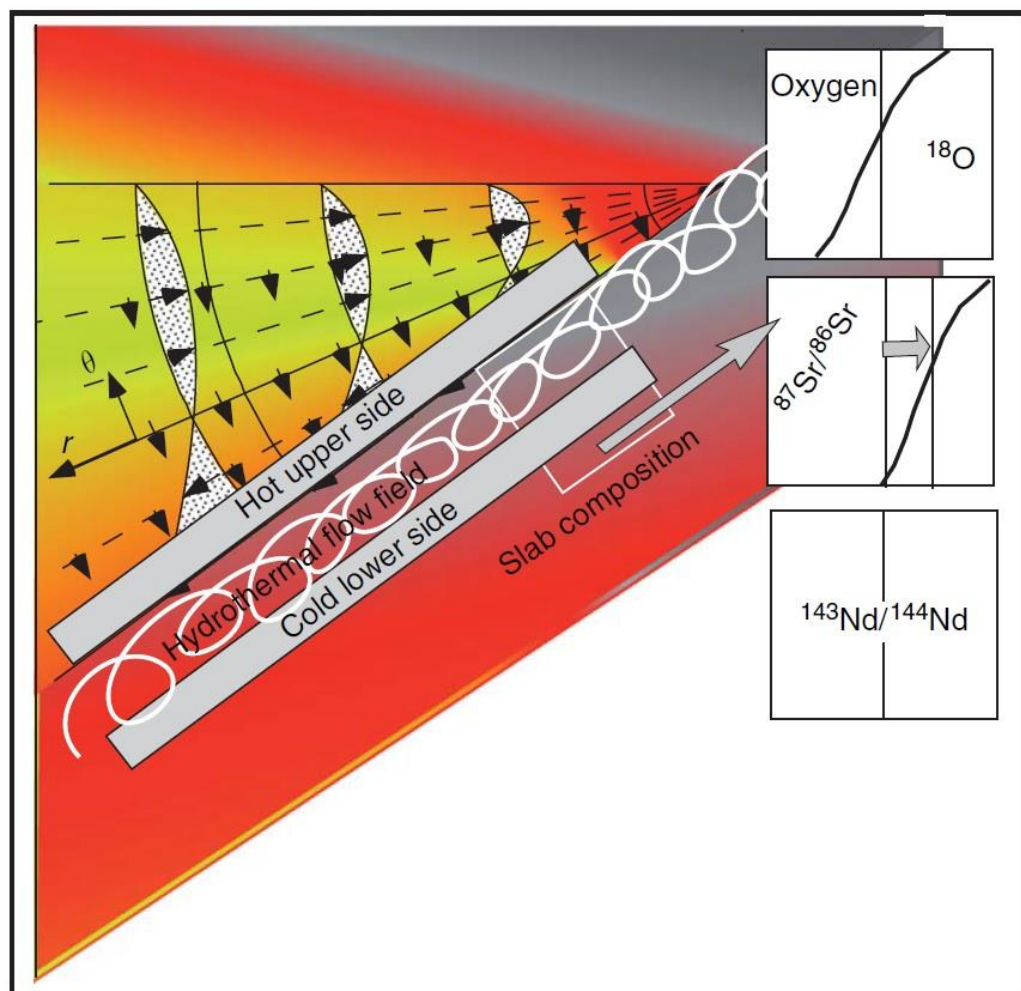


- At shallow depths, the subducting lithosphere is in fault contact with the adjacent arc lithosphere. The associated stress field creates a coupling between the two plates (welded together along the lithosphere-to-lithosphere fault contact).
- The depth extent of this contact is 70–100 km or longer in continental terrains, where the lithosphere is older and thicker.
- Below this depth of fault contact, the subducting plate is in contact with the mantle wedge, which adheres to the upper boundary with the arc lithosphere and at the same time is dragged downward by the motion of subduction.
- The flow in the mantle wedge has an upper part, which is essentially a parabolic flow between parallel plates (directed normal to the face of the subducting plate) and the lower half is a shear flow (streamlines parallel to the plate).
- Then, the corner flow continually brings into the wedge region hotter mantle material from greater depths, and this material is continually brought in contact with the subducting plate.
- Consequently, a cold thermal boundary layer develops near the corner and thickens down the plate of thickness δ_T .

$$\delta_T = C\sqrt{K}\left[\frac{x}{V}\right]^{1/2}$$

x = distance from the corner, which is measured by $x=t/V$, t = time, V = subduction velocity, K = thermal diffusivity, C = constant

Hydrothermal flow and magma formation in the subduction zones

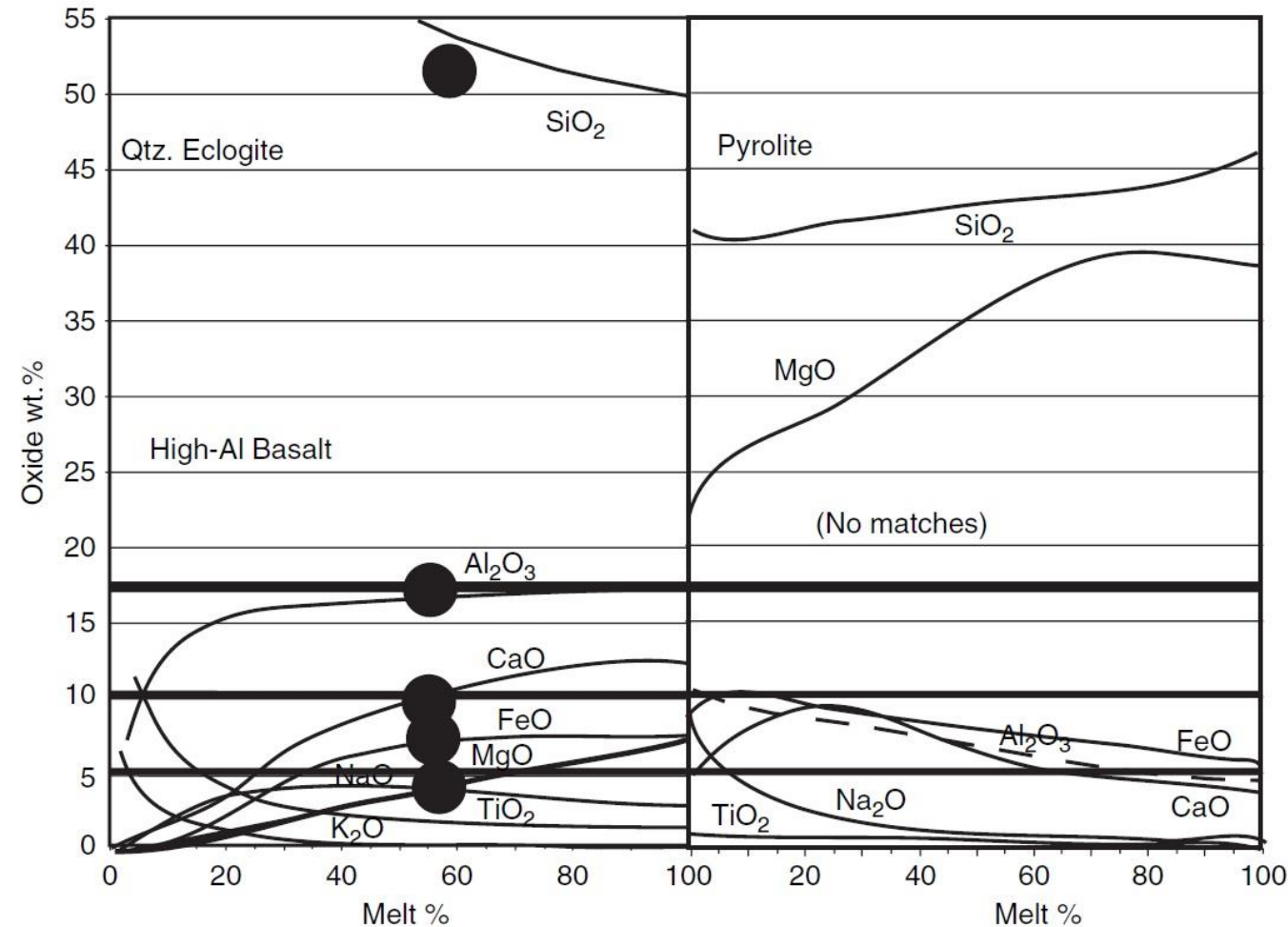


- The production of magma are the thermal regime and the transport of hydrothermal fluids.
- For seafloor rocks at 0°C brought up against mantle wedge at 1300°C , the crust interface of the subducting plate would achieve a T of about 650°C .
- However, we should consider that the contact between the two slabs extends to some depths (e.g., 125 km) and that frictional interaction produces heat, so that the oceanic crust may be brought to a temperature of about 1350°C and starts to melt.
- As the plate subducts and becomes heated, a new hydrothermal circulation system sets up due to phase transformations associated with dehydration. The form of this flow is horizontal and upward (helical).
- This flow is confined only to parts of the plate where there is sufficient permeability to allow flow (rocks are sealed beyond T of $\sim 700^{\circ}\text{C}$).
- This flow has a critical effect on the chemical nature of the subducting plate, increasing the amount of ^{18}O and the ratio $^{87}\text{Sr}/^{86}\text{Sr}$.

Magma composition of the volcanic arcs

Melt composition as a function of degree of melting at 3–4 GPa produced from:

a) quartz-eclogite subducted oceanic crust and b) pyrolite of the asthenospheric wedge.



- The composition of the predominant basalt found in island arcs has an exceedingly close chemical affinity to subducted MORB.
- No matches are found for melt from the peridotitic wedge material, but matches are almost complete for each of the major elements for slab melting.
- Arc volcanic HABs have a low MgO content (7.5wt.%) and elevated Al₂O₃ content (16–20 wt.%).

At P of 2–3GPa, the MORB mineral assemblage of plagioclase, olivine, clinopyroxene transforms to quartz, garnet, jadeitic pyroxene (a quartz-eclogite rock).

The straight horizontal lines show the composition of typical HAB of island arcs (Aleutians) and the large dots show matching points between the observed HAB and the calculated melts.

Volcanic eruptions

- Explosive behaviour of highly viscous magma is due to the fact that different gases in supercritical conditions can be trapped and delivered along with the magma to the surface.
- The most active volcanoes (> 450) and the most powerful eruptions (> 85%) are located within the Pacific Ring of Fire.
- The most powerful eruptions result in caldera formation, whose products are represented by rhyolites and dacites, which have much lower melting point (200-400 K) and higher viscosity than basaltic magma (which often do not reach the surface).



Yellowstone Caldera



- The **Yellowstone caldera**, about 80 km long and 50 km wide, is the youngest of three nested and overlapping calderas, and is filled with younger rhyolitic lavas. The calderas were formed during three episodes of magmatic eruptions about 2 million years ago ($\sim 2,500 \text{ km}^3$ Huckleberry Ridge Tuff), 1.3 million years ago ($\sim 280 \text{ km}^3$ Mesa Falls Tuff), and 0.6 million years ago ($\sim 1,000 \text{ km}^3$ Lava Creek Tuff).
- The formation of hydrothermal features around the Yellowstone Lake area is related to convective meteoric hydrothermal fluid circulation above a cooling magma chamber.
- The total fluid discharge from all thermal features of Yellowstone National Park averages an astounding $\sim 3,091 \text{ L/s}$ and temperature $\sim 345 \text{ K}$.
- Yellowstone National Park has the largest number of geysers in the world (> 400)

Kamchatka Peninsula

- The Kamchatka Peninsula is located in the Pacific Ring of Fire and is one of the most volcanically active regions on Earth. There are about 300 volcanoes and about 30 are still active.
- The largest Holocene eruption in the Kuril–Kamchatka resulted in an estimated tephra volume of 140–170 km³ with the formation of a caldera.



Kamchatka Peninsula

Kamchatka has a great geothermal potential, characterized by thermal waters (used for heating) and geysers (more than 200 in the Valley of Geysers).

Mutnovsky Volcano

| Site, feature | Temperature on surface (in K) | Power (in mW) |
|--------------------------------------|-------------------------------|---------------|
| Active vent | >973 | 700 |
| Northeast crater | 578 | 400 |
| North Mutnovsky springs (east group) | 371 | 19 |
| North Mutnovsky springs (west group) | 383 | 9 |
| Dacha springs | 371 | 73 |
| Pereval springs | 369 | 9 |
| Upper Zhirovskie springs | 369 | 18 |
| Voinovskie springs | 366 | 8 |
| Vilucha springs | 363 | 12 |

Valley of Geysers

| Geyser | Temperature (in K) | pH |
|--------------|--------------------|------|
| Pervenets | 371–372.6 | 8.35 |
| Troynoy | 371 | 8.5 |
| Sakharniy | 371 | 8.74 |
| Sosed | 371 | 8.2 |
| Konus | 371 | 8.2 |
| Maliy | 371 | 8.74 |
| Bol'shoy | 371 | 8.74 |
| Schel' | 371 | 8.74 |
| Velikan | 371 | 8.74 |
| Zhemchuzhniy | 371 | 8.74 |
| Fontan | 371 | 8.74 |
| Dvoynoy | 371 | 8.74 |

Iceland

Iceland located on the Mid Atlantic Ocean Ridge has about 130 volcanoes. Over the past 500 years, Iceland's volcanoes have erupted roughly a third of the total global lava output.



Iceland

- Geothermal gradients as high as 400 K/km were recorded in volcanically active areas, while ranges from about 50 K/km to about 150 K/km outside the volcanic zones.
- Hot springs are found throughout Iceland (but rare in the eastern basalt area). There are about 250 low-temperature geothermal areas (regions with temperatures ~ 423 K at depths $\sim 1,000$ m), with a total of about 800 hot springs, with an average temperature of water in a hot spring 348 K.
- Waters from hot springs and wells at Hveravellir geothermal field have a temperature in the range of 369–385 K and a pH of 9.32–9.44.

| Area/system | Number of production wells | Average production (in kg/s) | Reservoir temperature (in K) |
|-------------------------------|-------------------------------|---------------------------------|---------------------------------|
| Svartsengi (SW-Iceland) | 10 | 380 | 513 |
| Laugarnes (SW-Iceland) | 10 | 160 | 400 |
| Reykir (SW-Iceland) | 34 | 850 | 343–370 |
| Nesjavellir (SW-Iceland) | 11 | 390 | 553–613 |
| Hamar (N-Iceland) | 2 | 30 | 337 |
| Laugaland (N-Iceland) | 3 | 40 | 368 |
| Krafla (N-Iceland) | 21 | 300 | 483–613 |
| Urridavatn (E-Iceland) | 3 | 25 | 348 |
| Gata (S-Iceland) | 2 | 17 | 373 |
| Bjarnarflag (Námafjall field) | – | 12.5 | 553 |
| Reykjanes (SW-Iceland) | – | – | 548–583 |
| Theistareykir (NE-Iceland) | – | – | 613–653 |
| Húsavík (NE-Iceland) | – | – | 393–403 |

New Zealand

New Zealand is one of the most active regions in the Pacific Ring of Fire

Almost all of the highly active geothermal fields in New Zealand, are located within the geologically young Taupo Volcanic Zone, which has shallow reservoirs/aquifers (500–1,500 m depth) containing boiling or near-boiling water/ fluids (523–573 K).



Thermal characteristics of geothermal fields, thermal and pH characteristics of waters in the Taupo Volcanic Zone

| Geothermal field | Maximum reservoir temperature (K) | Water temperature on discharge (K) | pH |
|---|-----------------------------------|------------------------------------|---------|
| Atiamuri | 438–473 | 338–346 | 7 |
| Broadlands (Ohaki) | 581 | 371 | <3–8 |
| Golden Springs | – | 313–333 | 6.5–7 |
| Crater Lake (Ruapehu) | – | 308–371 | <3 |
| Horocho | >433 | 352 | 8.5 |
| Kawerau | >588 | 371 | 6.5–7 |
| Mokai | 597–599 | 371 | <2–7.6 |
| Orakeikorako | 538 | 338–371 | 7–9.2 |
| Orakeikorako | – | 313–363 | 2–3.5 |
| Mangakino (excl. Whakamaru or Ongaroto) | 527 | 371 | 8.5 |
| Moutohora | – | 371.5 | <4 |
| Reporoa (excl. Golden Springs) | 513 | 371 | <3–8.7 |
| Rotokawau (Rotokawa) (Rotorua) | 428 | 320–323 | 6.5–6.7 |
| Rotokawa (Taupo) | 593 | 313–363 | 2.5–5.5 |
| Rotorua | 523 | 363–371 | <2–9 |
| Taheke | – | 313–372 | 2–5 |
| Tarawera | – | 293–371.6 | 7–8.7 |
| Tauhara (excl. Wairakei) | 552 | 335–371 | 2–8 |
| Te Kopia | 514 | 318–378 | <2–7.5 |
| Tikitere (Hell's Gate, Ruahine Springs) | 503 | 313–371 | 2–7.6 |
| Tokaanu | 523 | 371 | 3.5–8.7 |
| Tongariro | 523 | 343 | <2.5 |
| Waihi (see Tokaanu) | 523 | 371 | 6.6–7.9 |
| Waimangu (and Rotomahana) | 543 | 328–371 | <3–9 |
| Waiotapu | 568 | 371 | 1.8–8 |
| Whakaari (White Island) | 1,073 | 573–773 gas | |
| Ketetahi | – | 327–411 | 2.0–6.5 |
| Waikite Valley | – | 371 | 8–8.7 |
| Wairakei | 544 | 371 | <3 |
| Whangairorohea | – | 311 | 7.4 |
| Okatina | – | 312 | 7 |
| Longview Road | – | 306–365 | 3.7–6 |
| Opakeke (Opateketeke) | – | 318–371 | 3–8 |
| De Bretts (Terraces) | – | 323–353 | 7.8 |
| Waiora Valley | – | 371 | 3 |
| Ngatamariki | 563 | – | – |
| Rotoma | 493–523 | – | – |

Japan

- Japan, another segment of the Pacific Ring of Fire, is a very seismically and volcanically active region and has one of the highest potentials for geothermal energy in the world, due to the large number of hot springs.
- The main water reservoirs in the Kuju-Beppu Graben are at least 500 m deeper (at 1,000–2,000 m), and they have a water temperature of about 30–80 K lower (493 K) than those in the Taupo Volcanic Zone in New Zealand.



Japan

| Hot spring | Temperature on discharge (in K) | pH |
|---------------------------------------|---------------------------------|-----------|
| Kannawa Ishimatsu | 373 | 7.7 |
| Kannawa Chinoike | 333 | 2.4 |
| Myoban Yamadaya | 340.5 | 1.7 |
| Kamegawa Shinoyu | 329.6 | 8.2 |
| Old-City Kimura | 328.6 | 7.4 |
| Hotta Hotta | 348.5 | 6.2 |
| Kankaiji Jizouyu | 323 | 6.9 |
| Nasu | 318.2–346.1 | 1.5–1.7 |
| Takao | 307 | 5.6 |
| Sandogoya | 365 | 2.6 |
| Nikko-yumoto | 328.2–342 | 6.0–6.8 |
| Yunohanzawa, Hakone | 306–350.2 | 2.0–2.6 |
| Ashi-no-yu | 313–314 | 6.4 |
| Kowakudani | 334 | 8 |
| Ubako | 318 | 5 |
| Kusatsu | 330–338 | 1.5–1.6 |
| Isobe | 288.5 | 6.6 |
| Ikao | 317.2 | 6.2 |
| Tanigawa | 315.3 | 8.2 |
| Tateyama | 315 | 1.2 |
| Arima | 334–364.3 | 6.0–6.6 |
| Hakone Springs: ^a | | |
| Zone I | 322.7 | 2.9 |
| Zone II | 330.5 | 8.1 |
| Zone III | 364.5 | 7.7 |
| Zone IV ^a | 338.5 | 8.4 |
| Zone IVb | 329 | 8 |
| Ground water | 286.1 | 7.2 |
| Obuki ^b (Tamagawa springs) | 369–372 | 1.22–1.42 |
| Owakudani | 309.3–369 | 2.48–4.25 |
| Azuma | 320–373 | Neutral |
| Zao | 338 | Neutral |
| Hijiori | 357 | Neutral |
| Akakura | 347 | Neutral |
| Akayu | 336 | Neutral |

^a Zonation of Hakone spring area after (Oki and Hirano 1974)

^b Fluctuation of temperature (1962–2000) and pH (1959–2000) after (Yoshiike 2003)

- Analysis of 3,686 hot springs in Japan showed that they belong to relatively low-temperature geothermal fields with a reservoir temperature (533 K).
- One of the largest hot spring fields in Japan is the Beppu Geothermal area. The source of heat for the high temperature manifestations is related to the middle-to-late Pleistocene Yufu volcanic formations (~0.32 Myr) and lava domes younger than 100,000 years.
- Japan has the highest number of hot springs in the world, but very few geysers, due to the greater of the main thermal sources than in active geyser regions, as well as the much lower temperatures at shallow depths than in such areas.

Rate of discharge of hot springs

- None of most powerful hot springs on the Earth comes close to the average discharge rates of the most powerful non-thermal springs on the Earth (> 2,800 L/s).
- Warm and non-thermal springs shows that non-thermal springs require higher flow rates, and in turn require the presence of more permeable aquifers than those in warm or hot spring waters.
- Low flow rates ensure more lengthy contact of water flowing through the aquifer provides better heating conditions for the spring water (larger temperature).

Rate of discharge of some hot springs and hot spring systems

| Spring or system of springs | Average rate of discharge (in L/s) | Magnitude of spring |
|---|------------------------------------|---------------------|
| Yellowstone National Park, USA: | | |
| Total for all thermal features in the Park | 3,091 | – |
| Upper Geyser Basin | 696 | 2 |
| Lower Geyser Basin | 662 | 2 |
| Midway Geyser Basin | 252 | 3 |
| Mammoth Hot Springs | 57 | 3 |
| Shoshone Geyser Basin | 66 | 3 |
| Heart Lake Geyser Basin | 104 | 3 |
| West Thumb Geyser Basin | 47 | 3 |
| Kamchatka Peninsula, Russia: | | |
| Paratunka geothermal field | 221 | 3 |
| Esso geothermal field | 168 | 3 |
| Anavgay geothermal field | 34 | 3 |
| Springs in vicinity of rivers Goryachaya and Zholtaya | 80–110 | 3 |
| Taloviye | 6 | 5 |
| Krayevedcheskie | 7 | 4 |
| Verkhne-Schapinskie | 40 | 3 |
| Valley of Geysers, Kamchatka Peninsula: | | |
| Total for the basin | 300–315 | 2 |
| River Geysernaya | 3,340 | – |
| Podskalniy | 5 | 5 |
| Bolshoy | 3 | 5 |
| Geyser Velikan | 2.4 | 5 |
| Deildartunguhver, Iceland | 180 | 3 |
| Hveravellir, Iceland | 35–40 | 3 |
| Gata, Iceland | 10–22 | 4 |
| Thorleifskot, Iceland | 70–80 | 3 |
| Yamagata prefecture, Japan, total of 140 springs | 916 | – |
| Zao, Yamagata prefecture, Japan | 92 | 3 |
| Bepu Geothermal Area, Japan, Total | 579 | 2 |
| Obuki, Tamagawa springs, Japan | 166 | 3 |

Oceanaic Vents

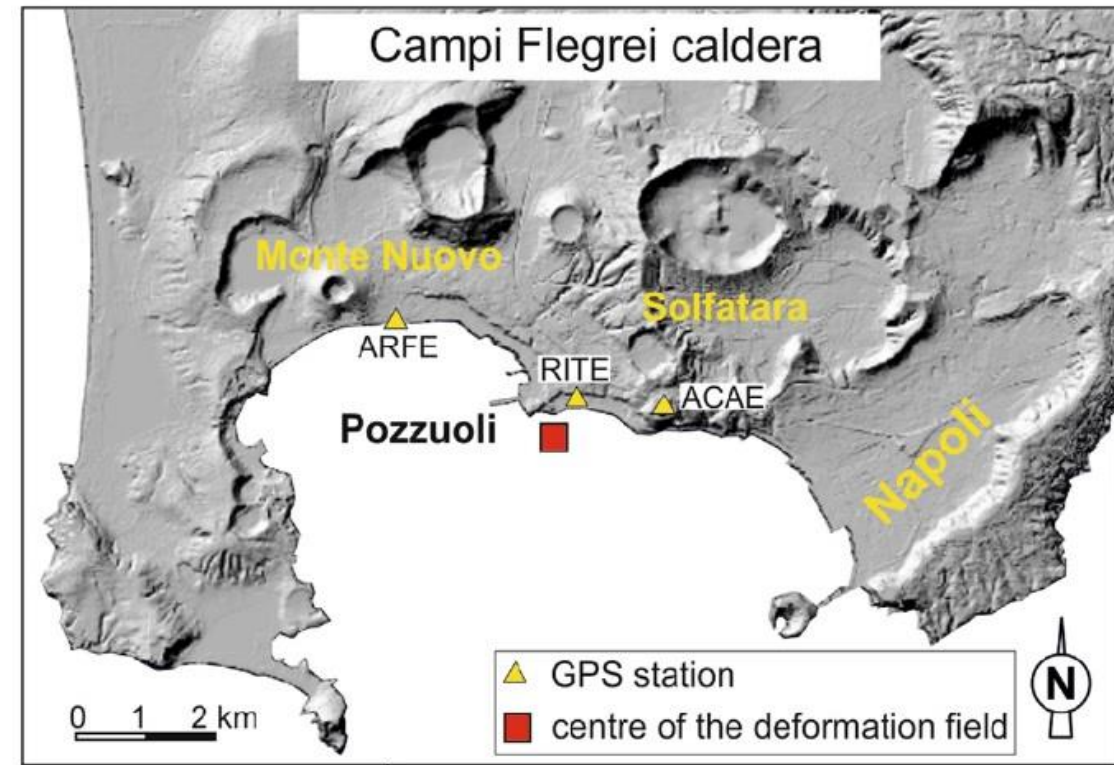
- Hot springs in oceanic areas or vents have a temperature ranging from 623 K in the axial zone of ridge to 293 K far away.
- Extra-hot vents are divided into ‘**black smokers**’, emitting hydrothermal fluids with temperatures mostly within the range of about 573–673 K, containing dissolved metals and ‘**white smokers**’, with temperatures of 523–573 K, with almost no metals in the ejected fluids.

Temperatures of discharged fluids for some black and white smokers

| Vent type and location | Temperature of water (in K) | References |
|---|-----------------------------|-------------------------|
| <i>Black smokers:</i> | | |
| Mid-Atlantic Ridge: | 613–663 | Petersen et al. (2000) |
| TAG at 26°N | 636 | Fouquet et al. (1998) |
| TAG at 26°08’N | 583–593 | Pedersen et al. (2010) |
| Loki’s Castle, 73°30’N and 8°E | 573–673 | McCaig et al. (2007) |
| low-angle detachment fault at 15°45’N | 600–654 | Herzig et al. (1998) |
| Black smoker complex (TAG-1) at 26°N | ~ 638 | German et al. (2010) |
| Rainbow hydrothermal field at 36°14’N 5°S | 680–737 | Haase et al. (2007) |
| East Pacific Rise: | | |
| Galapagos Rift, 21°N | 623 | Köhler et al. (1994) |
| East Pacific Rise | 603–678 | Tivey (2007) |
| 9–10°N | 656 | John et al. (2008) |
| Site #7 at 12° 49’N | 653 | Graham et al. (1988) |
| Site #10 at 10°58’N | 620 | Graham et al. (1988) |
| Central Indian Ridge: | | |
| Kairei vent field at 25°19.239S, 70°02.429 | 579–638 | Van Dover et al. (2001) |
| Edmond vent field at 23°52.689S, 69°35.809E | up to 655 | Van Dover et al. (2001) |
| Juan de Fuca Ridge: | | |
| Middle Valley in the area of active venting | up to 633 | Turner et al. (1993) |
| ‘Sully’ and ‘Puffer’ vents | 628–633 | Crone et al. (2006) |
| Izena Hole, Mid Okinawa Trough | 593 | Sakai and Nozaki (1995) |
| <i>White smokers:</i> | | |
| Mid-Atlantic Ridge: | | |
| TAG at 26°08’N, the Kremlin area | 533–573 | Herzig et al. (1998) |
| East Pacific Rise: | | |
| Galapagos Rift, 21°N | 543 | Köhler et al. (1994) |
| 9–10°N | 480 | John et al. (2008) |
| Site #9 at 11°14’N | ~ 543 | Graham et al. (1988) |
| Site #3 at 12°51’N | >483 | Graham et al. (1988) |
| Juan de Fuca Ridge: | | |
| Middle Valley in the area of active Venting | 457–553 | Turner et al. (1993) |
| Northern Okinawa Trough | 551 | Sakai and Nozaki (1995) |

Phlegrean Fields

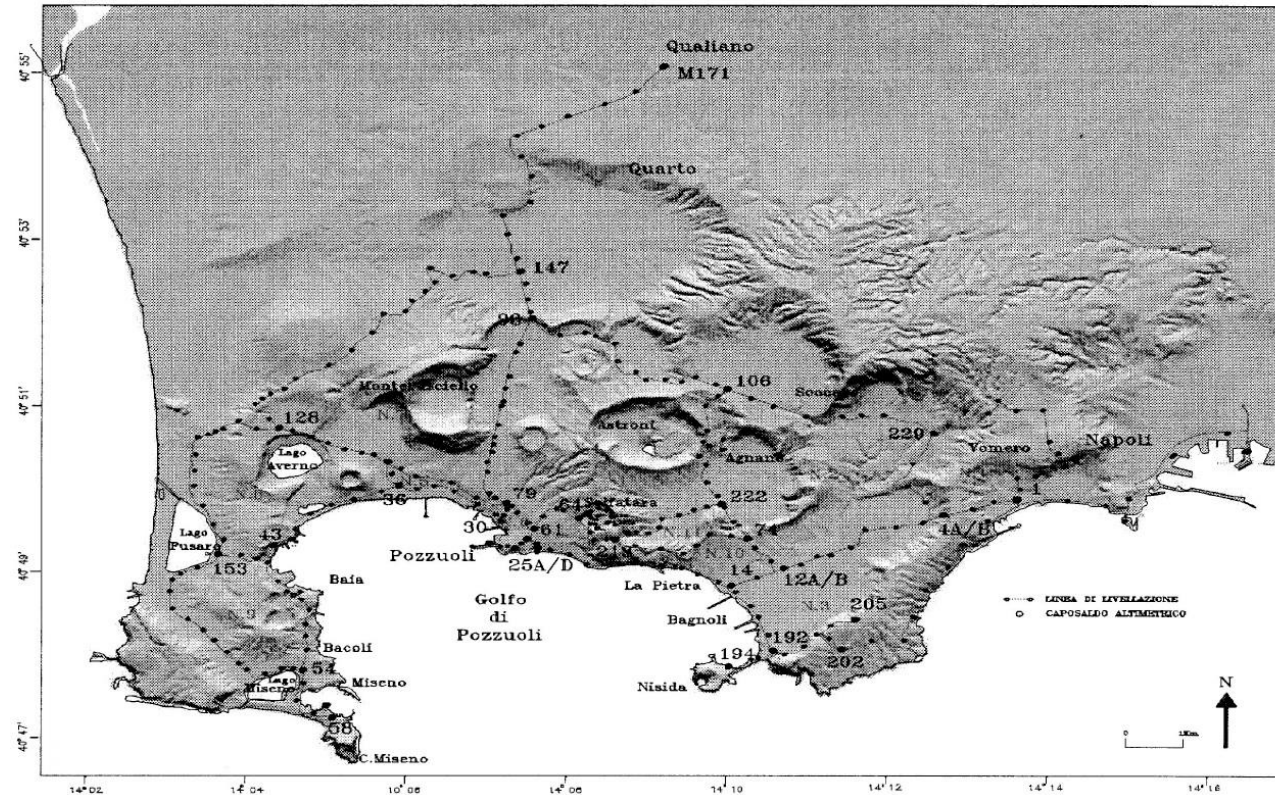
- The Phlegrean Fields consist of two nested calderas, resulting from the interaction between regional and local tectonics. The formation of the larger, outer caldera has been related to the Campanian Ignimbrite eruption (37.000 years ago, with 80 km^3 of material), while the inner caldera formed during the eruption of the Neapolitan Yellow tuff (12.000 years ago, with 10 km^3 of material).
- Most eruptive activity has occurred during two periods: from 10.000 to 8.000 years ago and from 4.500 to 3.700 years ago, while in the last 2000 years there was only one eruption occurred in 1538 which gave the birth to Monte Nuovo.



Phlegrean Fields

- During its history, the Phlegrean Fields have alternated phases of uplift and subsidence over a range of timescales. In the last century, there were two major uplift and seismic episodes ('bradyseisms') (1969–1972 and in 1982–1984, respectively). These episodes have produced a total vertical displacement of $3.8 \pm 0.2\text{m}$ and have been accompanied by thousands of shallow earthquakes.
- Since 1985, Phlegrean Fields have been slowly subsiding, which has been interrupted by a few minor uplift events. In 2005, there was new inflation, with a maximum vertical displacement of $\sim 23\text{ cm}$ by June 2014. This last stage was accompanied by weak seismicity, by a strong increase in fumarolic activity, and by compositional variations in the fumarolic effluents.

Levelling net

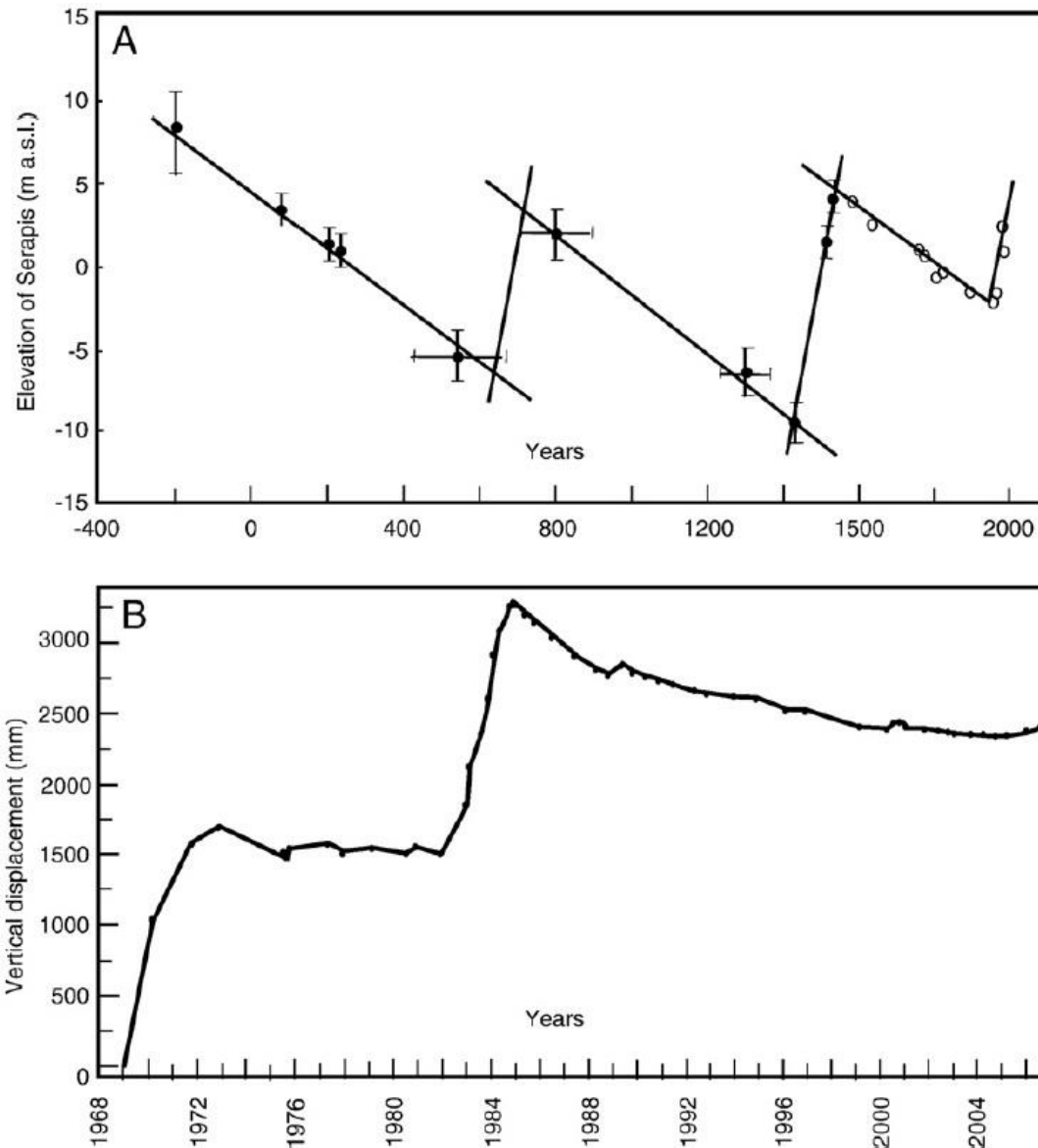


Phlegrean Fields

Vertical movements history at Serapis Temple in Pozzuoli.

Black circles: constraints found from radiocarbon and archaeological measurements.

Serapis Temple



Vertical ground displacements as recorded at Pozzuoli Harbour by levelling data (1969–2006).

Phlegrean Fields

Models of CF bradyseism fall into three main categories:

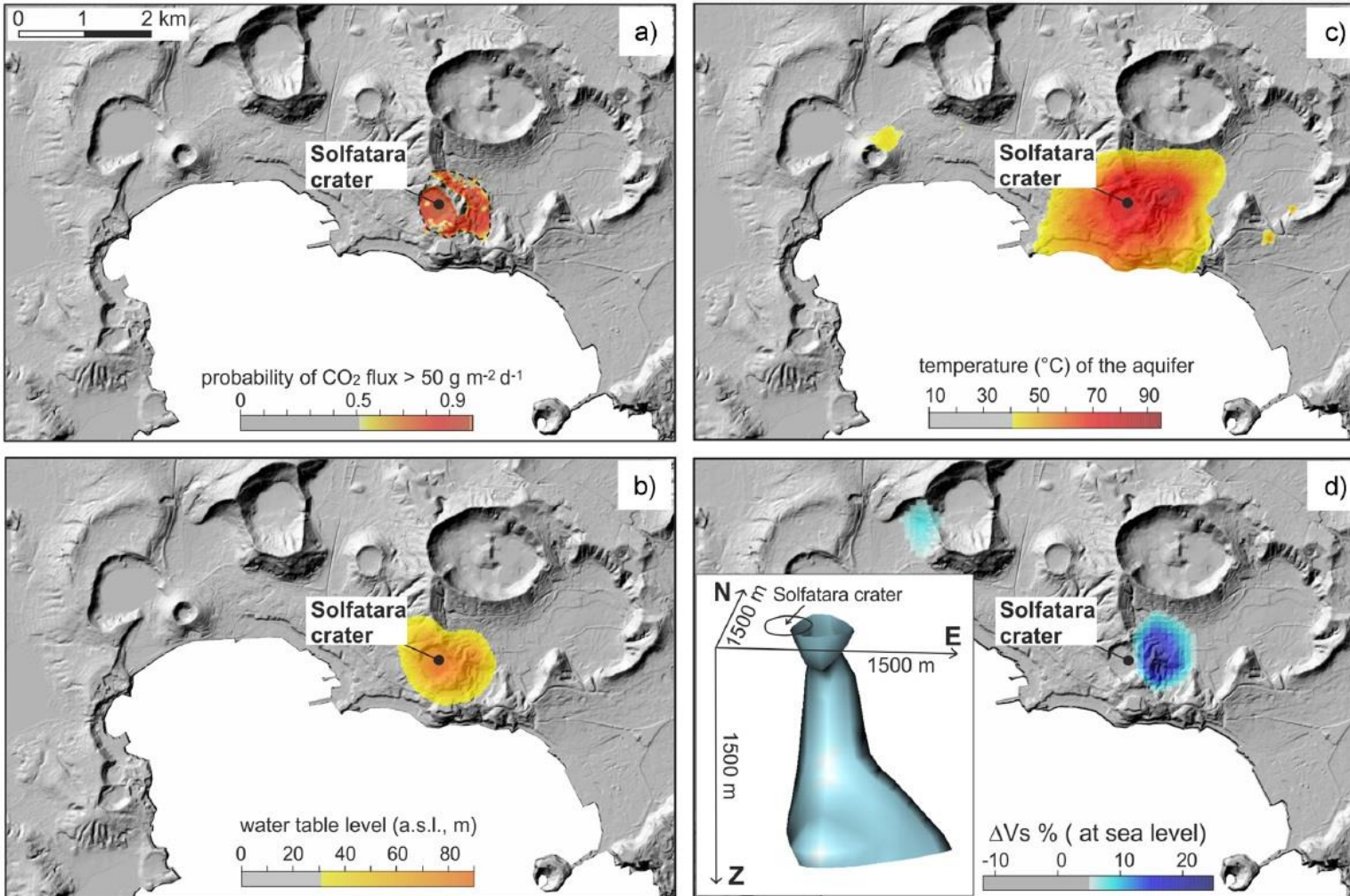
- One scenario relates ground surface deformation (GSD) directly to emplacement of a fresh batch of magma at shallow depth.
- A second class of models also involves fresh magma input (recharge) to trigger bradyseism not in response to pressurized magma, but rather due to the injection of magmatic fluids into overlying crust.

In the second case:

- Injection of high temperature magmatic fluid into the shallow hydrothermal reservoir system induces fluid overpressures that may cause the host rocks to **inflate** depending on the mechanical properties of the local crust.
- **Subsidence** results from a decrease in the flux of magmatic fluid entering the hydrothermal system or as a response to rapid permeability increases and hence pore pressure decreases.

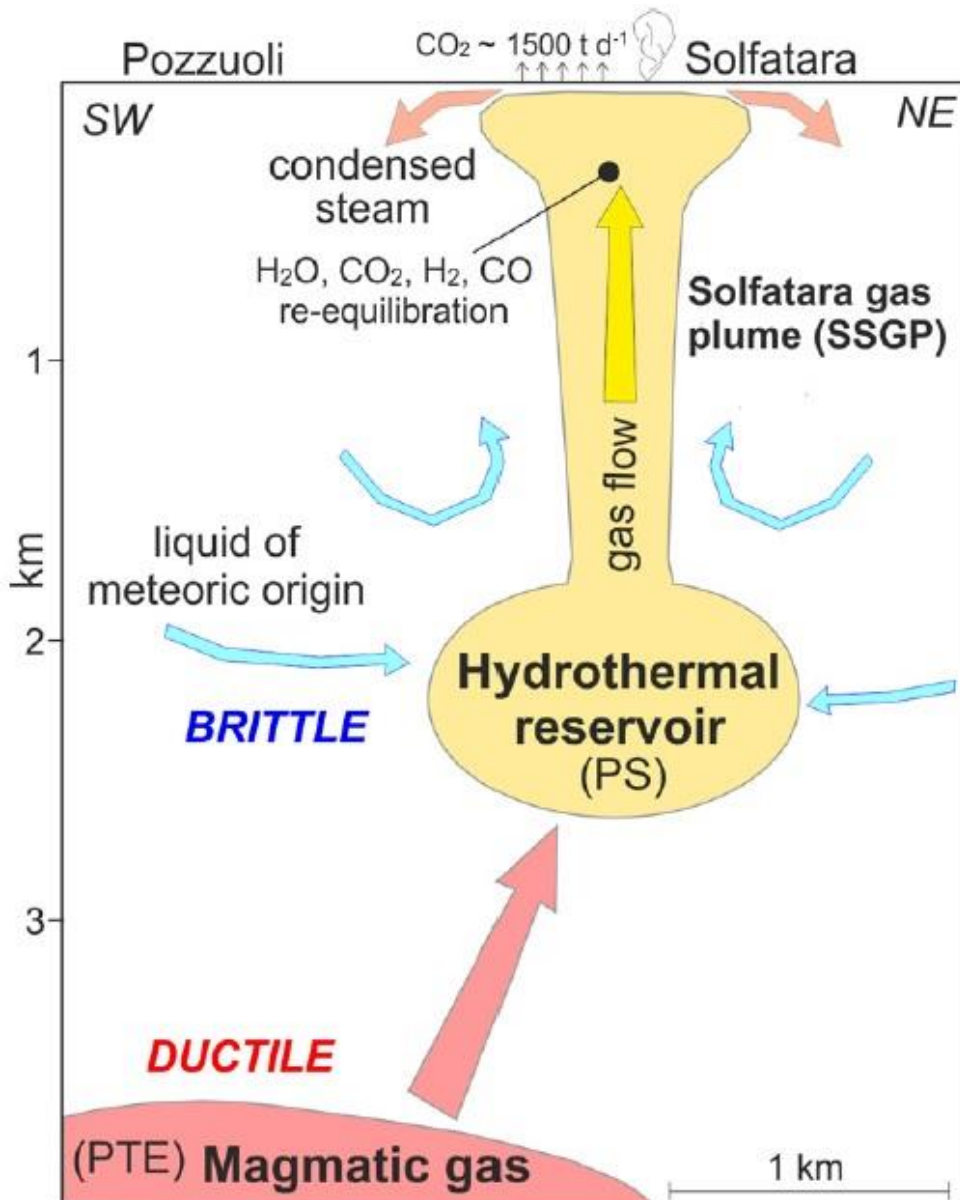
Phlegrean Fields

The presence of a gas plume in the subsoil of Solfatara crater is evidenced by different factors:



- The large amount of CO₂ release from diffuse degassing processes at Solfatara and its surroundings (~1500t/d), which is more compatible with the presence in the subsoil of a large zone where there is a gas phase, rather than with a boiling process of a liquid.
- At Solfatara, the aquifer is anomalously high for both the water table height and the water temperature. The height of the water level indicates that a pressurized gas plume sustains the aquifer here.
- The S-wave seismic velocity (*V_s*) models clearly delineate a vertical, high-*V_s* structure from the surface down to at least 1.5 km, which can be attributed to the presence of gas instead of liquid (*V_s* in dry samples of tuffs are systematically 10% to 50% higher than the same samples saturated with liquid).

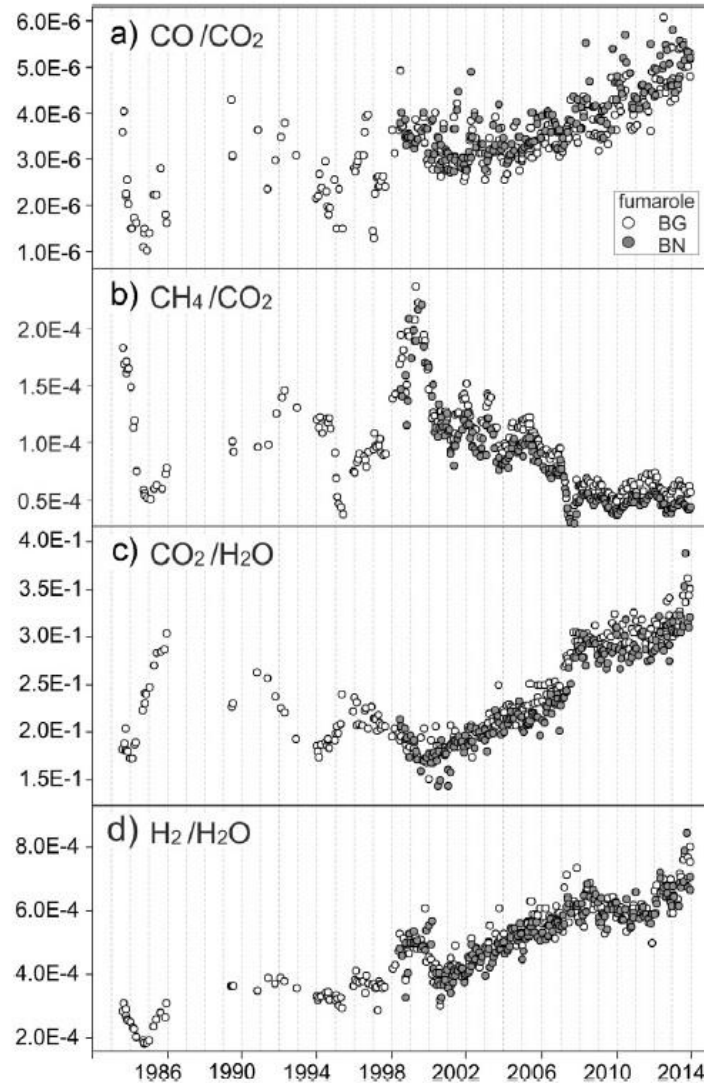
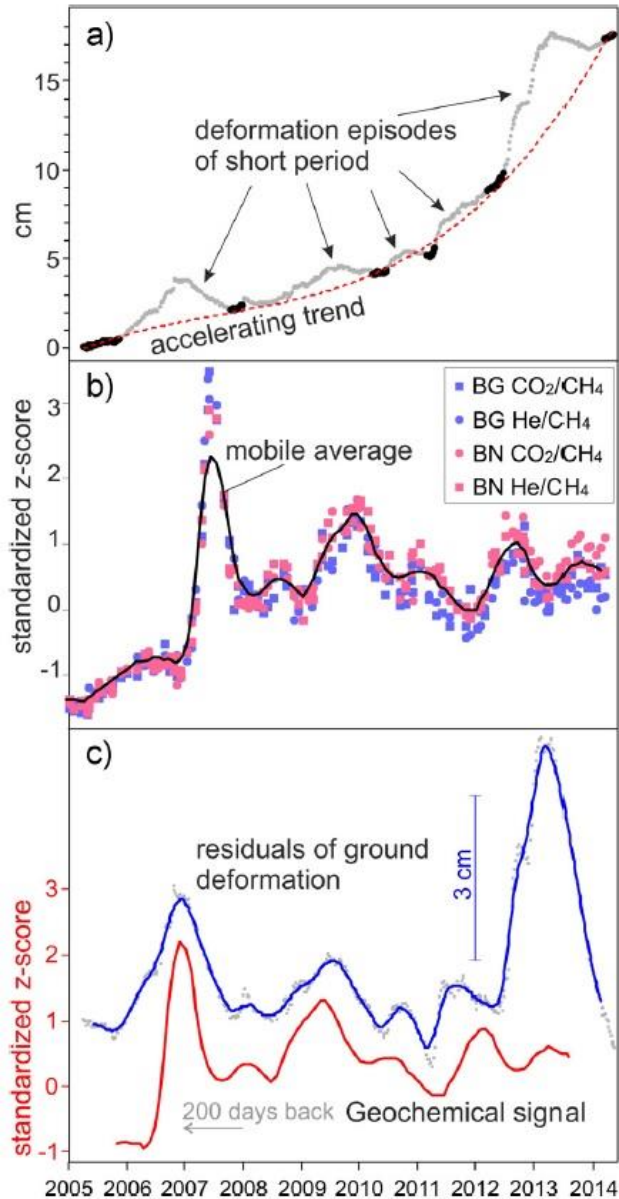
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- Hot gases separate from the magma at depth, ascend toward the surface, mix with boiling meteoric water to form a gas plume that feeds fumaroles and diffuse soil degassing at Solfatara.
- The system consists of: (i) a deep zone of gas accumulation, located at ~4 km in depth (where also small batch of magma may exist), which supplies hot gas to the system; (ii) a shallower reservoir (~2 km in depth), where magmatic fluids mix and vaporize liquid of meteoric origin that forms the Solfatara gas plume. The observed deformation would be controlled by pressure changes in these two distinct sources.

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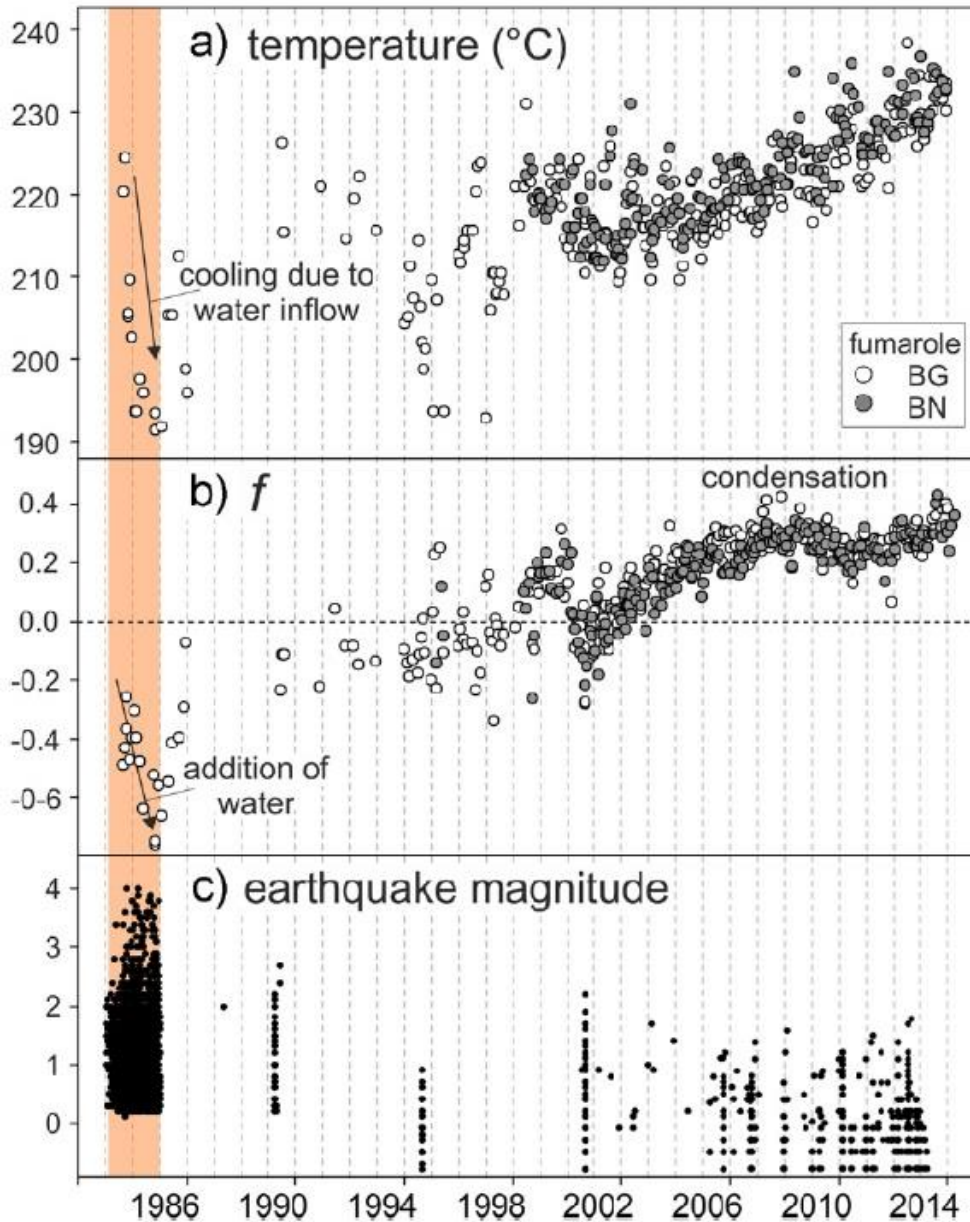
Comparison between ground deformation and geochemical signals



Chiodini et al., 2015, EPSL, 414

- During the phases of uplift, the CH₄ content of the fumaroles decreased, while the content of other gases of prevalent magmatic origin, such as CO₂ and He tend to increase.
- The sharp increase in the proportion of the magmatic component over relatively short periods, was interpreted as resulting from magmatic fluid injections into the hydrothermal system.
- The delay between the uplift and geochemical signals would represent the transfer time of the magmatic fluids from the input zone to the fumarolic discharge areas.
- CO and CH₄ qualitatively indicate a temperature increase for the entire hydrothermal system, which is likely caused by the increment in the flux of the magmatic fluids that enter the hydrothermal system.
- The increment in the flux of the deep hot fluids would cause water vapor condensation within and at the border of the gas plume and heating of the rock by the latent heat release during condensation. The condensation of the steam may causes the continuous increase in non-condensable gases (CO₂) in fumaroles.

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- From 1983 to 2014, the temperature estimations of the fumaroles range from 190°C to 240°C and show a continuous increasing trend during the 2005–2014 phase of ground acceleration.
- T and water fraction (f) have a marked negative anomaly during the 1983–1984 seismic, which might have been caused by a T decrease, as a result of the input of liquid water in the Solfatara plume (e.g., due to the increase the rock permeability by creating new fractures during the earthquakes).

From 2003-2014, the total amount of heat released by steam condensation was $\sim 6.2 \times 10^{12} \text{ kJ}$. This would produce a 5°C heat increase in a mass of $\sim 1.25 \times 10^{12} \text{ kg}$ of rock, which corresponds to a volume of 0.625 km³ (with a density of 2000 kg/m³).

For a volumetric expansion coefficient of $3 \times 10^{-5} / ^\circ\text{C}$, the volume increase would be of $\sim 0.94 \times 10^5 \text{ m}^3$.

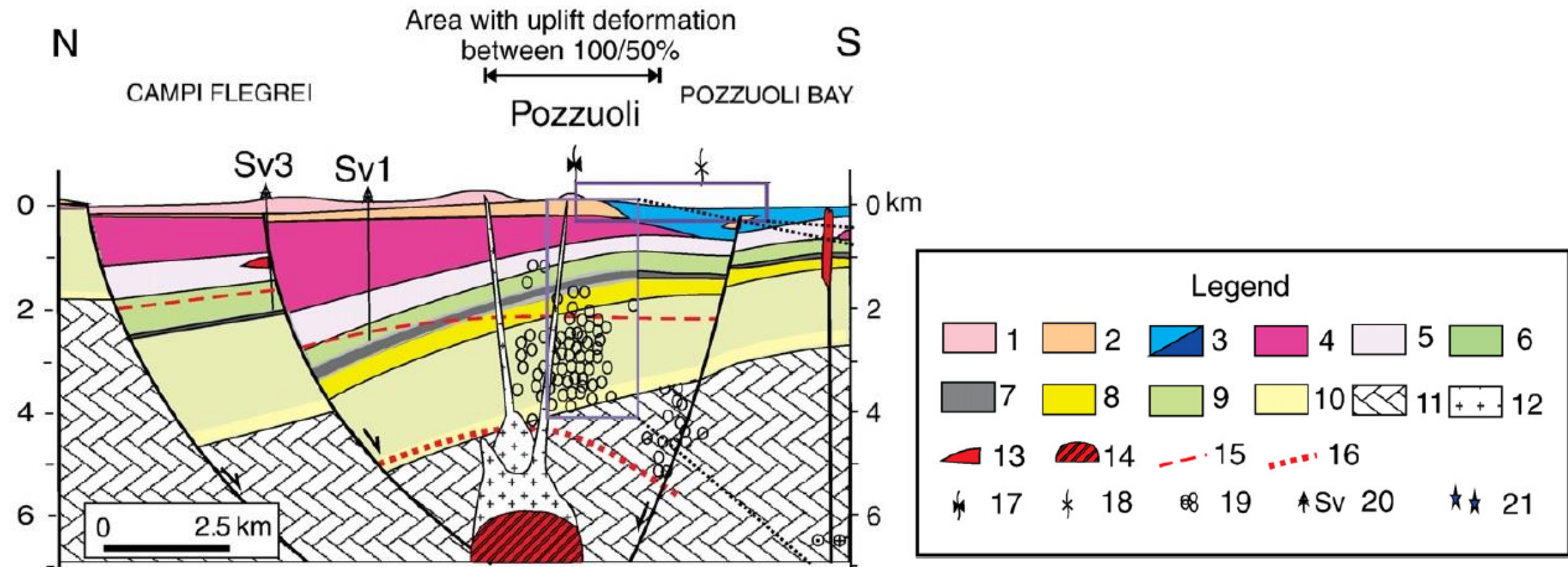
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- According to the model of Chiodini et al., 2015, two main processes contribute to the ongoing CF bradyseism:

- 1) Transient episodes of gas pressurization that are accompanied by fluid transfer from the deep magmatic gas zone to the shallower parts of the hydrothermal system, which trigger the short-term uplifting episodes;
- 2) A long-term process of heating of the system that causes (or contributes to) the 10-year-long pattern of accelerating ground deformation.

- An alternative model for subsidence and uplift of the CF of Lima et al., 2009 is based on the linkage between bradyseism and magma body cooling and concomitant crystallization and fluid phase exsolution from melt, with eventual expulsion of lithostatically pressured fluid into overlying country rock.

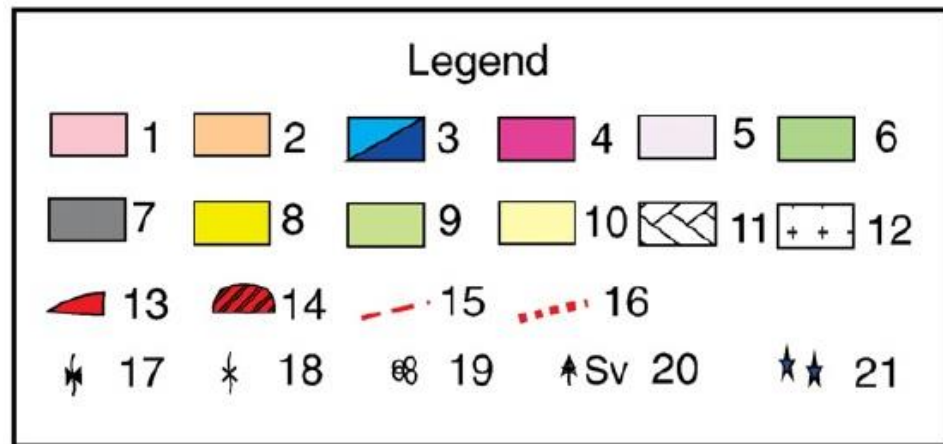
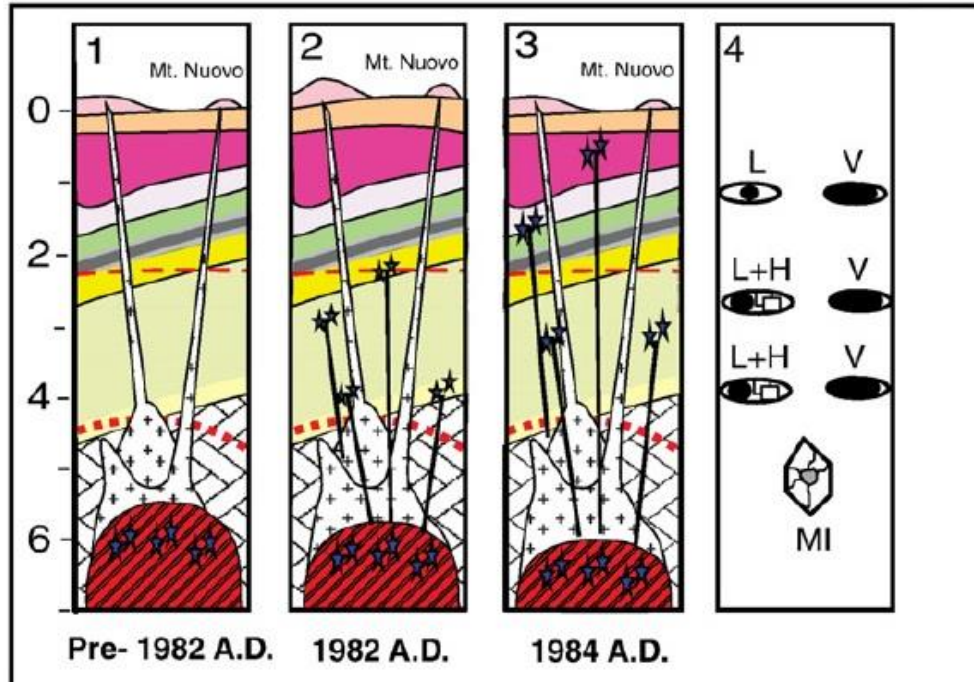
Phlegrean Fields



Lima et al., 2009, Earth-Science Reviews, 97

1. Holocene volcanics; 2. Neapolitan Yellow Tuff; 3. Main sediments post 39 ka; 4. Campania Ignimbrite and pre-CI tuffs; 5. Middle Pleistocene sandstones, siltstones and volcanics; 6. Middle Pleistocene marine sediments (sandstones and siltstones — unit C in text); 7. Fine grained Middle Pleistocene marine sediments (claystones and siltstones — Unit B in text); 8. Middle Pleistocene deep water debris flows; 9. Lower Pleistocene marine sediments; 10. Continental deposits and conglomerates; 11. Meso-Cenozoic substrate; 12. Crystallized magma; 13. Volcanic bodies; 14. Magma body; 15. Thermometamorphic boundary; 16. Impermeable zone surrounding the crystallizing magma body. 17. Pozzuoli Anticline; 18. Pozzuoli Bay Syncline; 19. 1983–84 earthquake hypocenters; 20. Deep geothermal wells; 21. Magmatic fluids.

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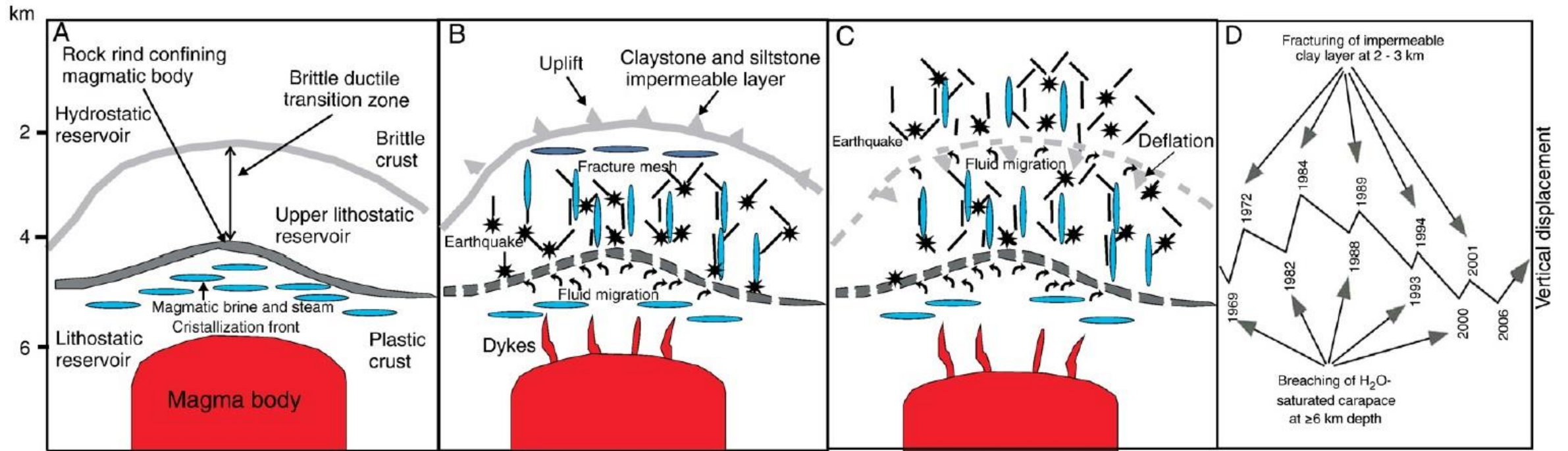


- (1) Following the 1538 eruption, the magma body became a closed system, and magmatic volatiles accumulated below impermeable crystallized carapace.
- (2) In 1982, the carapace confining the magmatic system fractured, allowing magmatic fluids to enter the overlying rocks beneath the low-permeability caprock (Unit B), causing vertical ground deformation.
- (3) Ground deformation that began in 1982 ended and deflation began when fractures penetrated the low-permeability cap rock, allowing the deep fluids to migrate into the shallow hydrostatic aquifers and flow toward the surface.

Lima et al., 2009, Earth-Science Reviews, 97

L = Liquid-rich inclusion; V = Vapour-rich inclusion; L + H = Halite-bearing fluid inclusion; MI = Melt inclusion).

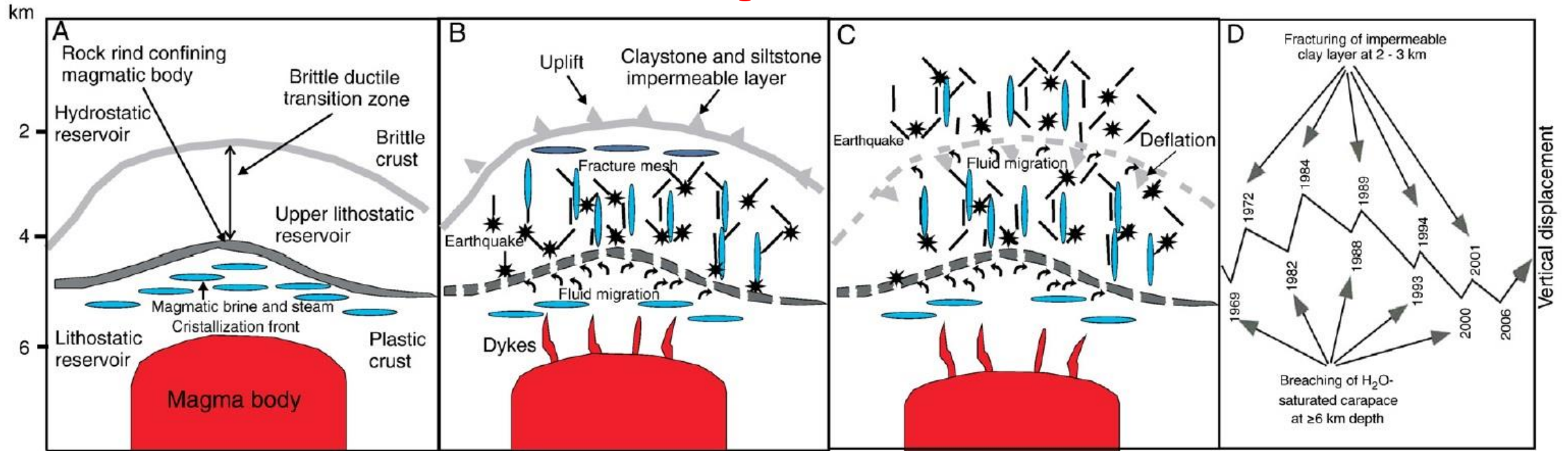
Phlegrean Fields



Lima et al., 2009, Earth-Science Reviews, 97

- During initial crystallization of the under-saturated melt, the volatile content of the melt increases as anhydrous phases precipitate (melt becomes saturated in volatiles and the pressure in the magma body increases). This fluid remains geopressured until fracture propagation enables the expulsion of magmatic fluids.
- High salinity brine and vapour may exsolve directly from the magma and due to its greater density and viscosity, the high salinity liquid usually “ponds” in the deeper portions of the system, while the lower salinity, low-viscosity vapour phase migrates to shallower levels of the system. Geopressured fluids can escape into the overlying rocks when rock rind around the magma chamber is breached (inflation).
- When the overlying impermeable rocks fracture, the low salinity gas moves towards the surface and interacts and mixes with meteoric water (or seawater) present in fractures and pores at shallow depths, to produce a low salinity boiling assemblage (deflation).
- According to the model, the magma body is presently isolated from the overlying aquifers and the fraction of exsolved fluid is building within the magma chamber.

Phlegrean Fields



Lima et al., 2009, Earth-Science Reviews, 97

- Uplift occurs when magmatic fluids enter in the rocks beneath the lower permeability cap rock (~2.5–3.0 km).
- Uplift ends and deflation begins when fractures penetrate the lower permeability cap rock, allowing the deep magmatic fluids to migrate into the shallow aquifers and flow toward the surface.
- Bradyseism is driven by the transient connection between the deeper lithostatic reservoir and an overlying more permeable hydrostatic one. Magma solidification and associated magmatic fluid generation occur on a timescale of 10^3 – 10^4 yr. Transient fracture propagation events that connect the lower lithostatic reservoir with the upper hydrostatic one, accompanying irreversible fluid decompression, occurs on a timescale of 1–10 yr.
- A magmatic eruption may follow if the reduced confining pressure on the magma leads to a runaway process of bubble formation, ascent and growth (if an overpressure is generated by the magmatic–hydrothermal fluid on rocks above a low-permeability cap rock).
- As the magma body cools and crystallizes and the H_2O -saturated carapace migrates to greater depth, the energy and volume change associated with volatile exsolution decreases and the magnitude of future uplift events is likewise expected to decrease.