Heat and mass transport in icy worlds
(melting/freezing, porous flow)

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The Solar System:

ring system of Saturn: not in scale

Scale (proportions)
20 000   40 000   60 000 km

Sun
Ø 1 392 684 km

Vesta

Juno

A s t e r o i d   b e l t

Ceres

Pallas

Callisto
Ganymede
Europa
Io

Iapetus
Hyperion
Titan
Rhea
Dione
Tethys
Enceladus
Mimas

Oberon
Titania
Umbriel
Ariel
Miranda

Triton

Eris

Makemake

K u i p e r   b e l t

Pluto
Charon

Haumea

Moon
Phobos
Deimos

Mercury **Venus** **Earth** **Mars**

Jupiter
Ø= 778 360 000 km
Ø 142 984 km
79 moons in total

Saturn
Ø= 1 433 500 000 km
Ø 120 536 km
62 moons in total

Uranus
Ø= 2 872 600 000 km
Ø 51 118 km
27 moons in total

Neptune
Ø= 4 499 400 000 km
Ø 49 528 km
14 moons in total

Moons with colored frames are not to scale.

Ø= = distance of the planet to the sun
Ø = average- or equatorial diameter

The major objects in the belt;
not in scale

Mercury
Venus
Earth
Mars
Jupiter
Saturn
Uranus
Neptune
(Pluto)

Scale representation of the distances of the objects to the sun
Scale: ▬ 1 AU (149.6 Mill. km)

The major dwarf
planets including the
biggest moon of Pluto;
in scale.
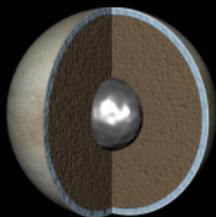
Europa
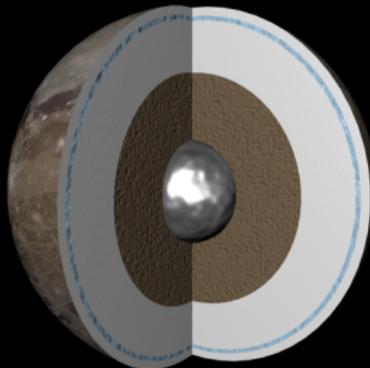
Ganymede
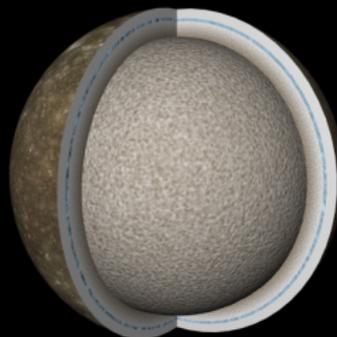
Callisto

Key
- ⬤ ice
- ⬤ water
- ⬤ rock
- ⬤ metal
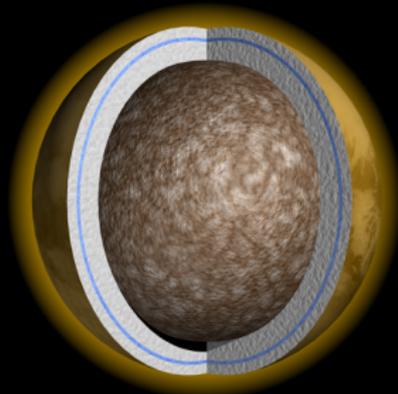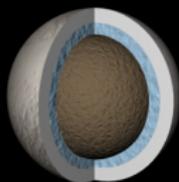
Eris
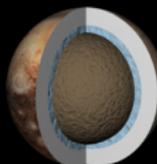
Pluto
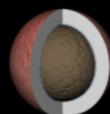
Sedna

Enceladus
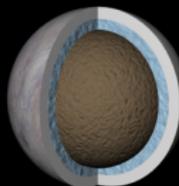
Titan

Triton
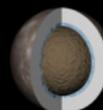
Titania
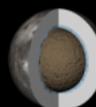
Oberon
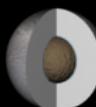
Rhea

Outlook

- •) solid state convection in planetary layers
- ••) balance laws and a boundary condition
- •••) solid-state convection with a phase change at the boundary
- •**v**) porous convection in the rocky core of Enceladus

Solid-state convection in planetary interiors

Conservation of mass:

$$\frac{\partial \rho}{\partial t} + \boldsymbol{\nabla} \cdot \rho \mathbf{V} = 0$$

Solid-state convection in planetary interiors

Conservation of mass:

$$\boldsymbol{\nabla} \cdot \boldsymbol{V} = 0$$

- ‣ incompressible fluids (not really true for large bodies), or at least anelastic approximation,

Solid-state convection in planetary interiors

Conservation of momentum:

$$\frac{1}{\mathrm{Pr}}\left[\frac{\partial \mathbf{V}}{\partial t} + \mathbf{V} \cdot \boldsymbol{\nabla}\, \mathbf{V}\right] = -\boldsymbol{\nabla} p^{*} + \boldsymbol{\nabla} \cdot \eta \left(\boldsymbol{\nabla}\mathbf{V} + \boldsymbol{\nabla}\mathbf{V}^{t}\right) + \mathrm{Ra}\, T \gamma_{\mathbf{z}}$$

- ‣ p^{*}: non-hydrostatic pressure,

- ‣ infinite Prandtl number fluids (diffusion of momentum way faster than diffusion of heat): non-linear inertial terms disappear
  ⇒ no mechanical turbulence, strongly simplifies dynamics!

- ‣ rheology is very complex... (temperature dependent, grain-size dependent and affects grain size, non-Newtonian)
  ⇒ strongly complexifies dynamics!
  ⇒ spectral approaches (spherical harmonics) lose elegance, grid-based methods used in practice.

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Rheology: exemple of ice I

Viscosity

$$\frac{1}{\eta} = \frac{1}{\eta_{diff}} + \frac{1}{\eta_{disl}} + \frac{1}{\eta_{gbs} + \eta_{bs}}$$

with

$$\eta_\bullet = \frac{d_\bullet^m}{A_\bullet \sigma_{II}^{n_\bullet - 1}} \exp\left(\frac{E_\bullet + pV_\bullet}{RT}\right)$$
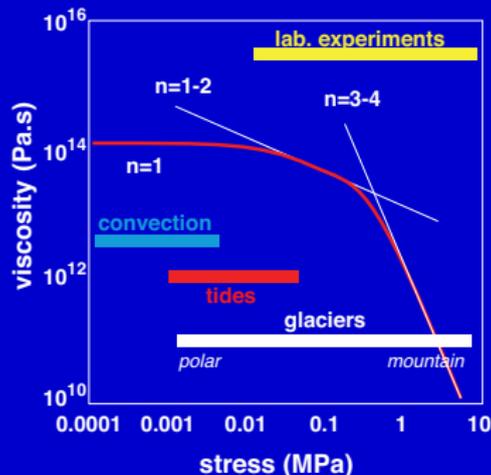
Solid-state convection in planetary interiors

Conservation of energy:

$$\frac{\partial T}{\partial t} + \mathbf{V} \cdot \nabla T = \nabla^2 T + \tilde{h}$$

- ‣ heat transfer as the sole control of time evolution,
- ‣ $\tilde{h}$: heat sources/sinks possibly involving secular cooling/heating, decay of long-lived radio-isotopes (mostly in rocks), tidal heating,
- ‣ typically solved with a *minmod* limiter approach for the advective flux.
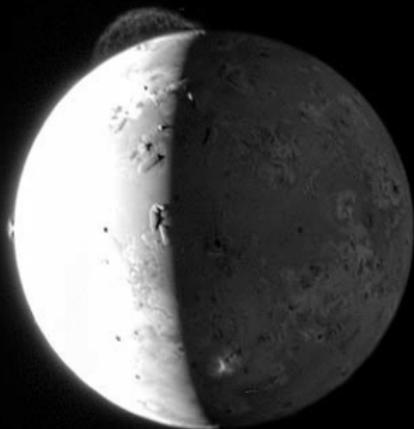
Convection with an open boundary

Heat pipe mechanism, Io *(O'Reilly and Davies, 1981)*

Fig. 1. Model of Ionian lithosphere. Magma at temperature T_m rises through vents and spreads across surface, cooling to surface temperature T_0. Weight of flows depresses lithosphere at subsidence velocity v, which equals resurfacing rate.

First Voyager observations paradoxical:

- very large surface heat flux (2 ± 1 $W.m^{-2}$ at the time, 2.4 ± 0.9 $W.m^{-2}$ today) $\Rightarrow$ thin conducting lithosphere (few kms),

- however large reliefs (*"apparently of silicate composition"*) in excess of 10 kms were observed...

Among other possibilities (dynamic support, low density roots), the authors favor the possibility of very efficient magmatic heat transport allowing a much thicker conductive lithosphere.

Heat pipe mechanism, early Earth *(Moore and Webb, 2013)*

On Earth, the > 3.2 Gyr old geological record often interpreted in terms of *proto plate-tectonics* or *vertical tectonics*. However, heat-pipe cooling could be a more appealing mechanism:

- ▸ Greenstone belts (Pilbara, Australia ; Barberton, South Africa) globally devoid of deformation (except for diapirs), longlasting rapid volcanic resurfacing, downward advection consistent with petrology of TTG plutons.

- ▸ Jack Hills zircons hosting diamonds with surface signature exhumed again by the ascent of TTG plutons (intersection of wet basalt and heat pipe geotherms in the diamonds stability field),

- ▸ abrupt cessation of heat-pipe volcanism coincides with onset of plate tectonics in numerical models.
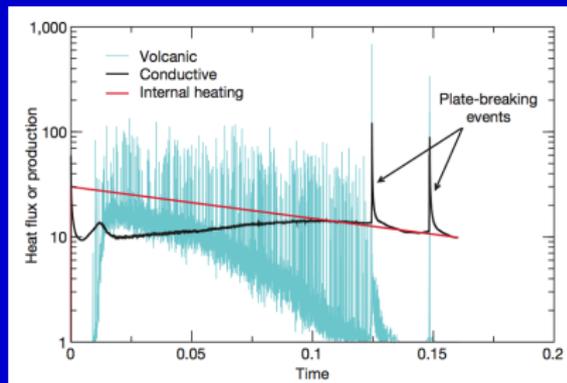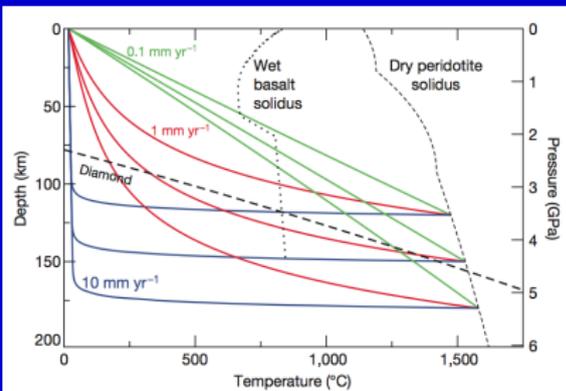
Heat pipe mechanism, Io *(Monnereau and Dubuffet, 2002)*
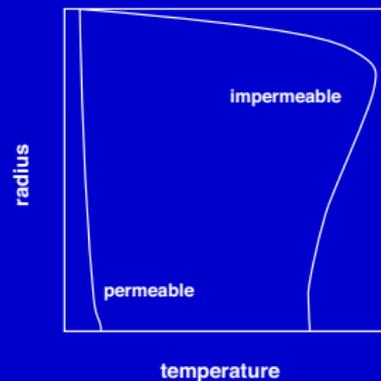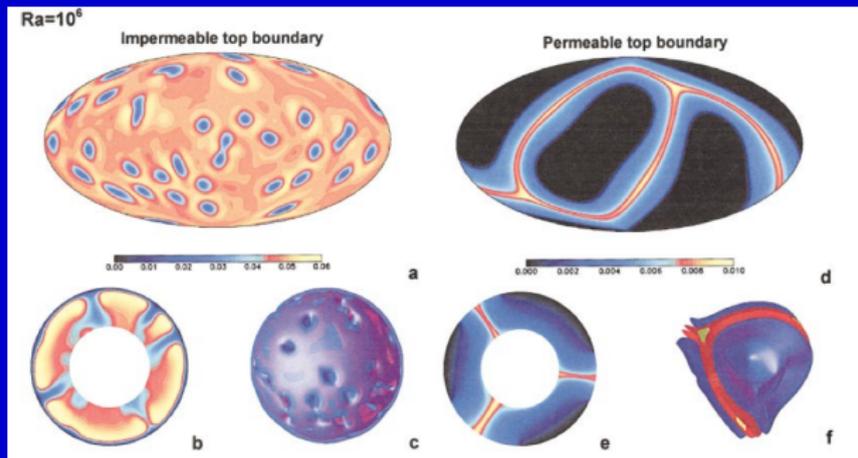
Boundary conditions

Monnereau and Dubuffet, 2002:

planet's surface
model boundary
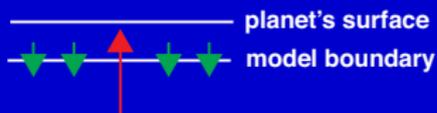
To mimic the heat pipe effect, a permeable boundary is introduced at the surface of the model (i.e. slightly below the actual planetary surface):

- $\tau_{rr} = 0$, $V_h = 0$,
- $\partial T / \partial r = 0$, with an upwind-like scheme:
  (cold) outside T prescribed when $V_r > 0$,
  (hot) inside T prescribed when $V_r < 0$.

Moore and Webb, 2013:

instantaneous
melt extraction

arbitrary
downward flux
solidus
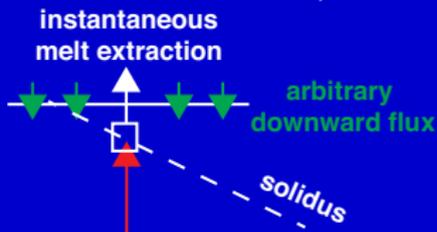
To mimic the heat pipe effect, downward advection with velocity v is prescribed:

- $\partial^2 T / \partial z^2 = v \partial T / \partial z - \tilde{h}$,
- melt is extracted instantaneously.

Convection in hot planets *(Ricard et al., 2014)*

A more general framework can be introduced with a free surface condiction subjected to erosion/sedimentation as well as magmatic cooling:
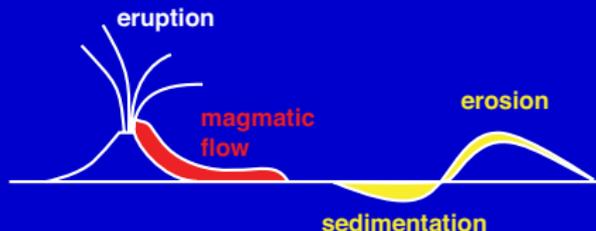
- ▸ free-surface formulation with a diffusing topography h:
  $\sigma_{zz} = -\rho g h$, $\tau_{xz} = \tau_{yz} = 0$, Eulerian vertical velocity $v_h = \nabla \cdot (D \nabla h)$,
- ▸ diffusivity $D = D_e + D_m$,
  $D_e = 10^{-6} - 10^{-3}$ m^2.s^{-1} (empirically determined),
  $D_m = 1 - 10^9$ m^2.s^{-1} (depending on magma viscosity),
- ▸ a secondary Rayleigh number is introduced for topography: $R = \rho g H^3 / \eta D$ (topographic resistance),
- ▸ this formulation continuously spans surface conditions ranging from a closed $(R \rightarrow \infty)$ to an open $(R \rightarrow 0)$ system.

Convection in hot planets *(Ricard et al., 2014)*

$Ra = 10^6$

$R = 10^5$

$R = 10^4$

$R = 10^3$

$R = 0$
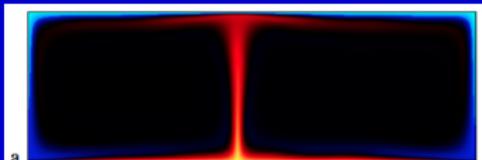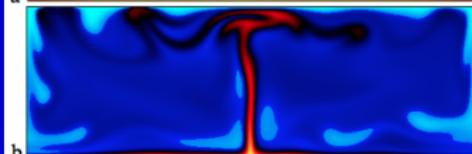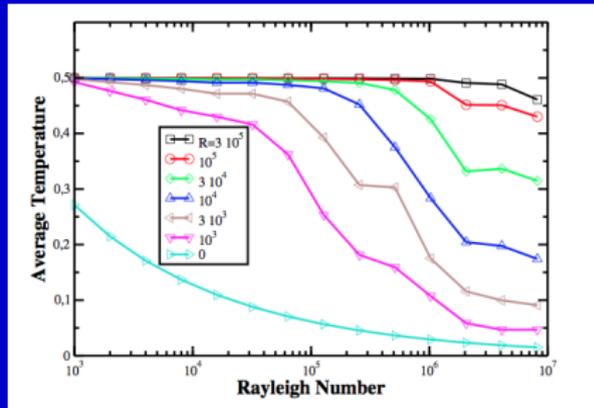
- resistance R controls the level of surface opening,
- a threshold value of $\sim Ra/10 - Ra/100$ is identified.

Phase change involved in major planetary boundaries

for each of the main planetary constituents, both a liquid and a solid phase may coexist

- ‣ solid and liquid metals in planetary cores,
- ‣ magma and rocks in planetary mantles,
- ‣ ices and liquid water in planetary hydrospheres.

Interface subjected to advection and phase change (Deguen, Alboussière)

- continuity of normal stress
(convective stresses equal weight of topography h):
$$-p + 2\eta \frac{\partial v_r}{\partial r} = \Delta\rho g h$$

- velocity of phase change V_r equal to radial velocity in solid v_r
(if h varies slowly):
$$V_r = -\frac{h}{\tau_\phi} - A\left(\frac{q_T^s}{\rho_l c_{p,l}} - q_\chi^s\right) \simeq v_r$$
where τ_ϕ is a characteristic time scale for the phase change
(involving heat transfer in the liquid)and $q_\bullet^s$ are fluxes (of heat and
solute) on the solid side.

Interface subjected to advection and phase change (Deguen, Alboussière)

both expressions combined lead to (dimensionless equation):

$$-p + 2\frac{\partial v_r}{\partial r} = \mathcal{P}\left[v_r - \frac{1}{S}\left(q_T^s + \frac{\kappa_\chi \Delta c}{\kappa_T \Delta T}q_\chi^s\right)\right]$$

where

$$\mathcal{P} = \frac{\Delta \rho g R \tau_\Phi}{\eta}$$

is the ratio between characteristic time scale for phase change (τ_ϕ) and the viscous time scale, indicative of the amount of permeability.

Translation of Earth's inner core

(Deguen, 2013)
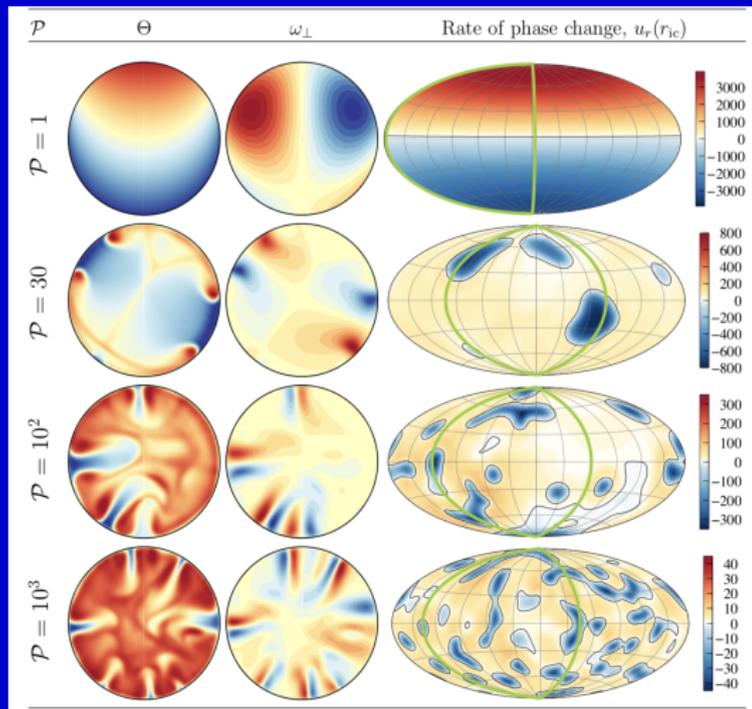
- for large viscosities (low $\mathcal{P}$) values, convective translation of the inner core is the most probable outcome, explaining paradoxical stratified layer observed by seismology at the inner-core boundary.

Main effects of a permeable boundary on thermal convection:

- ‣ the onset of convection possibly shifted to lower thresholds,

- ‣ convection pattern strongly affected (longer wavelengths promoted),

- ‣ heat transfer enhanced and cool layers with hot pipes.
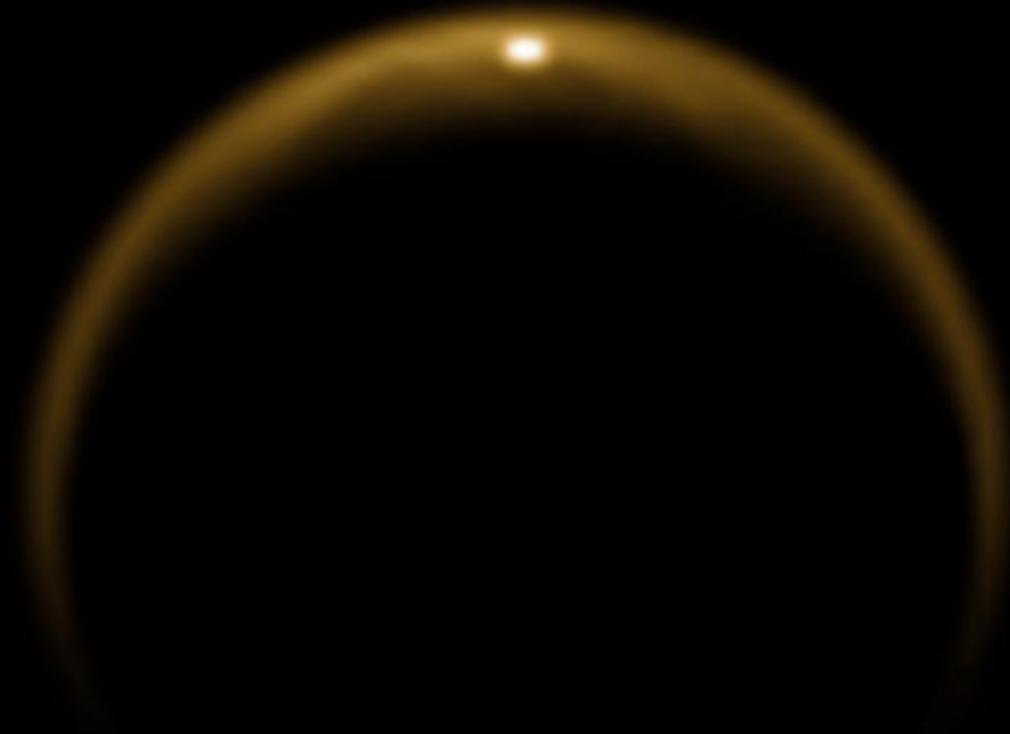
Heat and mass transfer in high-pressure ice mantles

Water phase diagram *(Choukroun and Grasset, 2007)*
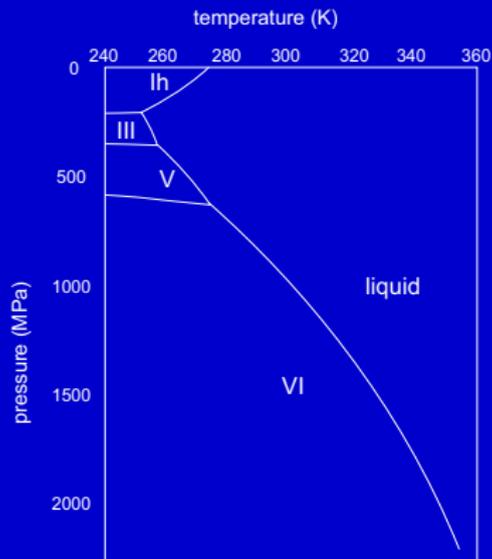
ice polymorphs present in Solar System's moons:

- ‣ hexagonal ice Ih
- ‣ high-pressure ices: II,III,V,VI

Water phase diagram

depth of water/rock contact:

moon	$R_s - R_{w/r}$ (km)	g (m/s^2)	$P_{w/r}$ (MPa)
Enceladus	50	0.11	10
Europa	100	1.31	130
Titan	550	1.35	1100
Ganymede	800	1.43	1600
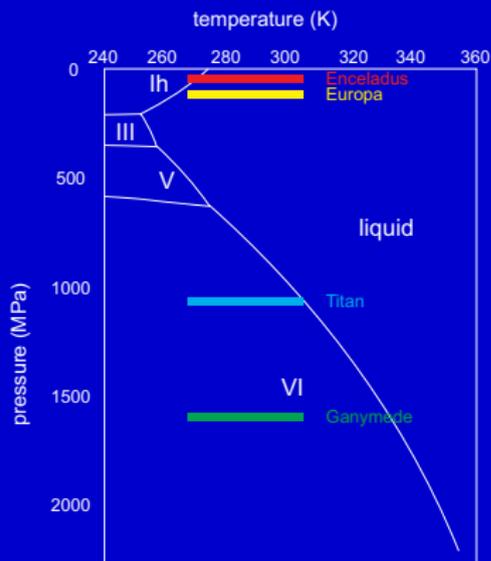
if ocean present beneath an icy crust, isentropic temperature profiles within ocean imply

▸ a direct contact with silicate core for Enceladus and Europa,
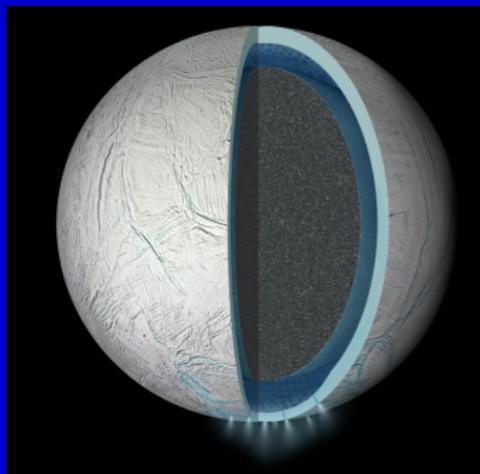
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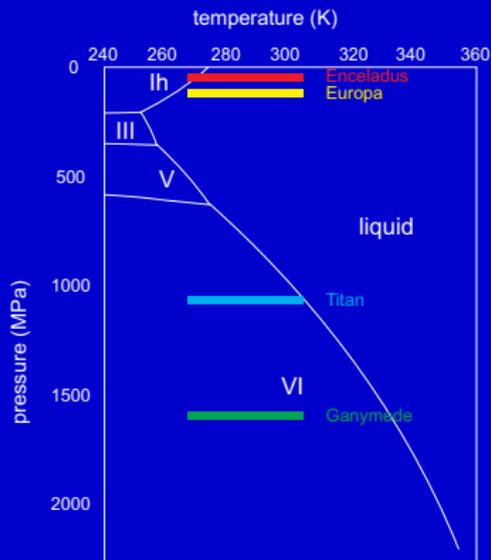
if ocean present beneath an icy crust, isentropic temperature profiles within ocean imply

▸ a solid HP ice (V or VI)/silicate core contact for Titan/Ganymede (not always distinguished in this presentation).
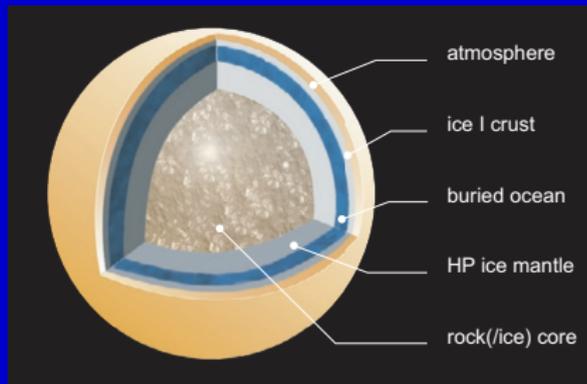
An ocean sandwiched between ice Ih and HP ices
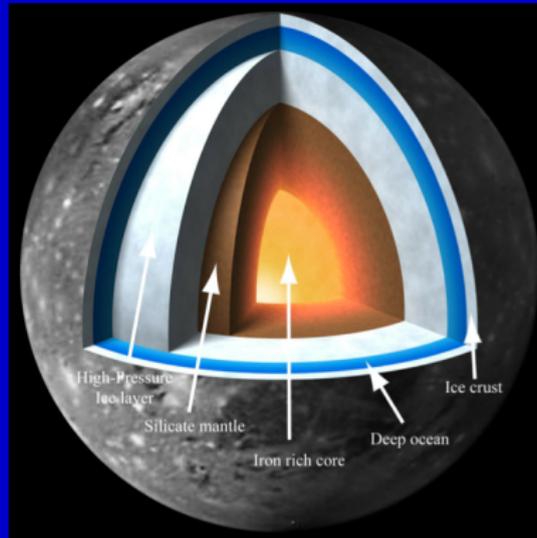
- broad questions on such habitats:
  ⟹ do such environments provide favorable conditions for deep habitats ?
  ⟹ such questions partly addressed by *JUICE*: scrutinize Ganymede as an emblematic example of this class of bodies.

An ocean sandwiched between ice Ih and HP ices
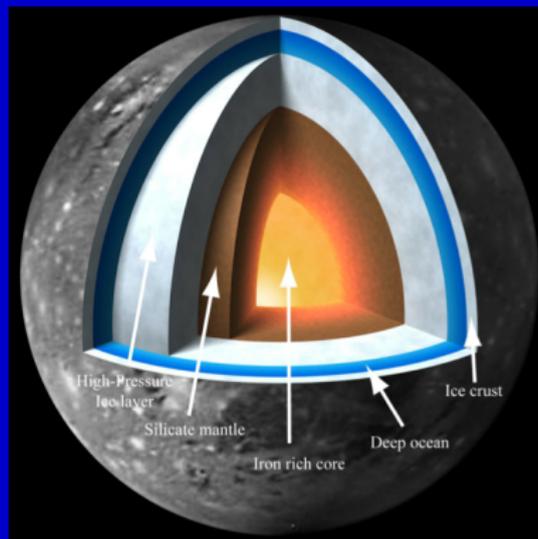
- ▸ more specific questions on dynamics and transport in the HP icy mantle:
  ⇒ what is the thermal state of the HP icy mantle ?
  ⇒ what is the vigour of convection in the HP icy mantle ?
  ⇒ what is the degree of chemical exchanges between the rock core and the ocean through the HP icy mantle ?

Surface evidence for mass/chemical exchanges ?

Titan:

- detection of radiogenic argon by *Huygens (Niemann et al., 2010)*
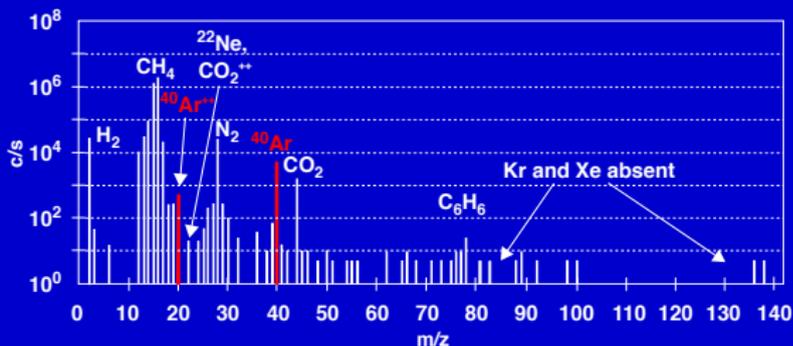
⇒ suggests geological activity in the rocky core as well as chemical transport up to the surface,

- evidence for a salty subsurface ocean from tides, obliquity and electric perturbations
  (Mitri et al., 2014 ; Baland et al., 2014 ; Beghin et al., 2012).

Ganymede

- evidence for a salty subsurface ocean from small oscillations of auroral ovals *(Saur et al., 2015).*

Numerical model for HP ice mantle (I)
(3D, simplified melt extraction, large viscosity $\sim 10^{16} - 10^{17}$ Pa s)

solid state convection

- extended Boussinesq approximation, spherical geometry,
- interface with ocean: free-slip mechanical boundary, temperature prescribed,
- interface with rocky core: no-slip mechanical boundary, heat flux prescribed,
- viscosity is temperature and pressure dependent,
- any melt is extracted instantaneously.
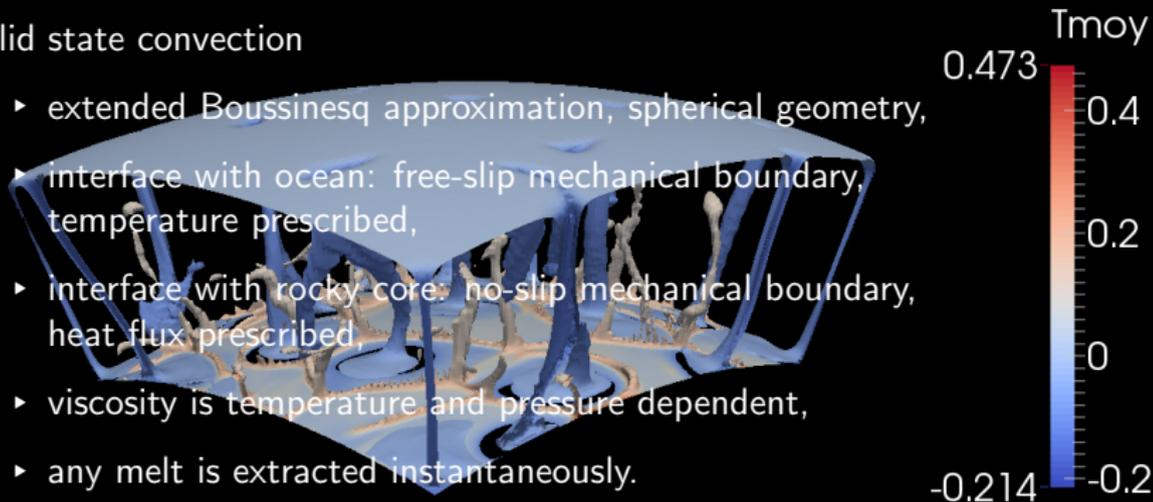
Choblet et al., *Icarus* (2017)

Effect of melting

melt not taken into account (case 1, Durham activation volume)
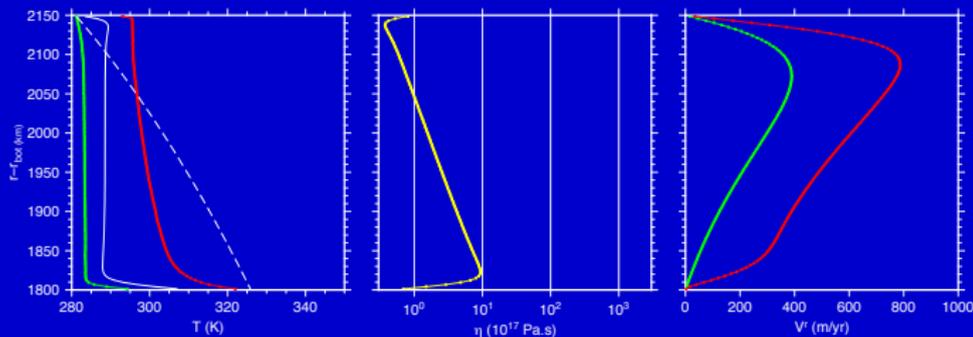
melt extracted (case 1, Durham activation volume)
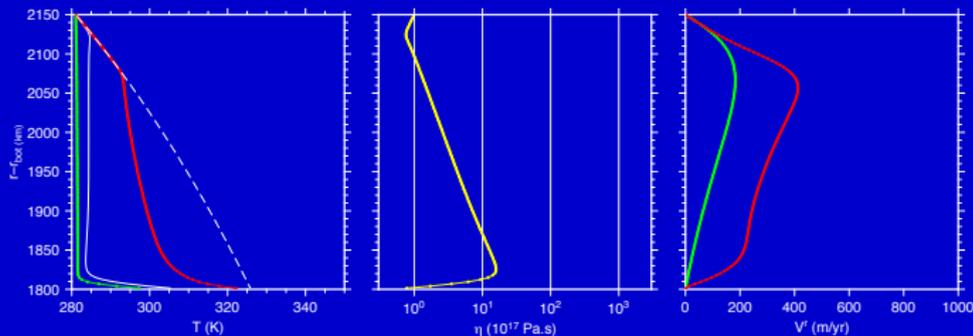
$\Rightarrow$ a slightly smaller ($\sim$ 5 K) temperature, velocities are reduced.

Effect of viscosity

case 1, Durham activation volume
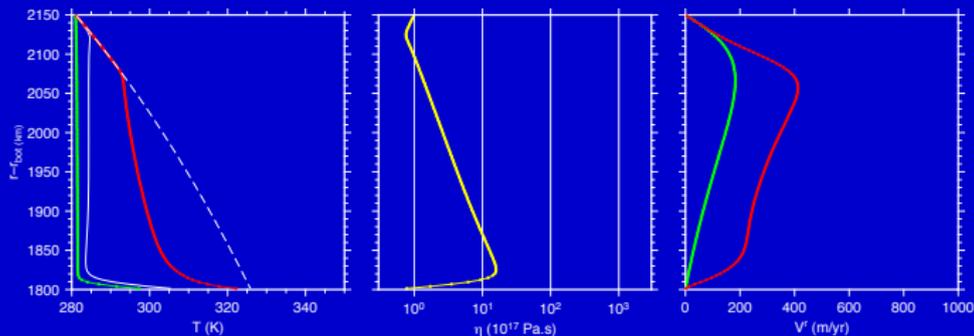
case 1, homologous temperature
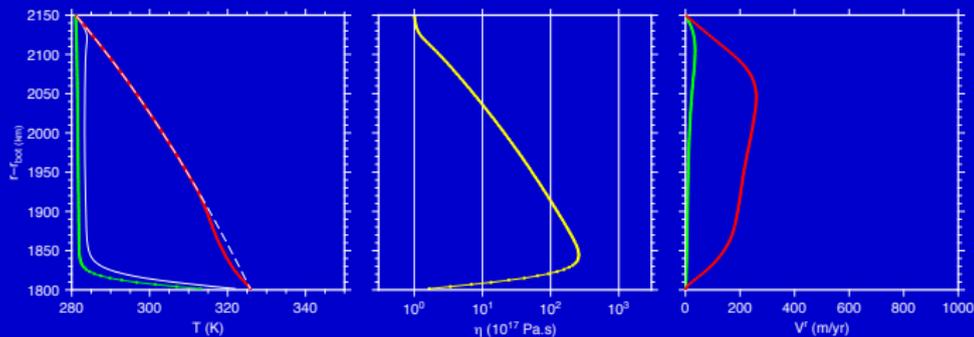
$\Rightarrow$ viscosity increase much more pronounced, maximal temperature much closer to melting curve.

Results: summary

‣ power from silicate core mainly transported by melt extraction,

‣ average temperature always cool ($\sim$ 5 K above surface temperature), mostly depends on layer thickness,

‣ melt occurs at HP mantle interface with rocky core,

‣ however, maximal temperature at all depths close to melting temperature, often leading to interconnection of melt path, unless HP mantle very thick or viscosity very low
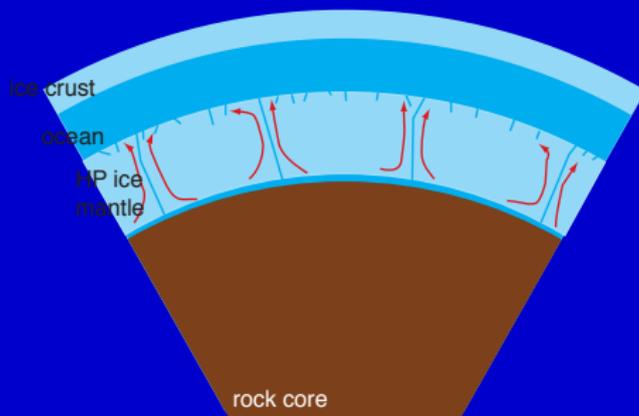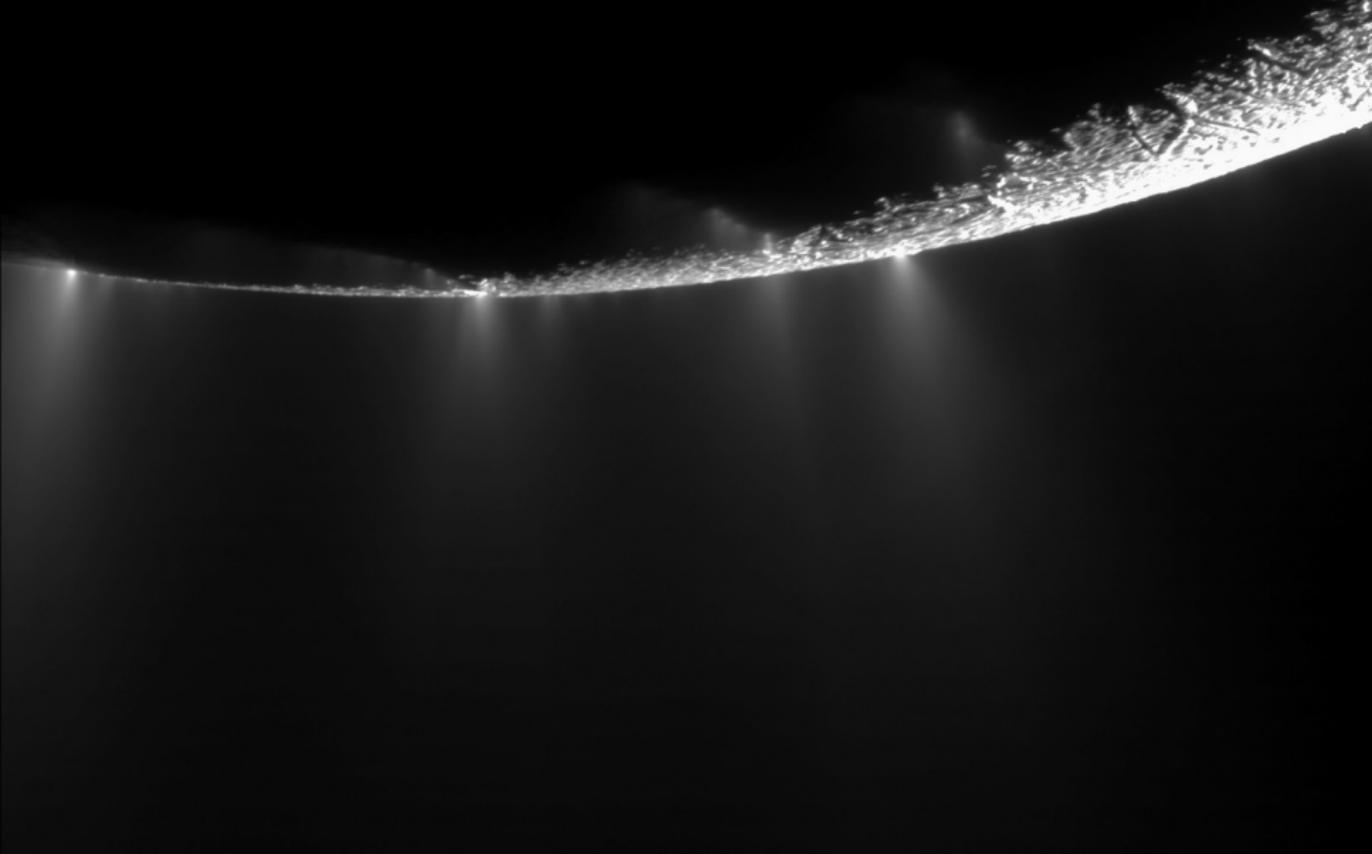
Convection in Enceladus porous core

Enceladus in the Saturn system
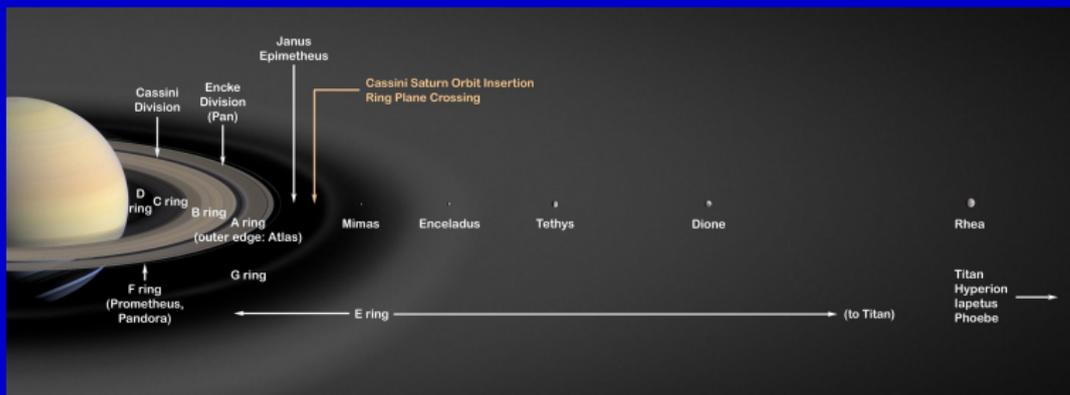
- Saturn's (♄) sixth moon, $R_s = 252$ km, $T \simeq 33$ h, $e \simeq 5 \times 10^{-3}$

- embedded in the densest part of Saturn's diffuse E-ring,

- *Voyager 2*: contrast between relatively young regions near equator and older, high latitude regions, very much unlike Mimas' ancient cratered surface,

$\Rightarrow$ is Enceladus the source of E-ring's material ? (Terrile and Cook, 1981)

Enceladus in the Saturn system
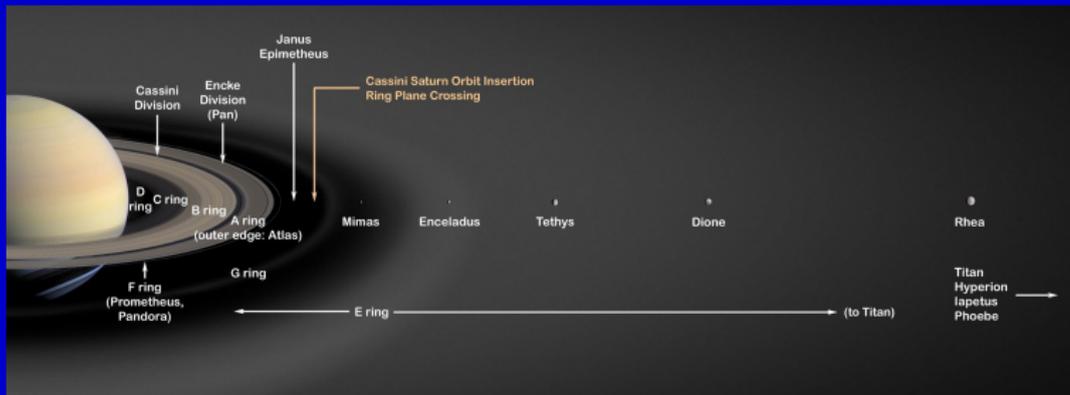
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$\Rightarrow$ is Enceladus the source of E-ring's material ? (Terrile and Cook, 1981)
yes

Voyager 2, April 26, 1981
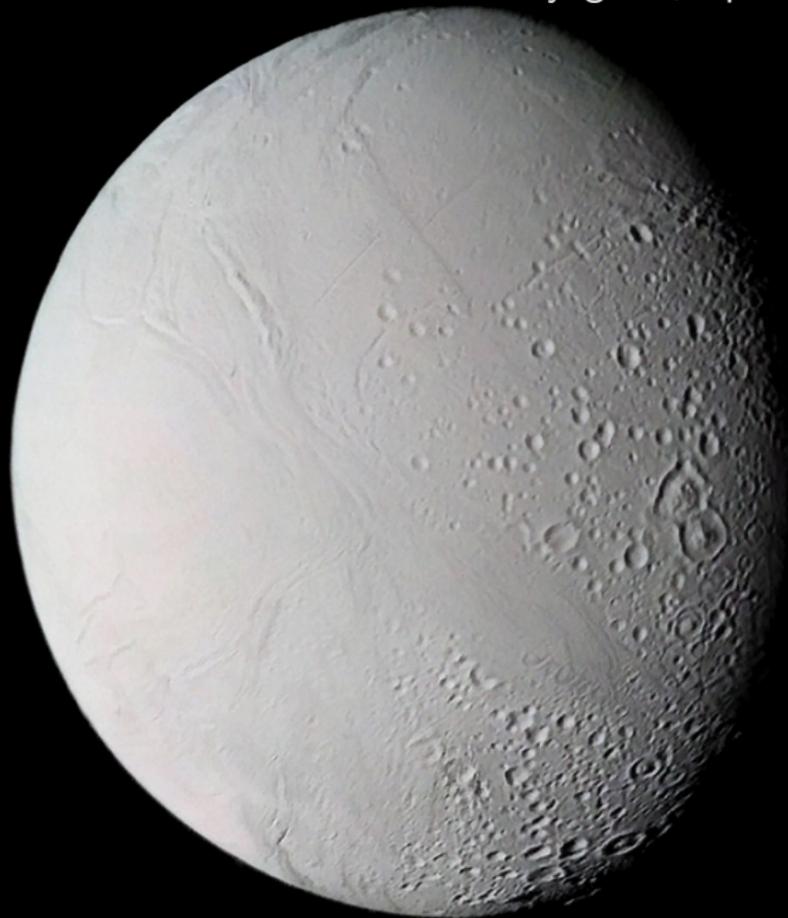
Cassini, January 16, 2005

Cassini, Feb. 2, 2010
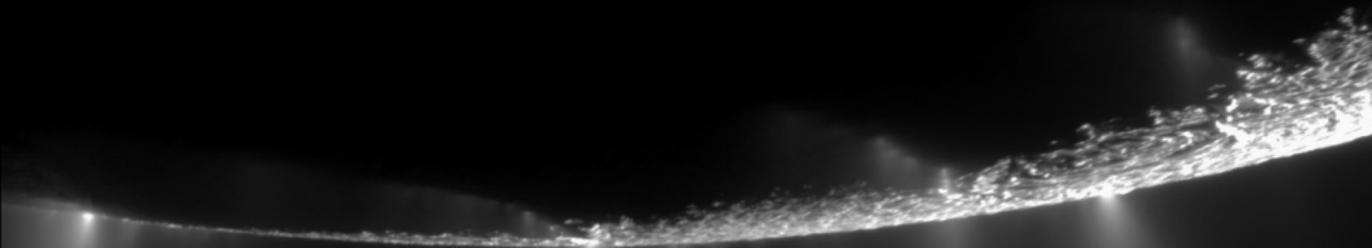
Cassini, July 14, 2005

80 79 80 81 91 87 78 74 78 74

What mechanisms power this intense activity ?

Enceladus' interior at the end of Cassini: structure
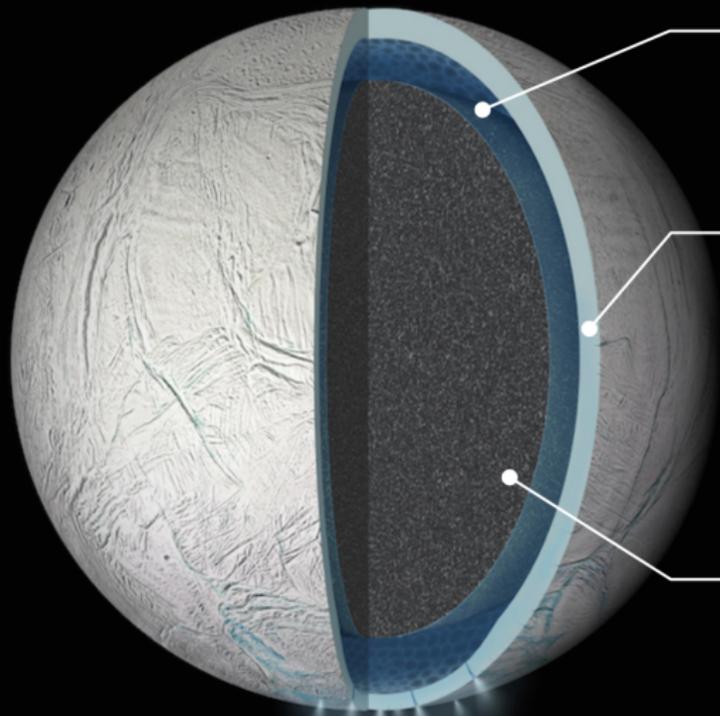
buried global ocean

30-40 km thick
(Thomas et al., 2016)

very uneven ice shell

25-30 km thick in average
up to 35 km at equator
less than 5 km at south pole
(Čadek et al., 2016 ;
Beuthe et al., 2016)

porous rock core

filled with 20-30 % water
(e.g. Roberts, 2015 ;
Waite et al., 2017)

Enceladus' interior at the end of Cassini: composition

ocean is salty

Na (and K) salts in E-ring
(Postberg et al., 2009)

salts and organics in plume
(Postberg et al., 2011)

deep hydrothermalism

nano-silica particles
in Saturn's environment
$\Rightarrow$ 90°C, > 40 km deep
(Hsu et al., 2015)

H_2 in the plume
$\Rightarrow$ hydrothermal reactions
(Waite et al., 2017)

Revisiting the heat budget

Ice shell thickness
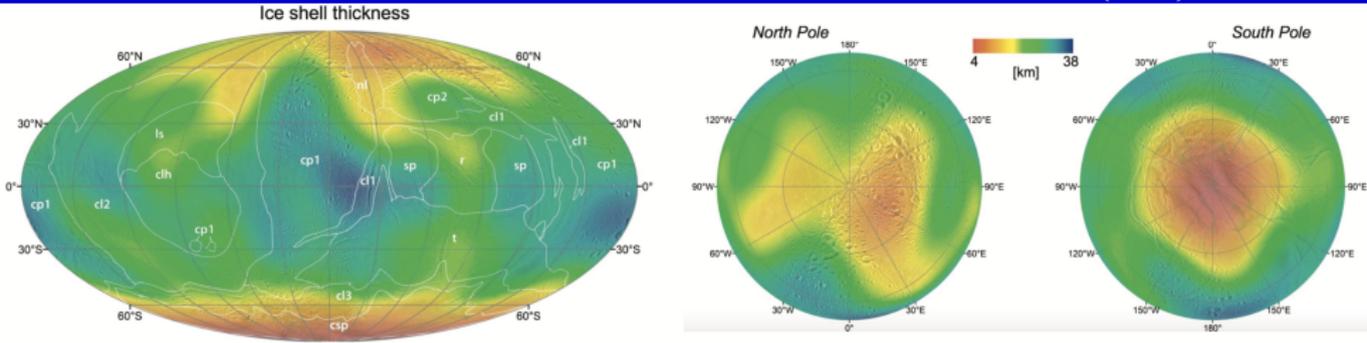
Diffusive equilibrium in a thin ice shell

- ‣ South Polar Terrain (SPT): heat flux $\geqslant$100-150 mW m^{-2}
  or a total power of 3-5 GW

- ‣ heat flux at low latitudes: $\sim$20-40 mW m^{-2}

- ‣ global heat loss outside the SPT: $\sim$20-25 GW

$\Rightarrow$ Total power lost by diffusion $\sim$ 23-30 GW
What/where are the heat sources ?

Tidal dissipation within an ice shell of varying thickness

- thinning at the South Pole strongly enhances tidal dissipation by viscous friction,

- optimal heat production obtained for ice shell thickness ranging between 1 and 3.5 km for ice viscosity between 10^{14} and 10^{13} Pa s,

- tidal friction further enhanced by the presence of faults,
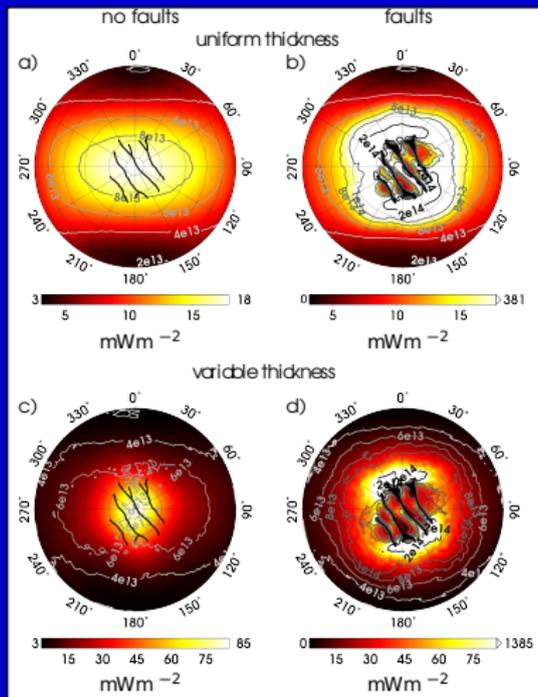
Běhounková et al. (2017)

$\Rightarrow$ Dissipation large enough to counterbalance heat loss in the SPT, but not at moderate/northern latitudes.

Tidal heat production in Enceladus' deep interior (1): the ocean
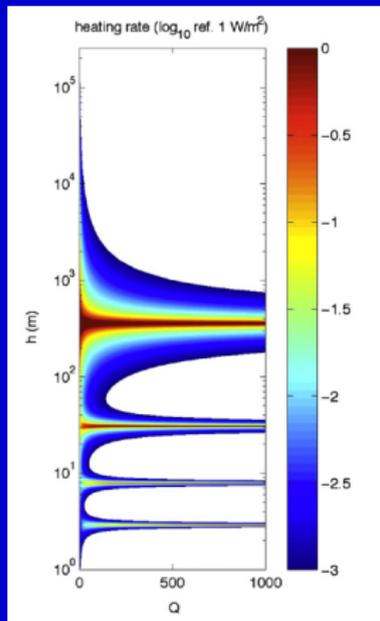
Tyler (2010)

‣ dissipation of resonant waves as a response of the subsurface ocean to both eccentricity and obliquity components of tidal forcing (Tyler, 2010 ; Matsuyama et al., 2018),

‣ significant heating obtained for very thin ocean and "inconsistent" pattern (Matsuyama et al., 2018),

⇒ overall, unlikely to generate 20-25 GW in the present geometry. . .

Tidal heat production in Enceladus' deep interior (2): the core

- due to low central pressure, Enceladus' core is likely unconsolidated,

- first gravity measurements (Iess et al., 2014) yield $\rho_{core} \simeq 2.4$ g cm^{-3} → porosity could be as large as 20-25 %,

- porosity in excess to 20, % weakens the core with ice/water controlling the deformation,

- at present, a few GW could be generated by viscous dissipation in the core filled with ice.

Roberts (2015)

⇒ what power could be produced by dissipation in a core filled with liquid water ?

Dissipation of a water saturated mixture of sand/gravel
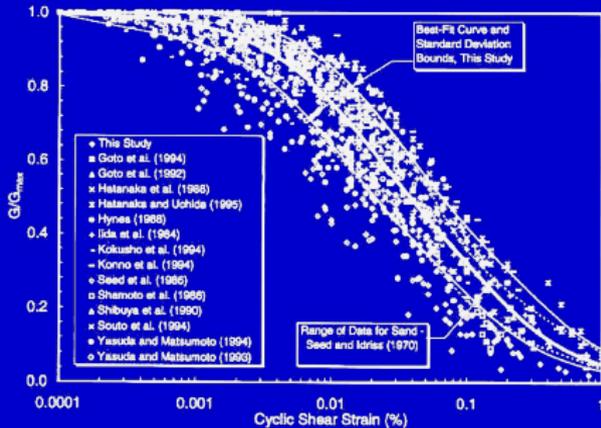
- possible mechanisms: inter-granular friction during grain rearrangements or/and frictional sliding along microcracks,

- anelastic properties of such materials classically parameterized with effective shear modulus and the damping ratio (or dissipation function),

- a strong decrease in elastic modulus (and increase in dissipation function) is expected when cyclic strain exceeds ∼ 0.01 % (typical value for Enceladus tidal deformation in the core, see after),

- laboratory mechanical tests performed however at larger frequency and lower pressure than those prevailing in Enceladus' core: we expect enhanced dissipation at the very low tidal frequency (creep effects) and a moderate effect of pressure.

Tidal dissipation in Enceladus porous core

(A) P_{tide} (GW)

(B)

(C) $q_{\text{tide}}/q_{\text{max}}$
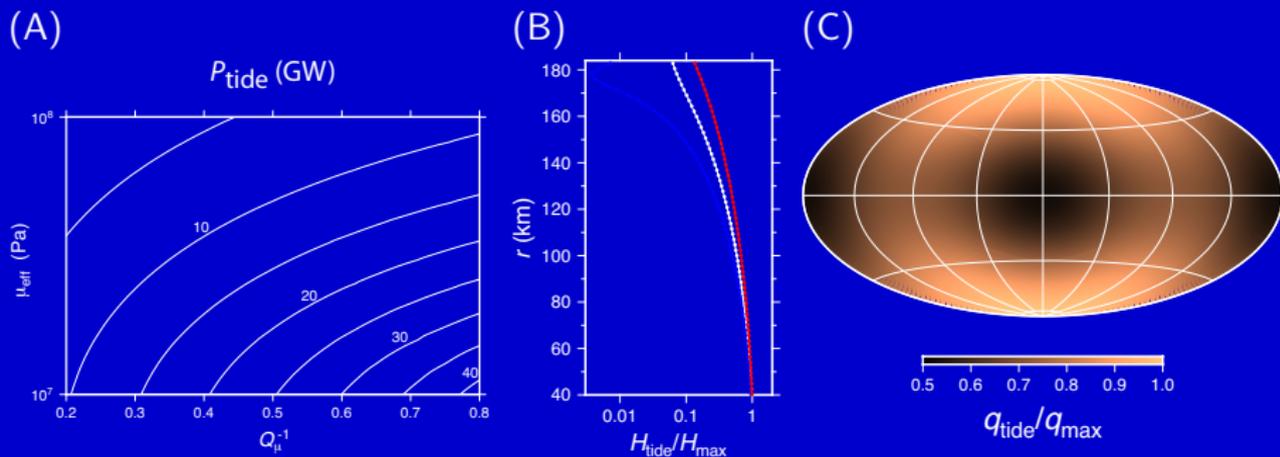
$\Rightarrow$ several 10s of GW can be produced with a slightly heterogeneous diffuse pattern: heating is maximal and homogeneous near the centre and decreases more slowly at the poles towards the surface
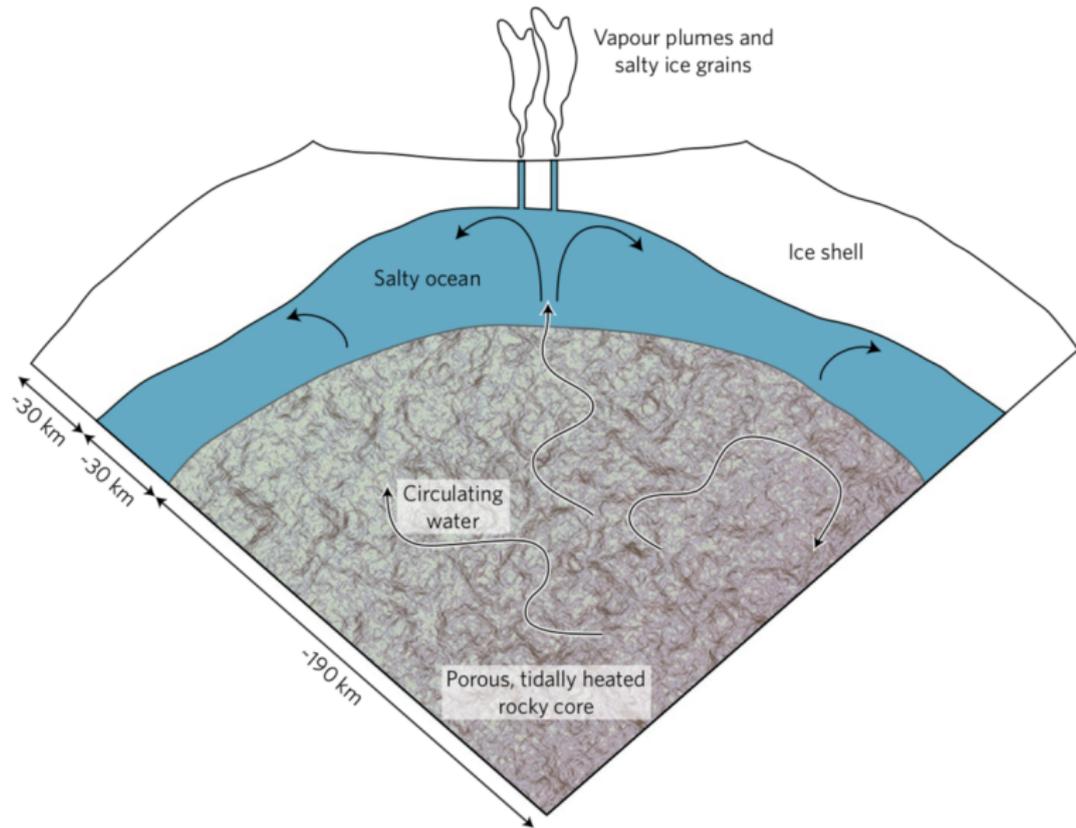
Fig. 1 | Sketch of the interior of Enceladus. Choblet et al. propose a tidally heated core driving hydrothermal circulation[6]. The background internal structure is based on the model presented in ref. [13].

Thermal convection of insterstitial water in a porous core

Conservation of mass:

$$\nabla \cdot \mathbf{V} = 0 \qquad (1)$$

Darcy equation:

$$\mathbf{V} = -\frac{K}{\eta_w \phi} \left(\nabla p_w + \frac{4\pi}{3} \rho_b \rho_w Gr \mathbf{e^r} \right) \qquad (2)$$

Conservation of energy:

$$C_b \frac{\partial T}{\partial t} + \phi \rho_w c_w \mathbf{V} \cdot \nabla T = k_b \nabla^2 T + H_{\text{tide}} \qquad (3)$$

Eqs (1) and (Eq. 2) can be combined to produce a pressure equation solved in our approach:

$$\nabla \cdot \left(\frac{1}{\eta_w} \left(\nabla p_w \right) \right) = -\gamma \nabla \cdot \left(\frac{\rho_w}{\eta_w} r \mathbf{e^r} \right) \qquad (4)$$

Porous convection with homogeneous heating
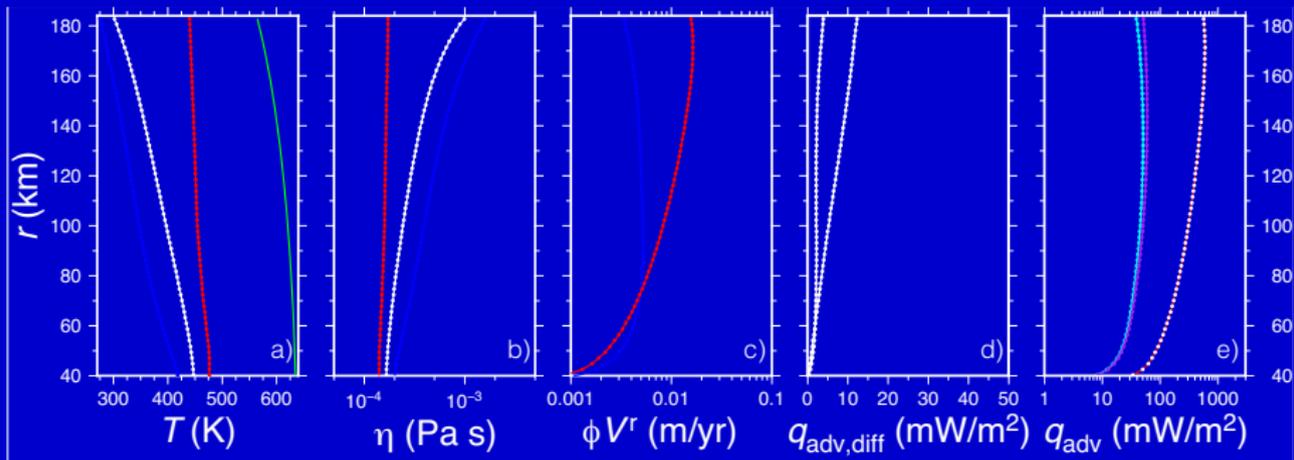
$\Rightarrow$ gradual heating of passive downwellings and buoyant rising in hot narrow upwellings)

Porous convection with homogeneous heating
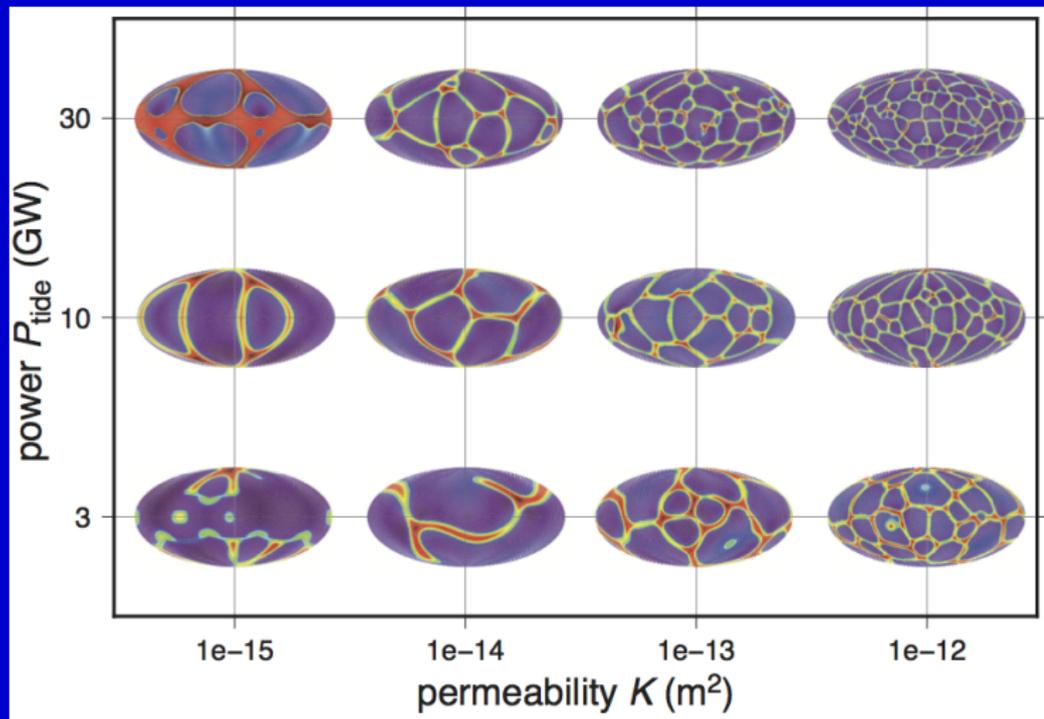
$\Rightarrow$ heat released by narrow, sheet-like upwellings of hot water ($>100^{\circ}C$)

Porous convection with heterogeneous (tidal) heating
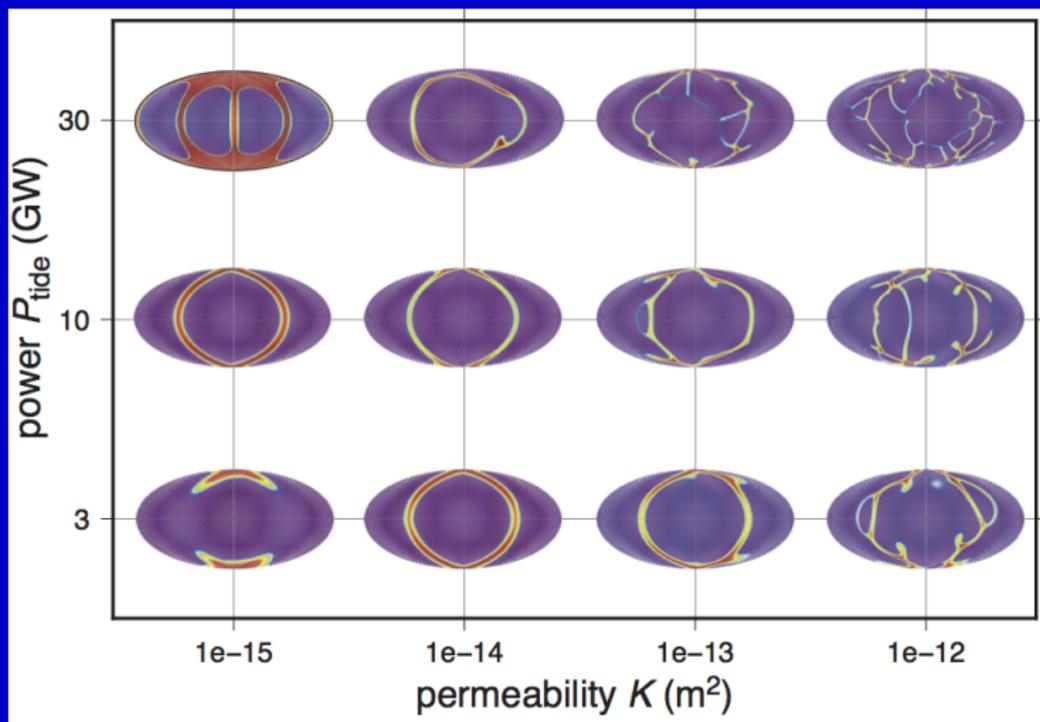
⇒ upwellings concentrated at the poles and trailing/leading meridians
where maximal dissipation occurs.

Hot spots at the seafloor
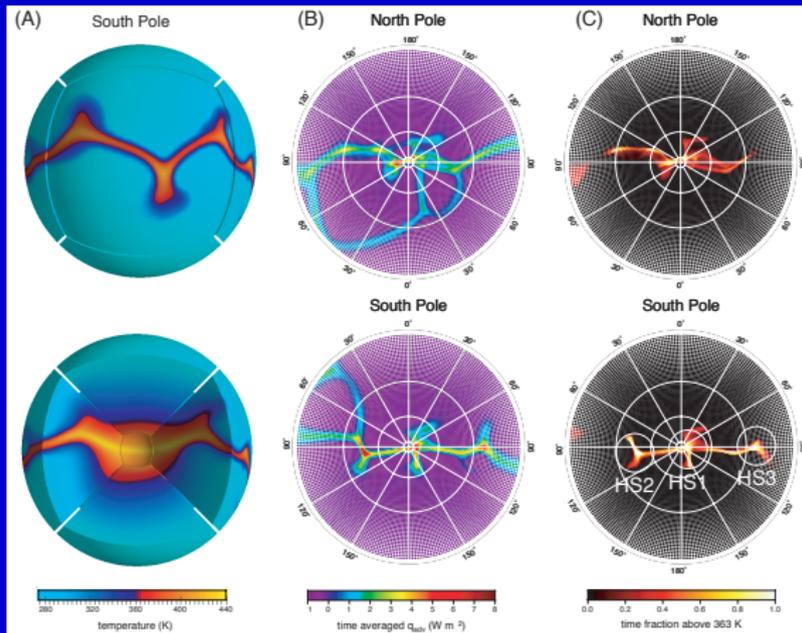
$\Rightarrow$ persistent hotspots (1-5 GW) are anchored beneath the poles.

Summary

porous interface
between the ocean
and the rock core

upwelling of interstitial
hot water (> 90°C)

seafloor hotspot
hot ocean plume

melting at the ice/ocean interface
jet activity
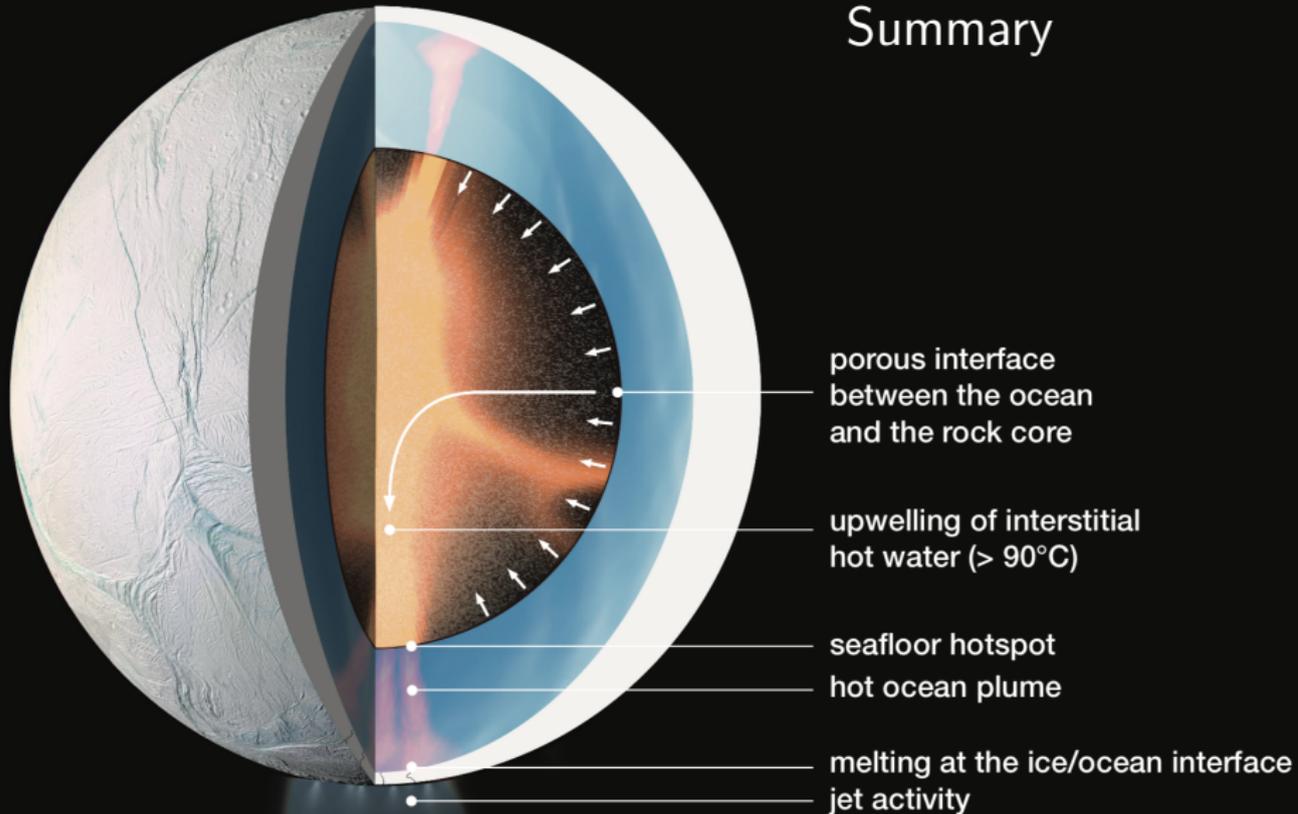
Cassini, Sept. 15, 2006

Porous convection of liquid water in the unconsolidated core

Water properties: density, viscosity
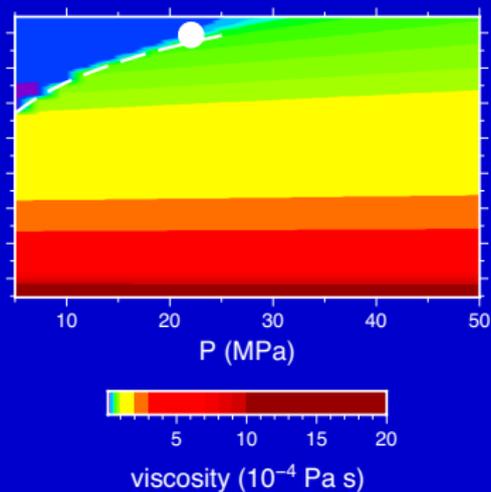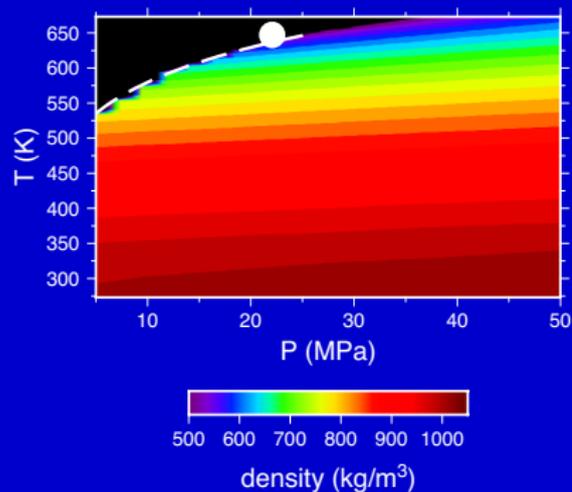
Porous convection of liquid water in the unconsolidated core

Water properties: density, viscosity

- effect of pressure (in the range [10, 40] MPa) is neglected
- critical point never reached in Enceladus' core
- boiling might occur (not treated consistently in our models)
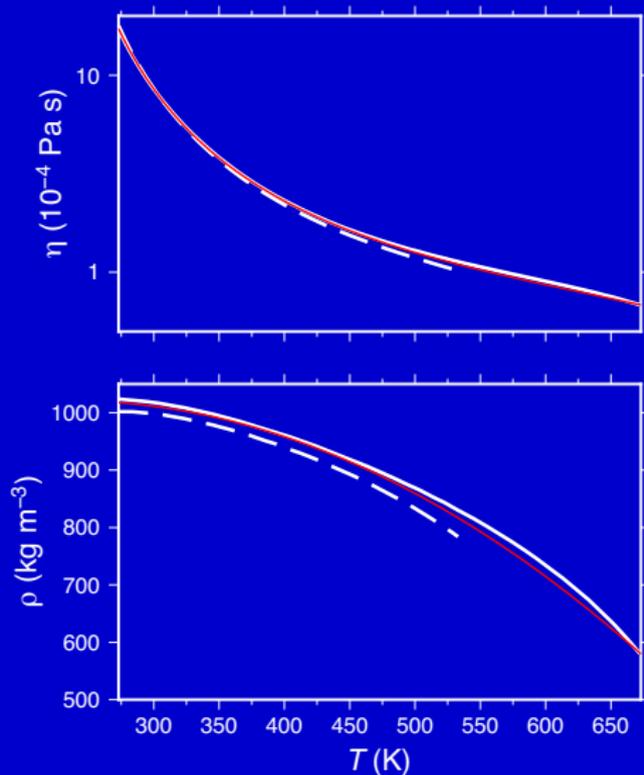
Porous convection of liquid water in the unconsolidated core

Permeability is the key (yet unknown) variable for porous flow convection
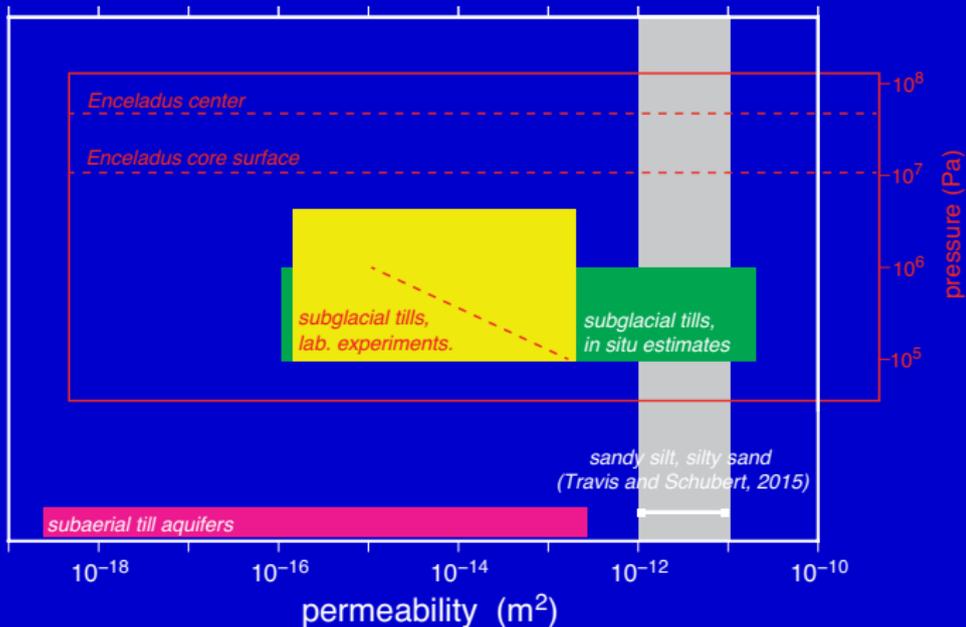
Synthesis of results

a trade off is observed between water temperature and velocity:

- ‣ low K: high T, lower V,
- ‣ high K: low T, higher V,
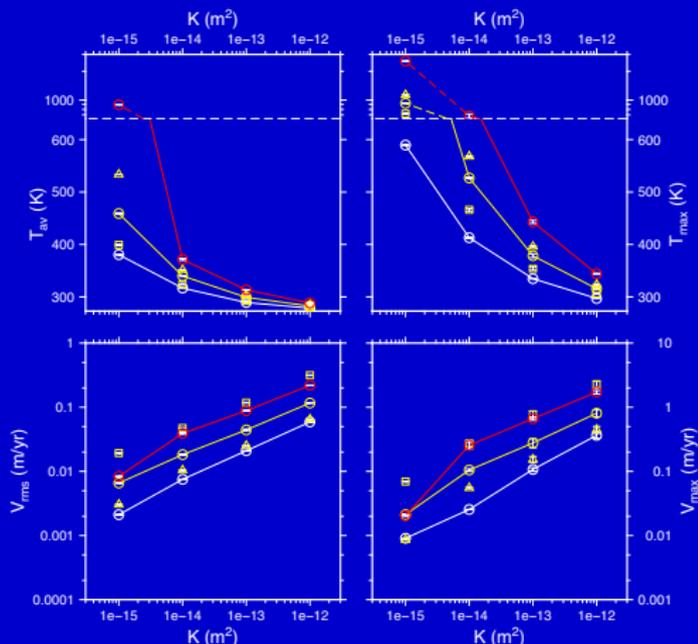
$\Rightarrow$ while the range of admissible values for permeability K is extremely large ($[10^{-18}, 10^{-11}]$) m^2, cf. terrestrial analogues), only a limited interval ($[10^{-14}, 10^{-13}]$ m^2) is consistent with findings on hydrothermal activity (cf. Hsu et al., 2015). . .

Synthesis of results
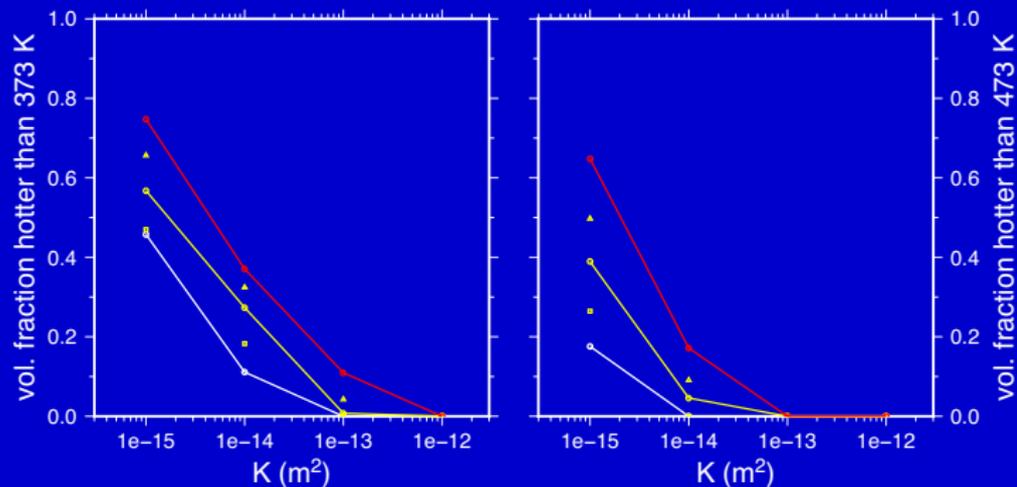
$\Rightarrow$ a significant volume fraction is above $>100°C$ implying efficient aqueous alteration

The very uneven ice shell as a witness

- refined analysis (based on shape up to degree $\ell=18$, Tajeddine et al., 2017) confirms earlier results for ice shell structure:

  - minimum thickness under polar regions (esp. at the South)

  - ice shell thinner along leading/trailing meridians at lower latitudes,

- predicted heat flux variations ressemble the pattern emitted at the sea floor,

a) Ice shell thickness

b) Heat flux on top of the ocean
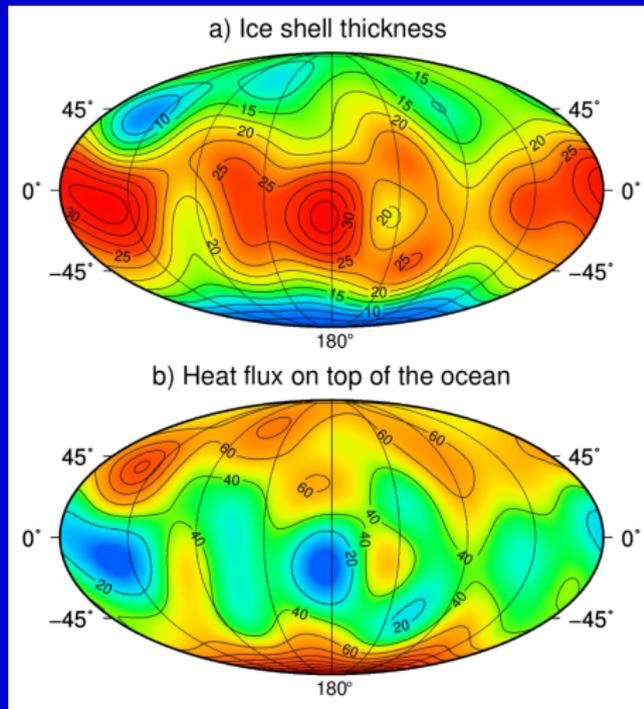
Polar ocean plumes: scaling results

	Goodman et al (2004) (scaling and laboratory experiments)	Goodman and Lenferink (2012) (numerical experiments)
l_{cone}/H	0.15	0.2
$V_p/(Hf')$	0.01	3×10^{-3}
$(\Delta T_p g_o \alpha_o)/(Hf'^2)$	3×10^{-4}	4×10^{-5}
l_{cone} (km)	8.25	11
V_p (cm/s)	6	2
ΔT_p (mK)	5	0.7

Table: Characteristics of the oceanic thermal vents.

Temperature:
0.14 mK

21 Europa days

40-60 km
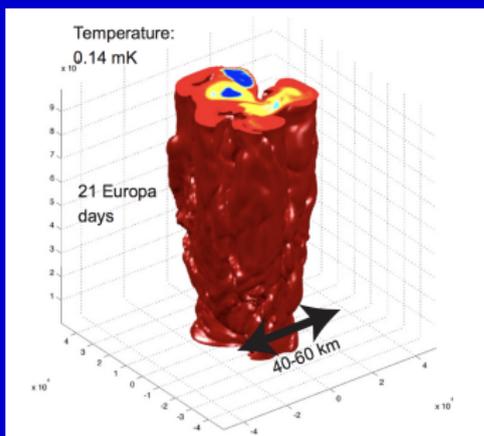
- temperature variations within the ocean of a few mK,
- estimated transport time from seafloor to the source of jets beneath SPT ice shell: a few weeks to months,
- compatible with the growth rate of nano-silica *(Hsu et al., 2015)*

From the seafloor to the ice/ocean interface: dedicated simulations

- at moderate latitudes bounded by the tangent cylinder, the signal may be blurred depending on the competition between rotation and buoyancy

  (A): Coriolis forces are dominant ⇒ thermal convection largely remains outside the tangent cylinder and is visible at the surface of the ocean

  (B): Archimedes forces are dominant ⇒ lesser effect of the tangent cylinder, convective instabilities deflected to higher latitudes

- in polar regions, ocean plumes always imprint the seafloor pattern at the ice/ocean interface.

prescribed basal heat flux

surface heat flux V^r

(A)

(B)

Temporal perspective

stability of the hydrothermal activity

- the latest estimate of Saturn's dissipation function yields an equilibrium heating of 10- 50 GW (Lainey et al., 2017 ; Fuller et al., 2016)

- hydrothermal activities could be sustained for billions of years for $P_{tide} > 15$ GW and for at least 20 Myr for $P_{tide} < 15$ GW.

mass/chemical flux in the hydrosphere

- the entire ocean would be processed in the core at temperatures higher than 363 K in 25-250 Myr,

- the entire volume of ocean would be exchanged with the ice shell through melting/freezing on the same time scale.

Čadek et al., (2017)

Solid-state flow of planetary constituents

- ▸ constitutive materials for planets:
  metals, rocks (silicates), water,

- ▸ the solid phase is known to possibly flow (creep).

metals: metal rolling (metal stock passed through a pair of rolls to produce a sheet of reduced thickness) probably very ancient (slitting mills around 1600 in Europe).

Solid-state flow of planetary constituents

‣ constitutive materials for planets:
  metals, rocks (silicates), water,

‣ the solid phase is known to possibly flow (creep).

ice: glacier flow observed in the Alps already in 17th century.

Solid-state flow of planetary constituents

- constitutive materials for planets:
  metals, rocks (silicates), water,

- the solid phase is known to possibly flow (creep).

rock: slow rise of Scandinavia above sea-level already observed by
Anders Celsius (around 1730, associated to evaporation), the
concept of post-glacial rebound appeared later on.

Mantle convection and continental drift

‣ once mantle viscosity exhibited from post-glacial rebound, mantle flow on geological time-scales is a possibility,

‣ Holmes (1931): connects solid-state mantle convection to continental drift,

‣ besides minor errors, modern understanding of mantle convection is proposed.

Holmes (1931)

Big vs. small, diversity: the seafloor
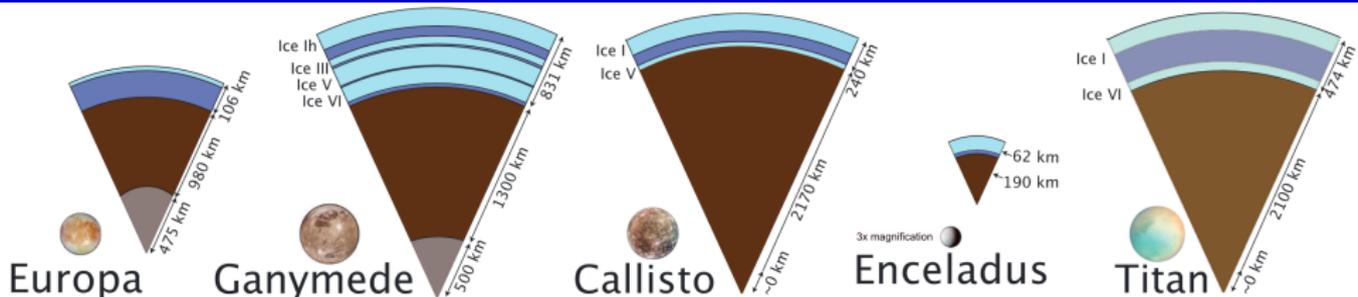
rock/hydrosphere interface:

- ‣ on large moons (Titan, Ganymede) the HP ice mantle is expected to constitute a barrier to chemical interactions between rocks and the ocean,

- ‣ conversely, on smaller moons (Enceladus, Europa) hydrothermal activity might provide a very promising mean for chemical interactions.

Big vs. small, resemblance: the ice ceiling

ocean/ice interface:

‣ while the seafloor has a different nature for the two classes (a phase boundary for large moons vs. a major compositional dichotomy for small moons), lateral gradients in heat flow (and possibly mass flux) could provide a forcing to ocean dynamics ,

‣ ocean/atmosphere interaction not obvious:
$\Rightarrow$ stagnant lid regime inherent to convection with strongly temperature-dependent viscosity has to be broken,
$\Rightarrow$ presence of liquid water in a partially molten ice crust might be episodic owing to an intrinsically larger density...

Heat power out of the rocky core

thermal history of Titan's rocky core *(Tobie et al., 2006)*:
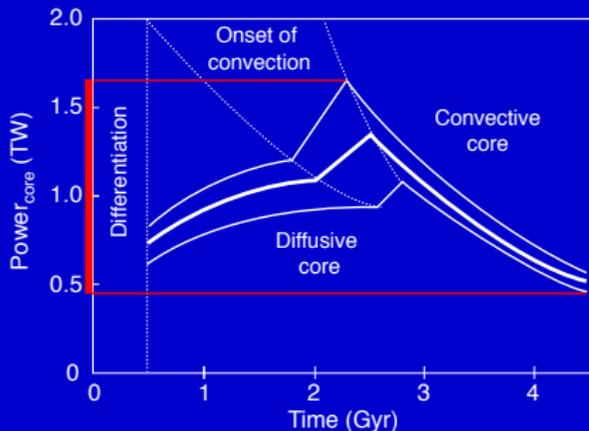
▸ three mineralogical compositions and densities reflecting different hydration states,

▸ initial conductive stage followed by convective cooling
(onset of convection $\sim$ 1.5-2.5 Gyr ago).

$\Rightarrow$ minimal power corresponding to the present-day $\sim$ 0.5 TW,
$\Rightarrow$ maximal power corresponding to the onset of convection > 1.5 TW.

Viscosity of the HP ice

rheology of ice VI measured *(Sotin et al., 1985 ; Durham et al., 1996)* although in limited conditions:
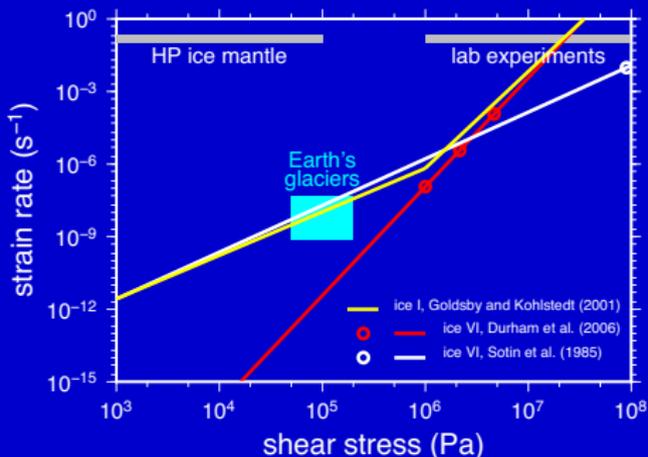
- considerably higher uncertainties than for ice Ih,

- apparently, dislocation creep $n > 1$ prevails, although GSS creep is a possibility,

- probably upper bounds on strength of planetary ice.

- activation volume not measured for $T > 250 K$

$\Rightarrow$ two end-member models for activation volume: either the estimate of Durham et al., 1996 (based on lower T, larger $\dot{\epsilon}$ experiments) or based on the "homologous" temperature formalism,

$\Rightarrow$ reasonable viscosity range near the melting point: $\sim 10^{14} - 10^{17}$ Pa s (low value corresponding to plausible change in creep mechanism at low stress).

Conclusions
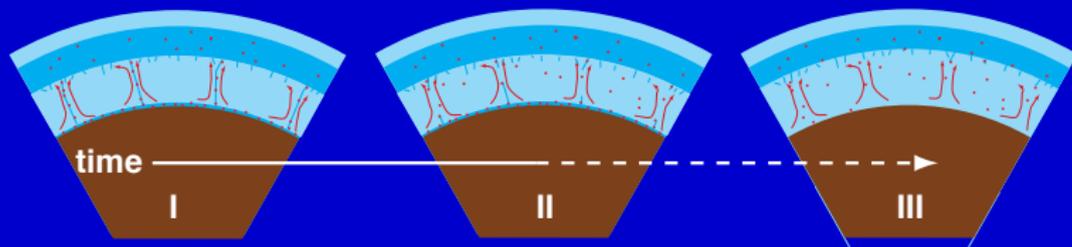
⇒ time evolution and cooling of the moons probably involves successive regimes for melt path,

⇒ due to alleged thinner hydrosphere, Titan more likely than Ganymede to have preserved melt connexion from rocky core to ocean, up to present day (or recent past),

⇒ ultimately, only a full description of heat transfer through HP ices/deep ocean/ice Ih will allow to derive thermal history tailored for each moon.