IS IO'S MANTLE REALLY MOLTEN? M. Monnereau[1] and F. Dubuffet[2], [1]Observatoire Midi-Pyrénées (14 avenue Edouard Belin, 31400 Toulouse, France - Marc.Monnereau@cnes.fr), [2]Minnesota Supercomputing Institute (1200 Washington Avenue South, Minneapolis, MN 55415, USA - fabien@msi.umn.edu).

Introduction: Io is one of the most remarkable terrestrial planets, because its intense volcanic activity contrasts with a size barely larger than our own moon. The tidal deformation, responsible for this activity, dissipates a prodigious amount of heat in Io's interior, several times larger than the heat released by the Earth's mantle dynamics. Infrared observations [1] and predictions of tidal energy dissipation in Io [2] indicate that Io has a large average heat flux, 2.5 W/m^2, amounting 10^{14} W for the whole planet. Such a heat flow suggests a strong internal activity and asks a puzzling question about the heat transfer mechanism and the related thermal state of Io's interior.

If the heat was evacuated close to the surface mainly by conduction, as on the Earth, the thermal gradient just beneath the surface should reach at least 750 K/km, such that a conductive lithosphere should be only 2 km thick. In that case, the sub-lithospheric mantle would be molten and stirred by a highly turbulent convection with a Rayleigh number lying between 10^{12} and 10^{16} [3]. However, this seems in contradiction with the topography observed by the Voyager spacecraft, mountains with relief in excess of 10 km requiring a thicker lithosphere. O'Reilly and Davies [4] point out this argument to discard the thin lithosphere hypothesis and to propose that the heat is advected by magma through isolated vents, what offers the possible existence of an arbitrarily thick and cool lithosphere at the Io's surface. This assumption is well supported by the fact that the power radiated by the hot spots is essentially the total heat flow. We propose to introduce this heat pipe cooling mechanism in mantle convection modelling just changing the classical impermeable top condition for an impermeable one.

Modelling: Melt migration inside the mantle is a shock wave-like process described by highly nonlinear equations which require a huge computational effort to solve. Thus, to introduce heat transport by magma circulation in mantle convection modelling has to be approached in simpler way. In the O'Reilly and Davies' model, the heat is carried out by the magma which rises from depth, then spreads out at the surface and cools. The solid lithosphere subsides under the weight of flows and is heated until its base melts. In that case, the conduction through the lithosphere is strongly limited by its subsidence and the heat is mainly evacuated by advection. This can be easily introduced in numerical models via an open boundary condition at the top surface, like in hydrothermalism modelling. Of course, as mentioned above, the two phase flow will not be taken into account, and the heat will not be advected by magma, but by solid viscous mantle. This leads to a permeable condition (the normal vertical stress vanishes), with a no slip condition to force the fluid velocity to be perpendicular to the surface (horizontal velocities are set to zero). Experiments are performed in spherical geometry, with pure internal heating. The aspect ratio (core radius/surface radius) is 0.52. The calculations have been done on spherical harmonic development up to degree 128 and on 128 regularly spaced radial levels.

The permeable boundary effects: To draw the main differences between closed and open boxes, first we compare experiments performed in spherical geometry with pure internal heating at a Rayleigh number of 10^6 and 10^7 with a bottom free slip condition. Experiments only differ in the top condition: in case of a closed box a free slip and impermeable conditions are set, while permeable and no slip conditions are used for an open box.

The planform: Experiments performed with an impermeable upper condition display an expected point-like downwelling pattern. No hot plume develops because hotter regions remain close to the surface convective boundary layer. In that case, the spectrum of thermal heterogeneities is mainly marked by high energy level localized only at shallow depth and high harmonic degrees. Also, the thermal field remains unsteady. Instead of the this classical pattern, the susbstitution for the impermeable condition by the permeable one provides a very low degree stable pattern. Here we find a convective mode 3 at Ra=10^6 and a mode 4 at Ra=10^7. The cells are delimited by hot planes dividing the mantle in equivalent domains where downwellings establish. The intersection axis of these planes is marked by the presence of hot cylindrical upwellings. In closed boxes, this remarkably stable pattern is only found at low Rayleigh numbers close to the critical threshold. Conversely to the open box, where downwellings, associated to the highest velocities, drive the circulation, in open box, downwellings play the role of return flow, and are, in a way, sucked by strong hot plumes.

The temperature profile: Not only the planform, but also the temperature profile is strongly affected by the change in the top boundary condition. The shape appears quite different. At the surface, the impermeable case profile is marked by a thermal boundary layer needed for the heat transfer. Below, the temperature reduces as depth increases, because of cold material piling up at the bottom surface. In the permeable case, since the heat transfer through the surface is advective, there is no thermal boundary layer at the top surface and the horizontally averaged temperature increases progressively with depth. Also the whole averaged temperature strongly reduces from the impermeable case to the permeable one, $4.5\ 10^{-2}$ against $3.8\ 10^{-3}$ respectively. As it was expected from the porous media studies, the permeable boundaries offers a more efficient cooling effect.

Experiments shows that permeable top boundary provides thermal field and dynamics close to the expected ones, as depicted by the O'Reilly and Davies' model, where the highest temperatures naturally occur at the base of the convecting layer and inside isolated plumes or planes. As a matter of fact, this allows the existence of the thick lithosphere required to support the topography observed on Io.

The temperature Rayleigh number relationship: The averaged temperature Rayleigh number relationship follows a power law function, but, while in case of an impermeable boundary the exponent has the expected value of -.25, in case of a permeable boundary it is twice steeper: the average temperature decreases in proportion to $Ra^{-1/2}$. This result is of a major interest because it suggests that Ra for Io, i.e. the vigor of the convection acting in the Io's mantle, may be much lower than the one expected from studies leaded with numerical experiments performed in classical way. Tackley et al. [3] estimates that to reduce the inner temperature of Io to values compatible with volcanism, i.e. 1600K, the Rayleigh number would be in the range of 10^{12} to 10^{16}. The present power law relationship for open box implies that it would be closer to the square root of the above estimation, i.e. between 10^6 to 10^8.

Is Io's interior really molten: Asssuming that the amount of heat dissipated in the Io's interior is 10^{14} W, and that the mantle thickness is 880 km, allows to aproach the discussion in a dimensionalized framework. The difference between permeable and impermeable boundaries cases reveals more straightforward: at $Ra=10^6$, the averaged temperature lies around 4000 K in the former case whereas it reaches 48,500 K in the latter. In case of permeable boundary, the melting point of rocks is reached until $Ra=6\ 10^7$, i.e. for a maximum temperature above 1570 K. For higher Rayleigh numbers, no more silicate volcanism will occur on Io. In fact, recent infrared observations of Io's surface have revealed several discrete brightenings with temperature up to at least 1500K [5,6], only consistent with silicate volcanism. This Rayleigh number threshold corresponds to a mean viscosity of $2.3\ 10^{21}$ Pas, what appears low compared to the 500 K temperature reached in average. At the other end, a mainly molten mantle, i.e. with an averaged temperature of 1570 K, occurs for a Rayleigh number one order of magnitude lower, with a corresponding viscosity, $2.\ 10^{22}$ Pas, too high for such a temperature. This suggests that the convective regime would lie in a restricted range of Rayleigh numbers, barely less than one order of magnitude.

Actually, the feed-back due to the temperature dependence of viscosity should stabilize the Rayleigh number and so the thermal state of the Io's mantle: the decrease in viscosity resulting from a temperature elevation will lead to a higher Rayleigh number with a more efficient cooling providing a decrease in temperature. The equilibrium should occur for a Rayleigh number close to 10^7. it corresponds to a bulk viscosity of 10^{22} Pas which appears consistent with the associated depth temperature profile. In that case, a 100 km thick molten zone lies between the core and the mantle whose the rigid part extends over the first upper 400 km. This conclusion appears slightly sensitive to the amount of internal heating. As a matter of fact, an increase by a factor of 2 would just affected the Rayleigh number threshold discussed above by a factor of 4.

Conclusion: Using spherical models of mantle convection taking into account a heat pipe cooling mecanism, we have shown that volcanism is an efficient mecanism to evacuate the tremendous tidal heat dissipated in Io's mantle. In that case, the mantle temperature varies as $Ra^{-1/2}$ and remains below the melting point of rocks throughout the main part of the mantle, melting reamaining localized close to the Core mantle boundary and inside upwellings. This allows the existence of a thick rigid mantle able to support the topographic features observed on Io.

References: [1] Weeder G. J. et al. (1994) *JGR, 99*, 17095-17162. [2] Segatz M. et al. (1988) *Icarus, 75*, 187-206. [3] Tackley P.J. et al. (2000) *Icarus*, submited. [4] O'Reilly T. C. and Davies G. F. (1981) *GRL, 8* 313-316. [5] Spencer J.R. et al. (1997) *GRL, 24*, 2451-2454. [6] Stransberry J.A. (1997) *GRL, 24*, 2455-2458.