Impact basin relaxation at Iapetus

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A B S T R A C T

We investigate impact basin relaxation on Iapetus by combining a 3D thermal evolution model (Robuchon, G., Choblet, G., Tobie, G., Cadek, O., Sotin, C., Grasset, O. [2010]. Icarus 207, 959–971) with a spherical axisymmetric viscoelastic relaxation code (Zhong, S., Paulson, A., Wahr, J. [2003]. Geophys. J. Int. 155, 679–695). Due to the progressive cooling of Iapetus, younger basins relax less than older basins. For an ice reference viscosity of 10^{14} Pa s, an 800 km diameter basin relaxes by 30% if it formed in the first 50 Myr but by 10% if it formed at 1.2 Gyr. Bigger basins relax more rapidly than smaller ones, because the inferred thickness of the ice shell exceeds the diameter of all but the largest basins considered. Stereo topography shows that all basins 600 km in diameter or smaller are relaxed by 25% or less. Our model can match the relaxation of all the basins considered, within error, by assuming a single basin formation age (4.36 Ga for our nominal viscosity). This result is consistent with crater counts, which show no detectable age variation between the basins examined.

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1. Introduction

Crater relaxation has been used as a probe of subsurface temperature structure for over 30 years, both on terrestrial bodies (e.g. the Moon – Mohit and Phillips (2006) and Zhong and Zuber (2000); Mars – Pathare et al. (2005) and Mohit and Phillips (2006); Venus – Grimm and Solomon (1988); etc.) and icy satellites (e.g. Dombard and McKinnon, 2006). The importance of satellite crater relaxation studies is that they provide a constraint on the heat flux out of the body since the crater was emplaced. Different craters forming at different times thus experience different relaxation histories. Hence, if the ages of different craters are known, the evolution of the surface heat flux can be determined. Conversely, craters with different degrees of relaxation can be assigned different ages if a model for the thermal evolution of the body exists.

Early investigations of icy satellite relaxation used purely viscous, Newtonian formulations (e.g. Parmentier and Head, 1981; Passey and Shoemaker, 1982). More recent work has employed finite-element models which, while slower, can cope with non-Newtonian rheology and viscoelastic or viscoplastic effects (e.g. Thomas and Schubert, 1988; Hillgren and Melosh, 1989; Dombard and McKinnon, 2000). The current state of the art is well-described in Dombard and McKinnon (2006).

In this paper, we couple thermal evolution and relaxation models together to derive model constraints on the ages of different basins on Iapetus. Because of both model uncertainties and the difficulty in obtaining absolute cratering fluxes in the outer Solar System, we cannot yet address issues such as whether some of the basins formed during a Late Heavy Bombardment (cf. Charnoz et al., 2010). Nonetheless, we do find that the observed degree of basin relaxation is consistent with previously-derived models of Iapetus’ thermal evolution, and with available crater counts.

The reason for focusing on Iapetus is that it is heavily cratered (Dones et al., 2009) and appears to have been tectonically inactive since shortly after its formation (Singer and McKinnon, 2010). Furthermore, because of its distance from Saturn, tidal heating is not expected to have played an important role once synchronous spin was attained. Thus, one might expect Iapetus’ thermal evolution to be relatively simple and monotonic, which is certainly not the case for some of the other saturnian satellites. Lastly, its synchronous spin period and 35 km flattening (Porco et al., 2005) place some constraints on allowable thermal evolution models (Castillo-Rogez et al., 2009; Robuchon et al., 2010).

The rest of this paper is organized as follows. In Section 2 we describe the relevant observations, and in Section 3 we outline the two numerical models we employed. Sections 4 and 5 present the results and accompanying discussion, while Section 6 gives our conclusions.

2. Observations

Two flybys of Iapetus have taken place during the Cassini mission (Porco et al., 2005; Roatsch et al., 2009). During the second
flyby, in 2007, stereo images were acquired and used by Giese et al. (2008) to study the topography of Iapetus’ leading side (see Fig. 1). These authors determined the relationship between the diameter and depth of smaller craters. They also give the depth and diameter of a few larger basins.

The degree of relaxation of these large basins is currently a matter of some debate. Giese et al. (2008) and White and Schenk (2011) both state that the larger impact basins on Iapetus have undergone viscous relaxation. On the other hand, Schenk and Moore (2007) state that these same basins are deep and unrelaxed. This apparent disagreement is mainly due to uncertainty in the depth an impact basin on Iapetus would have in the absence of any relaxation. Below we consider a range of possible relaxation states, including some in which some of the basins have undergone no relaxation at all.

There are two pieces of evidence that might argue for at least limited basin relaxation. First, the Giese et al. (2008) data show that the largest measured basin (I in their numbering scheme (see Fig. 1)) is shallower than slightly smaller basins. This observation suggests that basin I at least has undergone relaxation. Second, Giese et al. (2008) argue that basin IV shows domical uplift suggestive of viscous relaxation of the kind modeled by Parmentier and Head (1981). Unfortunately, however, the quality of the topographic data makes it hard to distinguish between domical uplift and the central peaks observed in other basins (which are not due to late-stage relaxation). Furthermore, none of our relaxation models show evidence for this kind of domical uplift, because of the presence of a rigid near-surface layer not present in the Parmentier and Head (1981) model (see Section 4).

Basin III in Fig. 1 requires further discussion. This basin was identified by Giese et al. (2008), but not by Schenk and Moore (2007). Of all the basins examined, this one has the least obviously circular outline, it sits at the edge of the stereo coverage area and is partially obscured by the later basin VI (see Fig. 1). We therefore regard its status as an impact basin as questionable. Although it could be a real basin, the degree of degradation makes its relaxation state hard to assess, and below we generally exclude it from our analysis.

Fig. 2 plots the present-day depths and diameters of the large basins discussed in the Giese et al. (2008) study. To quantify the degree of relaxation, we require a prediction for the initial depth of a basin, prior to relaxation. We adopted three different approaches, which are intended to span the likely range of possibilities. One approach is to use the best-fit depth-diameter relationship obtained by Giese et al. (2008) at smaller diameters and assume that unrelaxed basins follow this same line. This relationship is shown by the red line in Fig. 2. A second approach is to use the method of Thomas and Squyres (1988), in which the deep

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Fig. 1. From Giese et al. (2008) showing the locations of large basins. The largest and smallest basins (labeled I and VII) have diameters of 800 and 340 km, respectively.
basin Herschel on Mimas is assumed to be unrelaxed, and unrelaxed depths for other basins are calculated based on a scaling which takes gravity and basin diameter into account. They used the following expression:

\[ d_b = d_H \left( \frac{D_b}{D_H} \right)^{0.301} \left( \frac{g_b}{g_H} \right)^{-0.311} \]

where \( d \) is the crater depth, \( D \) crater diameter and \( g \) gravitational acceleration. The subscripts \( b \) and \( H \) refer to a basin on Iapetus and Herschel respectively. Here we assume that \( d_b = 10 \) km (Thomas and Squyres, 1988; Moore et al., 2004). The scaling exponents were derived from craters on the Moon and Mercury and thus may not be appropriate for icy satellites. Nevertheless, Thomas and Squyres (1988) successfully used these scaling exponents to estimate the initial depth of Odysseus. This relationship is shown by the blue line in Fig. 2. A final, empirical alternative is to assume that the envelope defined by the present-day basins (basins with largest central depth: II, V and VI) represents the unrelaxed curve; that is, some of the basins are completely unrelaxed. This assumption is shown as the green line in Fig. 2. Because craters on most bodies show a break in slope in the diameter/depth ratio as crater size increases, our first two methods for predicting initial depth may be overestimates. However, in this case the third method will still give a good estimate of the initial depth.

In all cases, the degree of relaxation \( f \) is defined as

\[ f = 1 - \frac{d_{obs}}{d_u} \]

where \( d_{obs} \) and \( d_u \) are the observed and unrelaxed depths, respectively. Thus, using the three estimates for \( d_u \), the degree of relaxation for basin I is between 20% and 40% and the smaller basins (except for basin III) have experienced less relaxation. These quantitative estimates may then be compared with our model results.

3. Model

We combined two different kinds of numerical models. We compute the thermal evolution of an undifferentiated Iapetus using OEDIPUS (Choblet, 2005; Choblet et al., 2007). We calculate the basin relaxation using CitcomVE, a finite-element viscoelastic code developed by Zhong et al. (2003). Here we describe the two models, their interactions and the benchmarks performed.

3.1. Thermal model

The thermal evolution model and results are described in a previous study by Robuchon et al. (2010) and will only briefly be summarized here. The results are then used as an input for calculating the corresponding relaxation of a model impact basin (Section 3.2).

A thermal evolution model similar to ours was used by Castillo-Rogez et al. (2007), who proposed that Iapetus developed a fossil bulge as it despin from a rapid initial rotational rate. That model, however, did not take into account the possibility of convection inside Iapetus. As our model is very sensitive to the internal temperature structure, below we use the thermal evolution results obtained by Robuchon et al. (2010).

That work studied the internal dynamics of Iapetus using OEDIPUS in a 3-D spherical shell where only one sixth of the sphere was used. Since the grid mesh cannot be extended down to the center of the sphere, the numerical domain was bounded by an internal sphere corresponding to 20% of the radius. A constant temperature of 90 K was prescribed at the surface while a zero heat flux was prescribed at the inner boundary.

The viscosity \( \eta(T) \) was computed using an Arrhenius law. For Iapetus, the volumetric rock fraction is very small (~6%) (Thomas et al., 2007) and will not significantly affect the bulk viscosity (Friedson and Stevenson, 1983). Therefore, we assumed an ice shell density equal to the average density and following Arenson and Palmer (2005) that the viscous creep was mainly controlled by the ice. So, in this case we have:

\[ \eta(T) = \eta_m \exp \left( \frac{E_a}{R \exp(\frac{T_m}{T} - 1)} \right) \]

with \( T_m = 273 \) K and \( \eta_m = \eta(T_m) \left(=10^{14}\right) \) Pa s for our nominal model, the reference temperature and viscosity, \( E_a = 50 \) kJ mol\(^{-1}\) the activation energy and \( \gamma \) the gas constant. For numerical reasons, a cut-off was prescribed for the viscosity function: \( \eta = \min(\eta(T), 10^6 \eta_m) \).

We adopted a Newtonian model partly for simplicity, but also because the low stresses associated with convection are likely to result in ice deformation via diffusion creep, which is Newtonian (Goldsby and Kohlstedt, 2001). The stresses associated with impact basins are larger (a few MPa), and it is thus less clear that a Newtonian assumption is correct when modeling relaxation (see below).

The model incorporated decay of radioactive isotopes including the short-lived nuclides \( ^{26}\text{Al} \) and \( ^{60}\text{Fe} \). The amount of dissipated energy associated with despinning is relatively small compared to the radioactive energy (they obtained a maximum of \( \sim 1.7 \) GW at 10 Myr when at this time the radiogenic power is equal to \( \sim 70 \) GW) and is neglected in the thermal evolution calculation.

In order to calculate the despinning rate a Burgers rheology was assumed, in which a short-term viscosity \( \eta_b \) is defined in addition to the long-term viscosity \( \eta \). The Burgers model can be represented by coupling in series a Maxwell and Kelvin–Voigt spring-dashpot analogs. The short-term viscosity \( \eta_b \) stands for the Kelvin–Voigt element and is responsible for the anelastic response. This short-term element allows a second dissipation peak at low temperature and low rotational period. The shape evolution of Iapetus as despinning and lithospheric thickening takes place was calculated following the approach of Čadek (2003) and Tobie et al. (2008).

The two most important parameters in the Robuchon et al. (2010) model are the initial formation time of Iapetus (which controls the initial amount of \( ^{26}\text{Al} \) present) and the viscosity ratio \( \eta_b/\eta \), which is assumed constant. To obtain the observed present-day flattening and spin rate, there is a trade-off between these two
model parameters. Smaller initial amounts of $^{26}$Al result in a colder, less dissipative body; as a result, a lower short-term viscosity $\eta_{B}$ is required to still allow despinning to occur. Very low amounts of $^{26}$Al result in a flattening smaller than that observed, irrespective of $\eta_{B}$.

In warm terrestrial ice shelves, $\eta/\eta_{B} \approx 17$ (Reeh et al., 2003). It is not yet clear how to extrapolate this value to colder ice temperatures; thus Robuchon et al. (2010) considered a wide range in the ratio $\eta/\eta_{B}$ spanning 8 orders of magnitude. Here we restrict this ratio to only one order of magnitude. This corresponds to an initial concentration of $^{26}$Al of 46 72 ppb, roughly a factor of two variability. Below we will adopt a baseline model with 72 ppb; models with less $^{26}$Al will result in older basin formation times (see Section 5 below).

Because of its small size, the gravitational accretion energy of lapetus is small compared to the radiogenic heating. Furthermore, we are interested in the long-term relaxation of basins that formed some time after lapetus accreted. As a result, the initial temperature conditions (which are calculated based on the gravitational energy of accretion) are unlikely to affect the results presented here.

3.2. Viscoelastic model

Our viscoelastic model is based on the code CitcomVE developed by Zhong et al. (2003). We model lapetus as a self-gravitating incompressible viscoelastic (Maxwellian) sphere with a constant density $\rho_{0}$. Previous works (e.g. Dombard and McKinnon, 2000, 2006) have used a non-Newtonian rheology. However, for sufficiently small degrees of relaxation, this can be approximated by a Newtonian rheology (see the Appendix for details). Here, we use a Newtonian rheology in order to be consistent with the thermal evolution. The topographic response of the surface in response to surface loads (e.g. impact basins) are governed by the equations for conservation of mass and momentum, and for the gravitational perturbation (Wu and Peltier, 1982).

$$u_{ij} = 0$$

$$\sigma_{ij} + \rho_{0} \ddot{\phi}_{ij} - (\rho_{0} g u)_{i} = 0$$

$$\phi_{ii} = 0$$

where $u_{ij}$ is the radial component of the displacement, $\sigma_{ij}$ is the stress tensor, $\rho_{0}$ is the gravity and $\phi$ is the perturbation to the gravitational potential. (Einstein notation is used in Eqs. (4)–(6), with summation over repeated indices, and where a comma in the subscript denotes differentiation with respect to the spatial coordinate denoted by the following index.)

The rheological equation for a Maxwell body (i.e. an incompressible viscoelastic medium) is:

$$\sigma_{ij} + \frac{H}{\eta} \left( \epsilon_{ij} - \frac{1}{3} \sigma_{kk} \delta_{ij} \right) = 2 \mu_{e} \epsilon_{ij} + \left( k - \frac{2}{3} H \right) \epsilon_{kk} \delta_{ij}$$

where the over-dot is a time derivative, $\mu_{e}$ is the shear modulus, $\eta$ is the viscosity, $\epsilon_{ij}$ is the strain tensor and $\delta_{ij}$ is the Kroenecker delta.

Eqs. (4)–(7) are solved with a finite-element method and the displacement is computed for each point of the grid mesh (see Zhong et al. (2003) for additional details). We use a 2-D axisymmetric spherical mesh with 192 points in latitude between the pole and the equator and 80 points radially. Each point is equally spaced in latitude but radially the grid spacing changes, having three different grid sizes from very small close to the surface to larger close to the base of the numerical domain. The minimum radial resolution close to the surface is 2.42 km and the maximum is 10.95 km at large depth. Owing to the axisymmetry of the problem we set the basin center at the pole (note that we can nevertheless model a basin anywhere on the surface). We extended the grid mesh down to 10% of the radius of lapetus in order to prevent any numerical problems at the center. The exact position of the bottom boundary does not matter as long as the mantle thickness is larger than the basin diameter. We prescribed null shear stress and free-surface conditions at the surface and bottom boundaries, respectively. $\phi$ is continuous across the boundaries, and its gradient changes by $\nabla \phi^{+} - \nabla \phi^{-} = 4\pi G \rho_{B}(\mathbf{u})$, where $\rho_{B}$ is the density difference across the boundary, $\mathbf{u}$ is the displacement vector and $G$ is the gravitational constant.

All variables already used in OEDIPUS (e.g. the density or the surface temperature) are taken to have the same values here. The code also defines the density of the layer below our numerical domain which is set at the mantle value for the case of lapetus. Other values are given in Table 1.

In the numerical model dimensionless values are used: the time scale is the Maxwell time $t_{M} = \eta_{ref}/\mu_{ref}$ where $\mu_{ref}$ is the ice rigidity and $\eta_{ref}$ is the reference viscosity. In order to have a time step large enough to complete our simulation in few days, we choose $\eta_{ref} = 10^{22}$ Pa s. This time step is typically too large to resolve the initial rapid rebound associated with low-viscosity ice. However, by comparing results of short-term (<1 Myr) relaxation with smaller time-steps (with $\eta_{ref}$ from $10^{14}$ to $10^{21}$ Pa s (time step equal to ~3.5 h) to $10^{21}$ Pa s (time step equal to ~4000 years)), we were able to verify that the long-term (>1 Myr) relaxation calculated using $\eta_{ref} = 10^{22}$ Pa s was accurate to within 1%.

The original code has been modified in two ways: (1) the viscosity structure may evolve with time, based on the thermal evolution; (2) a topographic depression (basin) is introduced into the surface at the first time step in CitcomVE. This depression is modeled as a negative surface load of the same density as that of the underlying mantle, i.e. a mass deficit. The code computes the deformation at the surface due to this load and deforms the mesh accordingly. Temperature profiles are the average temperatures at each depth and different time steps calculated by OEDIPUS. As the grid mesh and time step of OEDIPUS do not match with those of the relaxation code, we perform two linear interpolations in time and space. Having determined the temperature field, we compute the viscosity following the Arrhenius law (Eq. (3)). Thus, a new viscosity structure is defined in CitcomVE every time-step. Note that we are assuming that the deformation due to the basin relaxation does not significantly affect the temperature field; this is reasonable, since the maximum basin depth is only a small fraction (<2%) of the radius of lapetus. We discuss the possible effects of impact heating on the temperature field in Section 5 below.

The initial shape of our basin is computed following a parabolic shape for which we can write the depth $z$ as:

$$z(\theta) = \frac{H}{R^{2}} r^{2} - H$$

with $H$ the initial basin central depth, $R$ basin radius and $r$ arc length from the pole to the colatitude $\theta$. When $r$ exceeds $R$, $z(\theta)$ is set to 0. For lapetus, we set $H$ to 10 km for bigger basins (400–800 km) and 7 km for smaller ones (100 and 200 km). These initial depths were

Table 1

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Associated quantity</th>
<th>Value</th>
<th>Unit</th>
</tr>
</thead>
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<tr>
<td>$R$</td>
<td>Average radius</td>
<td>735.6</td>
<td>km</td>
</tr>
<tr>
<td>$\rho_{0}$</td>
<td>Average density</td>
<td>1083</td>
<td>kg m$^{-3}$</td>
</tr>
<tr>
<td>$T_{w}$</td>
<td>Reference temperature</td>
<td>273</td>
<td>K</td>
</tr>
<tr>
<td>$E$</td>
<td>Surface gravity</td>
<td>0.22</td>
<td>m s$^{-2}$</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>Thermal expansion</td>
<td>1.70 $\times$ 10$^{-4}$</td>
<td>K$^{-1}$</td>
</tr>
<tr>
<td>$\kappa$</td>
<td>Thermal diffusivity</td>
<td>1.23 $\times$ 10$^{-6}$</td>
<td>m$^{2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$\mu_{ice}$</td>
<td>Shear modulus of ice</td>
<td>4.00 $\times$ 10$^{2}$</td>
<td>Pa</td>
</tr>
</tbody>
</table>
derived following the study of Giese et al. (2008). We emphasize here that for a Newtonian fluid the relaxation timescale is independent of the initial basin depth.

3.3. Benchmarks

To check our modifications, we performed some simple benchmarks. Analytical solutions are given by Melosh (1989) to compute basin relaxation in a viscous fluid with a constant or depth-dependent viscosity. The basin profile is decomposed into $J_0$ Bessel functions so the depth $z$ can be represented as a function of distance and time by:

$$z(r, t) = \int_0^{\infty} \zeta(k) \exp \left( -\frac{t}{t_R} \right) J_0(kr) dk$$

(9)

where

$$\zeta(k) = \frac{HD^2}{128} (kD)^2 \exp \left( -\frac{k^2D^2}{16} \right)$$

(10)

with $H$ initial basin central depth, $k$ wavenumber ($k = 2\pi/\lambda$) and $D$ basin diameter. The relaxation time $t_R$ is function of the viscosity profile. For a constant viscosity profile $t_R = 2\eta/(\rho g D)$ where $\rho$ is the average density and $g$ is the surface gravity. A more complicated solution exists for cases in which the viscosity varies exponentially with depth (Brennen, 1974).

In order to compare our model with the analytical solutions we need to bear in mind two differences: (1) the analytical solution is for a cartesian case while our code is spherical and (2) the analytical solutions are for a viscous case while our code is viscoelastic. To avoid problems due to sphericity, we chose small craters (a few degrees in diameter). To minimize effects arising from viscoelasticity, we carry out our comparison at times much greater than the Maxwell time, thus avoiding the effect of any initial elastic response. Care must also be taken to pick the correct reference level when determining the basin depth in CitcomVE, since conservation of mass results in slight deformation of the entire sphere when the basin is introduced. The results for a depth-dependent viscosity case with a $2^\circ$ radius crater are shown in Fig. 3. Our code gives a very good agreement with the analytical solution. We obtained a similarly good agreement with a constant-viscosity case.

4. Results

We used an initial basin profile described by Eq. (8) with a range of basin diameters (from 100 to 800 km) in order to represent the basin population seen on Iapetus. As the time of basin formation is poorly constrained, we tested different times of formation from 50 Myr to 1.2 Gyr after Iapetus' formation, representing a range from close to the end of the accretion to after the putative Late Heavy Bombardment.

4.1. Iapetus case

The baseline thermal evolution model is shown in Fig. 4a. The evolution is characterized by an onset of convection at 5.3 Myr after the start of the model. The surface heat flux is 6 mW m$^{-2}$ at 50 Myr and 1.6 mW m$^{-2}$ at 1 Gyr. Convection prevents the ice from melting and a maximum temperature of 256 K (corresponding to a viscosity of $4 \times 10^{14}$ Pa s) is obtained. The lithospheric thickness (defined by the 140 K isotherm) is a minimum at the start of the model and then progressively increases up to $\sim$18 km after 50 Myr and 70 km at 1 Gyr.

Fig. 4b shows the evolution of an 800 km basin forming at 50 Myr. The initial relaxation at the center is very rapid (1 km in 2 years, blue curve on Fig. 4b). The relaxation then begins to slow down and another kilometer of uplift is obtained after more than 5500 years. A rim begins to form, but the uplift is only tens of meters. Finally, a central depth of 7 km (30% relaxation) and a rim of $\sim$100 m are reached at the end of our simulation.

The initial rapid relaxation is a consequence of the high initial temperature. After 50 Myr the mantle temperature beneath the lithosphere is roughly 240 K (see Fig. 4a) and so a viscosity of about $10^{15}$ Pa s is expected. Based on the analysis of Section 3.3, the corresponding relaxation timescale ($\sim \tau/\rho g D$) is of order a year, in agreement with the initial model behavior. The subsequent slower relaxation occurs because the lithosphere is more viscous, and thus takes longer to relax. We note that, unlike Parmentier and Head (1981), a central domical uplift is not obtained, because of the presence of this high-viscosity lid.

Fig. 4c shows the present-day relaxation fraction $f$ as a function of basin diameter and formation age. For the model results $f$ is defined as $1 - (z/H)$ where $z$ is the present-day model depth at the center. For all basins the maximum relaxation occurs for the earliest formation time, when the ice shell viscosity was low. A relaxation of 30% is obtained at 50 Myr for the largest basins while the smallest relax by about 5%. Subsequently, the lithosphere thickness increases while the interior viscosity also increases due to cooling via convection. At progressively later formation times, the total relaxation fraction is correspondingly reduced. An 800 km diameter basin forming at 1200 Myr is expected to relax by only 10%.

In general, bigger basins relax more rapidly than smaller ones. This behavior is due to the fact that wider basins exert a pressure perturbation to greater depths, resulting in both a greater effective channel thickness and a lower average viscosity, and hence more rapid relaxation. Basins with widths exceeding the shell thickness should show different behavior, because the bottom boundary restricts the rate at which flow can occur. The largest basin we modeled (800 km diameter) is comparable to the radius of Iapetus (735.6 km). The initial relaxation rate of this basin is comparable to that of the 600 km diameter basin, suggesting that the influence of the bottom boundary is important in this case.

4.2. Comparison with observations

Fig. 5 plots our model basin relaxation fractions compared with the observations (Section 2). Because of the uncertainties in the
observed relaxation fraction, it is just possible to find a single model age (200 Myr, equivalent to an age of \( \approx 4.36 \, \text{Ga} \)) that fits all the observations (with the exception of basin III – see above). If we exclude the most conservative relaxation estimates (black squares), the implication is that basin I is older than basin IV, and both are a few hundred Myr older than basins II, V and VI. Because of the crater counts (see below), we do not favor this interpretation, but note that basin I must be older than basin IV due to the fact that basin IV is superposed on basin I (see Fig. 1).

As discussed below, the absolute model ages that we derive depend on the assumed viscosity and \(^{26}\text{Al}\) concentration. However, the relative model ages will not be affected by these parameters. These relative ages can then be compared with those derived from crater counting as a check on the model.

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Fig. 4. (a) Iapetus thermal evolution used as input into the relaxation computation (from Robuchon et al., 2010). Vertical black bars give the assumed formation times of model impact basins. (b) Example of relaxation history of a basin with diameter 800 km, initial depth 10 km, and formation time 50 Myr after CAIs. The lowest curve (red) shows the initial profile. Progressively higher are later profiles at 14.47 days, 57.9 days, 217 days, 1.98 years, 39.6 years, 5545 years, 0.119 Myr. Here the reference viscosity \( \eta_{\text{ref}} \) (which controls the time step) was progressively increased from \( 10^{14} \, \text{Pa s} \) (time step equal to 3.47 h) to \( 10^{19} \, \text{Pa s} \) (time step equal to \( \approx 39.6 \, \text{years} \)). (c) Present-day relaxation fraction versus time of basin formation for each different basin diameter tested. Relaxation fraction is defined by \( 1 - z/H \) where \( z \) is the present-day central depth and \( H \) is the initial central depth.
Based on craters with diameters >20 km counted by one of us (MK), the crater densities for basins IV, II and V are 65 ± 16, 84 ± 32 and 84 ± 38 per 10^6 km^2, respectively. Assuming that the basins are not saturated, all three basins thus have identical ages within error. That conclusion is consistent with the conclusion we drew above based on our model results that all basins examined could be the same age. In short, although we cannot exclude a more protracted epoch of basin formation, neither the crater counts nor our modeling require it.

5. Discussion

The results shown in Figs. 4 and 5 are for a nominal lapetus model with 72 ppb initial 26Al. As discussed in Section 3.1, there are alternative models with lower 26Al values which fit the despinning and flattening constraints equally well. Because such models undergo more rapid cooling, relaxation is less efficient. A model with 46 ppb initial 26Al – at the bottom end of acceptable values – results in relaxation fractions which are roughly one-third smaller for a given time and crater diameter than the values shown in Fig. 5. Again, all basins could have formed at the same time, ~4.49 Ga in this case. The model basin ages are thus sensitive to variations in initial 26Al. An even later accretion for lapetus (i.e. less 26Al) would result in even more unrelaxed basins – but would not then be consistent with the acquisition of a fossil rotational bulge as observed by the Cassini mission (see Robuchon et al. (2010)).

As previously mentioned we choose a thermal evolution which assumes an undifferentiated lapetus. Although differentiation remains unlikely (see Robuchon et al. (2010) for discussion), sequestration of heat-producing elements into a rocky core would lead to higher maximum temperatures in the icy mantle compared with the results presented here. However, whether these higher temperatures would lead to longer-lived convection (or indeed a subsurface ocean) is less obvious, but critically important to understanding the long-term relaxation behavior. Thus, further numerical studies will be required if lapetus is indeed differentiated.

Our initial topography profile is a parabolic shape with no rim (cf. Eq. (8)). This shape is similar to the Bessel function shape used by Melosh (1989) and to the fourth-order polynomial used by Dombard and McKinnon (2006), although the latter also included a crater rim. We find that the initial shape does not significantly affect the final results. As expected for a Newtonian fluid, for a given diameter the relaxation timescale is essentially independent of the initial basin depth.

The effect of the crater formation on our initial temperature profile introduces additional complications. The shock wave generated by the impact creates a pressure peak which is almost uniform in the region (the isobaric core) near the impact point. In this volume, the material is strongly heated (e.g. Senshu et al., 2002), creating a temperature anomaly below the impact point. The size of the isobaric core is less than 10% of the size of the basin diameter (37 km for a basin diameter of 400 km using Eq. (12) from Zahnle et al. (1998) and Eq. (2) from Senshu et al. (2002)). We carried out some additional simulations (not shown) to investigate the consequences of impact heating. Although there are significant local effects (resulting in the formation of a central peak, as proposed by Dombard et al. (2007)), the thermal anomaly diffuses rapidly and the final overall relaxation state of the basin is essentially unaffected. This conclusion is consistent with the existence of apparently unrelaxed basins, such as Herschel on Mimas and Odysseus on Tethys.

We assumed an incompressible Maxwell-type viscoelastic body. We did not include plastic behavior, in contrast to the study of Dombard and McKinnon (2006). Nevertheless, as noted in their study for a case with zero heat flux at the surface, the low stresses at the surface of lapetus should prevent any plastic deformation and are thus unlikely to significantly affect our conclusions.

We used parameters for pure water ice, but in the case of lapetus the volumetric fraction of silicate is estimated at 6.3% (Thomas et al., 2007). It is known that the presence of rock particles may slightly decrease the creep rate (Arenson and Palmer, 2003), but this effect remains small with such a small rock fraction (Friedson and Stevenson, 1983; Mangold et al., 2002).

An apparent shortcoming of our model is its use of Newtonian viscosity, which is likely appropriate for the low stresses associated with convection, but may not be appropriate for the higher stresses (~MPa) associated with basin relaxation. However, in the Appendix we demonstrate that for small degrees of relaxation, non-Newtonian relaxation may be adequately modeled by using a case with a Newtonian viscosity which equals the initial non-Newtonian effective viscosity. Thus, in the case of lapetus (small relaxation), our adoption of a Newtonian rheology is a smaller source of uncertainty than that of the appropriate reference viscosity.

All the model runs of Robuchon et al. (2010) assumed a nominal reference viscosity of 10^14 Pa s. As that paper points out, the self-regulating nature of convection ensures that, irrespective of the reference viscosity, the actual viscosity will adjust itself so that heat production and convective heat loss are roughly in balance (cf. Tozer, 1965). This logic breaks down if tidal heating (which is viscosity-dependent) is important, but in the case of lapetus the dominant source of heating is radioactive decay. Thus, varying the reference viscosity will affect the temperature structure much more strongly than the viscosity structure. Since it is the viscosity structure which controls the relaxation rate, we do not anticipate that reasonable variations in reference viscosity will change our conclusions.

Neither our modeling nor crater counts can resolve age differences between the different basins examined. However, examining some of the younger, less heavily-cratered basins may provide additional information. For instance, Engelier has a diameter similar to that of Falsaron (basin IV), but is much less heavily cratered (6 ± 6 craters >20 km per 10^6 km^2) and therefore much younger. Although absolute cratering rates are poorly-constrained in the outer Solar System (Zahnle et al., 2003), it seems clear that basins like Herschel and Odysseus have an age less – perhaps much less – than 3 Ga (Dones et al., 2009). Assuming that Engelier has a similar age, the predicted relaxation fraction from Fig. 4 is <5%, implying a depth in the range 8–9 km (Fig. 1). Should Engelier turn out to be shallower than this prediction (e.g. similar in depth to basin IV), this would imply that basins with very different crater densities must have similar ages, unless regional variations in heat flux are
important. In the former case, this would imply an early and rapid decrease in cratering flux, likely consistent with the observation that lapetus has many more 100 km-scale basins than one would predict based on current impact rates (Dones et al., 2009). Thus, measuring the depths of some of the younger basins on lapetus could provide important constraints on the history of the impact flux.

6. Conclusion

The main message of this paper is one of self-consistency. A model developed to explain the global characteristics of lapetus (spin state and flattening) also reproduces the slightly-relaxed state of the big impact basins there. All the basins examined could have the same model age, which is in agreement with the available crater count constraints. Thus, the relatively simple picture of lapetus’ evolution presented in Robuchon et al. (2010) appears to be consistent with other observational constraints.

Modeling of the kind presented here would obviously benefit from better constraints on impact basin ages. Relative ages can be established with crater counting, but absolute ages are more difficult, and perhaps impossible, to establish. Indeed, in theory one could use relaxation models of the kind presented here to actually estimate basin ages, though in practice the modeling uncertainties probably make this goal unrealistic. In the more immediate future, it will be of interest to apply the models developed here both to younger basins on Iapetus, and to other bodies. In particular, Tethys has a young, unrelaxed basin (Odysseus), and some independent constraints on its thermal evolution (Giese et al., 2008; Chen et al., 2008). Some of Rhea’s basins appear to have undergone partial relaxation, while Dione has a 350 km diameter basin that is both relaxed and young (Schenk and Moore, 2007). Combining crater relaxation studies with thermal evolution models is likely to throw additional light on the histories of these interesting satellites.

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Appendix A

We develop in this section a heuristic calculation to illuminate the difference between Newtonian and non-Newtonian viscous relaxation. In the case of Newtonian relaxation we can write the basin depth \( h \) as a function of the initial central depth \( h_0 \) and relaxation time \( t_{en} \):

\[
h(t) = h_0 \exp \left( -\frac{t}{t_{en}} \right)
\]

where for a constant viscosity half-space the relaxation time is given by

\[
t_{en} = \frac{2\pi \eta N}{\rho g D}
\]

with \( \eta N \) viscosity, \( \rho \) density, \( g \) surface gravity and \( D \) crater diameter. In the non-Newtonian case, the effective viscosity depends on the stress \( \sigma \):

\[
\eta_{en} = \frac{1}{3A} \exp \left( \frac{Q}{RT} \right) \sigma^{1-n}
\]

with \( A \) and \( n \) material constants, \( T \) temperature and \( Q \) activation energy.

Making the heuristic assumption that the characteristic stress \( \sigma \approx fgh \), with \( f \) a dimensionless constant, we can use Eq. (A.1) to write:

\[
\frac{dh}{dt} = -\frac{D}{2\pi c} h^{n-1}(\rho g)^n h^n = -kh^n
\]

with

\[
c = \frac{1}{3A} \exp \left( \frac{Q}{RT} \right) \quad \text{and} \quad k = \frac{D}{2\pi c} h^{n-1}(\rho g)^n
\]

Integration of Eq. (A.4) with the initial condition \((t = 0 \rightarrow h = h_0)\) gives:

\[
h(t) = \left( \frac{h_0}{1 + (n-1)kh_0^{n-1}} \right)^{\frac{1}{n}}
\]

We can also write the relaxation time \( t_{enN} \) which is defined by the time at which \( h/h_0 = 1/e \):

\[
t_{enN} = \frac{(e^{n-1} - 1)2\pi \exp \left( \frac{Q}{RT} \right)}{(n-1)(h_0)^{n-1}D(\rho g)^n}3A
\]

where the constants \( k \) and \( c \) have been written out in full.

To verify this approach, we compared our Newtonian and non-Newtonian relaxation times (Eqs. (A.2) and (A.7)) with the numerical results shown in Figs. 5 and 6 of Thomas and Schubert (1987). Fig. 6 shows that our analytical solution does a good job of matching the numerical results for \( f \approx 0.15 \). We also verified that Eq. (A.6) reproduced the individual non-Newtonian crater evolution plots shown in Fig. 12 of Thomas and Schubert (1987) (not shown).

If we further define the relaxation fraction \( r = 1 - \frac{h}{h_0} \), then we can compare the time for a Newtonian and non-Newtonian rheology to reach the same relaxation fraction. Defining an effective non-Newtonian viscosity using Eq. (A.3) and making use of Eqs. (A.2) and (A.7), it can be shown that the ratio of the two timescales is given by

\[
\frac{t_{enN}}{t_{en}} = \frac{\eta_{enN}}{\eta_{en}} \left( \frac{1 - r}{1 - n \ln(1 - r)} \right)
\]

Clearly, when the relaxation fraction \( r \) is small (as is the case for lapetus), the ratio of the two timescales approaches the ratio of
the effective viscosities, depending on the value of $n$. At larger relaxation fractions, non-Newtonian relaxation becomes increasingly slow compared to the Newtonian case, because the reduction in driving stresses increases the effective viscosity.

References