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Tectonothermal evolution of solid bodies: terrestrial planets, exoplanets and moons

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Tectonothermal evolution of solid bodies: terrestrial planets, exoplanets and moons

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A framework for understanding the tectonothermal evolution of solid planetary bodies has historically been lacking owing to sparse observational constraints. Developments in simulating the physical interiors and tectonic behaviour of terrestrial planets have allowed insights into the relevant physics and important factors governing planetary behaviour. This contribution summarises the critical factors in determining a planet's tectonic regime, and the application of this framework to understanding terrestrial planet evolution. Advances in modelling have led to the identification of new, unmapped tectonic regimes, such as episodic convection, which has relevance to our understanding of the evolution of the early Earth, Venus and Saturn's moon Enceladus. Coupling of tectonic and atmospheric models for planetary evolution has contributed to our knowledge of Martian and Venusian degassing histories, and recent debate on the tectonic regime of exosolar planets informs outstanding questions on their habitability. Ultimately, a framework for terrestrial planet evolution will couple available cosmochemical, geochemical and astrophysical constraints into an emerging generation of simulation tools, facilitating the mapping of terrestrial planet behaviour over a wide parameter space.

KEY WORDS: geodynamics, planets, tectonics, mantle convection.

INTRODUCTION

Plate tectonic processes on Earth exert a first order control on the geological record, and also on long time-scale atmosphere, hydrosphere and even biosphere cycles (e.g. Kastings 2010). However, of the terrestrial planets, only Earth is known to possess plate tectonics; and so for comparative planetology, plate tectonics would seem to be the exception, rather than the rule (O'Neill *et al.* 2007a). Given its influence on the Earth, a discussion of planetary tectonics is of fundamental importance in understanding the surface conditions of other solid planets or moons.

However, some workers have argued that throughout much of its history, Earth did not possess plate tectonics (e.g. O'Neill *et al.* 2007b; and see Condie & Pease 2008, and papers within). While there is evidence for subduction and recycling up to *ca* 3.0 Ga (Calvert *et al.* 1995; O'Neill *et al.* 2007b; Condie & Kroner 2008), it is not clear whether this represents steady-state subduction or discrete events (O'Neill *et al.* 2007b), and Stern (2008) even suggests modern plate tectonics only initiated in the late Proterozoic. This raises questions about the interconnectivity of the biosphere, atmosphere and hydrosphere, with Earth's active tectonic system over long time-scales. Indeed, many smaller bodies and icy moons in the solar system seem to be able to sustain liquid water oceans (albeit subsurface) without the operation of plate tectonics but powered by vigorous tidal heating (Greenberg 2005). So the question rather becomes what end-members of tectonic activity are to be expected on solid planets? And how are these tectonic regimes able to affect the surface—or subsurface—conditions?

This contribution will explore the factors controlling the tectonic regime of a planet, and how they evolve through time, with specific focus on Australian contributions to the debate. We will also explore the impact surface tectonics—or lack thereof—has had on the atmospheric, hydrospheric and geological evolution of terrestrial planets and moons.

THE BALANCE OF POWER

Active tectonism on a terrestrial planet occurs primarily through brittle faulting or shearing of an otherwise highly rigid lithosphere. Moresi & Solomatov (1998) was one of the first works to explore this style of convection at planetary scales—made possible through advances in finite element modelling of temperature-dependent mantle convection (Moresi & Solomatov 1995). The forces powering these systems are highly coupled convective stresses, which arise primarily owing to thermal buoyancy anomalies within the mantle. Traditional plate driving forces, such as ridge-push, and slab-pull, are implicitly part of this convective stress balance. The temperature dependence of viscosity

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means the cold highly viscous lithosphere is resistant to flow. This, however, leads to amplification of stress in the shallower lithosphere, which eventually may exceed the yield stress, allowing brittle failure to ensue (Figure 1).

The simulation of surface plate dynamics has a long history within the geodynamics community (e.g. Moresi & Solomatov 1998; Tackley 2000, 2008; Bercovici 2003; Stein *et al.* 2004). While some form of plastic or stick-slip rheology is minimally required to allow the cold, viscous surface plates to deform (Bercovici 2003), generating plate-like behaviour involves shear-weakening behaviour (Tackley 2000; Bercovici 2003), low viscosity asthenospheric channels (Richards *et al.* 2001), and elastic–plastic behaviour of plates (Tackley 2000). The physics of damage-dependent rheologies is an ongoing area of research (Bercovici 2003).

Within the complexities of plate rheology, there are three possible regimes of modelled plate behaviour. First, if the convective stresses of the system are insufficient to exceed the yield stress of the lithosphere, no surface deformation or tectonism will occur. The lid will effectively stagnate—a regime known as stagnant lid convection, and applicable to Mars or the Moon (Moresi & Solomatov 1998). The weakly convecting interior of these two bodies is unable to overcome the resistance offered by the thick, cold lithosphere, which had remained stagnant for much of their evolution.

At the other extreme is the case where the convecting interior is able to generate stresses far in excess of the lithospheric yield stress. This results in failure of the lithosphere along fault zones, and facilitate its recycling into the mantle beneath. This regime is known as mobile-lid convection, and plate tectonics on Earth is an example (Moresi & Solomatov 1998). A key feature of this regime is that recycling of the lid enables efficient cooling of the mantle and the core—temperatures in the interior are far lower than in an equivalent stagnant lid regime.

At the boundary between these two regimes, when the convective stresses are of similar magnitude to the plate strength, the system enters a highly nonlinear state. Called episodic convection by Moresi & Solomatov (1998), the system essentially oscillates between long periods of quiescent, stagnant behaviour and periods of rapid plate overturn and recycling (Figure 1). While Figure 1 shows an example for purely basal heated

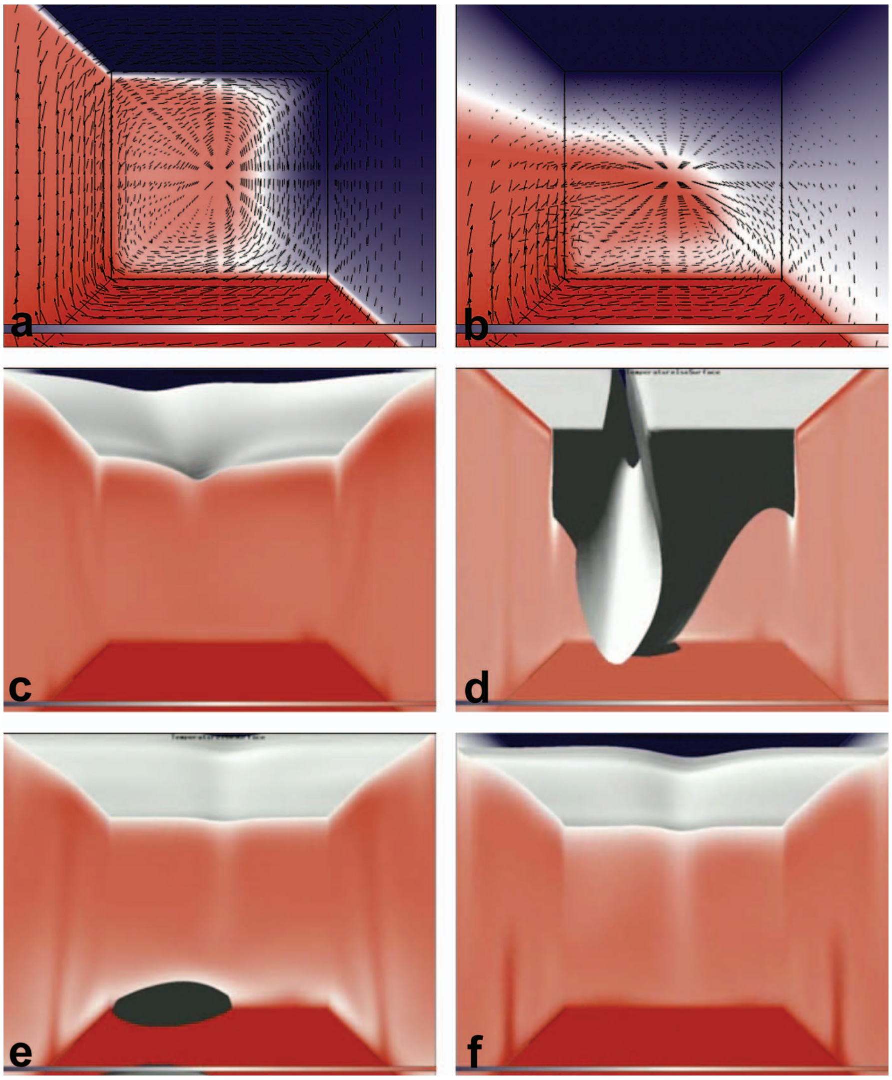

Figure 1 Examples of Mobile (a), Stagnant (b), and Episodic overturn (c–f) regimes. Simulations run using Underworld (Moresi *et al.* 2007). Simulations run at Ra = 1e7, with purely bottom heating and reflecting side boundary conditions, and a resolution of $32 \times 32 \times 32$, for a unit box, otherwise as per Moresi & Solomatov (1998), but in 3D. All the models employ a temperature-dependent Frank-Kamenetskii viscosity, with a contrast of 3e4 over the mantle. Differences in tectonic behaviour arise purely owing to differences in non-dimensional depth-independent yield stress (2e5 for mobile, 4.5e5 for episodic, and 8e5 for stagnant). Arrows in (a) and (b) represent velocity, and temperature bar shows temperatures between 0 and 1 (non-dimensional). In (c)–(f), the isosurface represents the $T = 0.2$ contour.

convection, for illustrative purposes, episodic behaviour has been demonstrated for predominantly internally heated systems also (Stein *et al.* 2004; O'Neill *et al.* 2007b). Initial observations of the surface of Venus from the Magellan mission suggested that, owing to the uniformity of the cratering distribution, the entire surface of the planet had experienced a resurfacing episode approximately 750 Myr ago. Turcotte (1993) first suggested an episodic mechanism to explain the surface age distribution, based on isoviscous scaling theory. Moresi & Solomatov (1998) offered a mechanical explanation based on a plausible rheology for the Venusian surface. In their model, the lithospheric strength on Venus exceeds that of Earth owing to a lack of surface water, placing Venus in an episodic convection regime.

A BESTIARY OF TERRESTRIAL PLANETS

O'Neill *et al.* (2007a) developed a scaling relationship for convective stresses and lithospheric strength, with the view to developing a framework for delineating tectonic regime boundaries. In their model, convective stresses scale with mantle viscosity and velocity gradients:

$$\tau_{xz} = \frac{v_x \eta}{\delta_{\mathrm{vel}}} \tag{1}$$

Here τ_{xz} is the shear stress, v_x the sublithospheric velocity, η the viscosity and δ_{vel} the thickness of the velocity boundary layer. The boundary between mobile and stagnant regimes is defined by equating the convective stresses with the lithospheric strength. At higher temperatures within the mantle and lithosphere, mantle material primarily deforms via viscous flow, and the response to stress is governed by a temperature-dependent viscous flow law. At lower temperatures, the stress required to produce a nominal viscous strain becomes much greater, owing to the temperature dependence of the viscosity. At low pressures, brittle failure is the dominant deformation mechanism. However, the yield stress, beyond which brittle failure occurs, is pressure dependent, and becomes greater with increasing depth. The maximum stress the lithosphere is able to support occurs at the intersection of the viscous flow law with the pressure-dependent yield criterion. This point of intersection will depend on the size and thermal state of a planet. Hotter, more vigorously convecting systems will have a thinner lithosphere than equivalent, colder systems. This tends to move the point of intersection up, to lower yield stresses. The lithospheric strength can therefore be considered to scale linearly with boundary layer thickness—which itself follow a well-known relationship with the (rheological) Rayleigh number:

$$\tau_{c,xz} = AB \frac{\eta \kappa}{d_{\mathrm{mantle}}^2} \frac{Ra_{\mathrm{rh}}^{\beta+\lambda}}{\theta^{\beta+\alpha}}. \tag{2}$$

Here, κ is the thermal diffusivity, d_{mantle} the mantle thickness, Ra_{rh} the rheological Rayleigh number and η the viscosity contrast; other symbols as per Equation 1. The constants α, β and γ depend on the fluid problem as discussed in O'Neill *et al.* (2007a), A and B are scaling constants.

Equating the terms for lithospheric strength, and convective stresses, O'Neill *et al.* (2007a) derived a relationship for the transition from stagnant to mobile systems:

$$\mu \rho g (c\delta) \approx \left(\frac{v\eta}{\delta_v}\right) \approx \frac{A(\kappa/d_{\mathrm{mantle}})\,(Ra/\theta)^{\beta}\eta}{f\delta}$$

$$\left(\frac{1}{d_{\mathrm{mantle}}}\right)^2 (c\delta)^2 = \left(\frac{1}{d_{\mathrm{mantle}}}\right)^2 \frac{Ac}{f}\frac{\eta\kappa}{\mu\rho g}\frac{1}{d_{\mathrm{mantle}}}\left(\frac{Ra}{\theta}\right)^{\beta}$$

$$\frac{d_{\mathrm{BDT}}}{d_{\mathrm{mantle}}} \approx E\left(\frac{\eta\kappa}{\mu\rho g}\right)^{\frac{1}{2}}\left(\frac{1}{d_{\mathrm{mantle}}}\right)^{3/2}\frac{Ra^{\beta/2}}{\theta^{\beta/2}}. \tag{3}$$

Symbols are per Equations 1 and 2; μ is the coefficient of friction, ρ the density, g surface gravity, δ boundary layer thickness and d_{BDT} the depth to the brittle ductile transition. A, c, f and E are scaling constants (O'Neill *et al.* 2007a). While this scaling arguments ignores non-linearities in the system (such as time-varying velocity near the tectonic transition, or changing internal viscosities through time—which will be discussed in the following sections), it allows a simple mapping of the transition from mobile to stagnant planets and, given estimates for their interior velocities and elastic lithospheric thickness, a basic diagram of terrestrial planet behaviour, shown in Figure 2.

While the Moon and Mercury fall squarely in the stagnant-lid field for their entire evolution, Earth, as the most strongly convecting system plotted, falls in the mobile field over its lifetime. Venus currently plots near the transition, consistent with the suggestion it is currently in an episodic regime at the transition between mobile lid and stagnant lid convection. While Mars is stagnant now, it may have been propelled into the mobile lid regime early in its history if it had water on the surface, lowering the yield strength of the lithosphere. Furthermore, Io and Europa, Jupiter's innermost two moons, plot close to the transition between mobile and stagnant lid. Io is currently the most volcanic body in the solar system, which obscures most evidence of any tectonism. However, about 2% of Io's surface consists of rugged, non-volcanic mountain belts, exhibiting large scarps and other evidence of active tectonics—a hint of surface tectonic processes consistent with being at the boundary of active lid behaviour (Carr *et al.* 1998; Figure 3). Europa has an extremely young surface, approximately *ca* 80 Myrs old, with surface activity apparently declining since then into a current period of cryovolcanism and primarily tidal cracking (Figueredo & Greeley 2004; Figure 3). This time-dependent resurfacing, however, is again consistent with its placement in Figure 2 near the boundary of active surface tectonics.

STAGNANT LID EVOLUTION AND THE DECLINE OF THE MARTIAN ATMOSPHERE

Mars is perhaps the benchmark for the decline of a planet within a stagnant lid regime. The geomorphology and surface mineralogy of Mars both demonstrate the

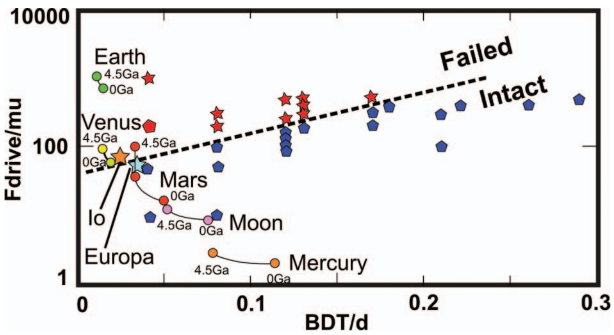

Figure 2 Plot of the driving stress powering plate dynamics—parameterised as the product of the sublithospheric viscosity and velocity, divided by the coefficient of friction (mu), *vs* the ratio of the brittle-ductile transition (BDT) to the depth of the convecting system—a measure of plate strength (modified from O'Neill *et al.* 2007a). Small stars and polygons show results from numerical simulations (O'Neill *et al.* 2007a). Planetary bodies are also labelled, for both present conditions (0 Ga), and at 4.5 Ga, assuming velocity scales with Ra and sublithospheric viscosities remain constant.

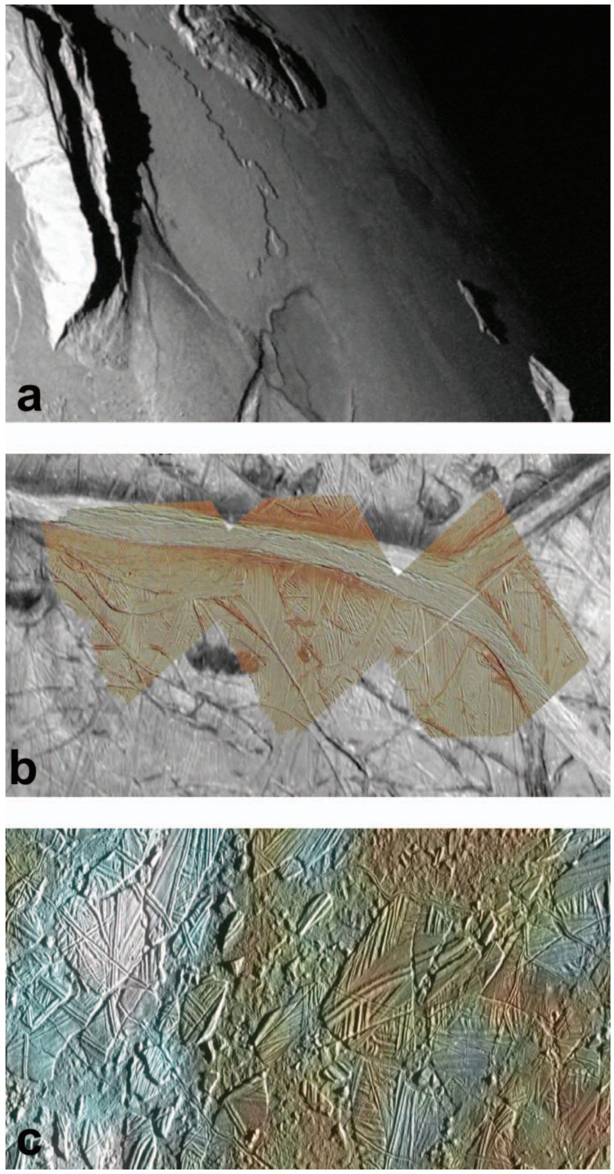

Figure 3 (a) Non-volcanic mountain ranges on Io. Image taken from the NASA Galileo mission, when the sun was low, so the scene is illuminated from the left. Mongibello Mons are the mountains in the left of the image. (b) Convergence feature on Jupiter's moon Europa. Shown is Agenor Linea (high-resolution colour images embedded in low-resolution background). The centre of the image is at 41°S/189°W. North is to the top right, and the image is ~135 km across. (c) Ice crust mobility in the Conamara region of Europa, taken during the Galileo mission in 1996. Distinct ice rafts have been fractured and remobilised during a later episode of surface activity. North is to the top; the sun illuminates the image from the right. The centre is at 9°N/274°W and is ~70 km wide. Image credits: NASA/JPL/UA/ASU.

existence of liquid water on its surface early in its evolution, during the appropriately dubbed Noachian period (Jakosky & Phillips 2001). Previous work has indicated that Mars began with a sizeable magma ocean (Elkins-Tanton *et al.* 2005; Debaille *et al.* 2007; Caro *et al.* 2008). Neodymium isotopic constraints suggest a delayed freezing and overturn of an initial Martian magma ocean—a scenario possible if a thick steam atmosphere existed for several hundred million years after formation of the planet (Debaille *et al.* 2009). This implies an unusual longevity of the thick primordial atmosphere on Mars; and the existence, and blanketing effects, of a thick steam atmosphere is discussed in Abe (1997). It would also leave a small window of time between primordial mantle overturn, and extraction of crustal-contaminated shergottiites (Norman 1999). The Noachian saw the emplacement of most of Mars's crust (e.g. Norman 1999), and the gigantic Tharsis rise, the most prominant volcanic construct in the solar system (Jakosky & Phillips 2001). There has been a suggestion of early surface activity on Mars, on the basis of observed magnetic lineations in the southern highlands, and indeed existence of the paleomagnetic field record observed in the oldest crust, requiring an active geodynamo, which has been suggested to have required active mantle cooling and lid resurfacing (Connerney *et al.* 1999; Nimmo & Stevenson 2000).

The presence of liquid water on the surface of Mars in the Noachian would require an atmosphere of several bars at this time (O'Neill *et al.* 2007c). The current surface atmospheric pressure is just 6.8 mbar, implying the vast majority of Mars' early atmosphere has been lost (O'Neill *et al.* 2007c). Analysis of gas inclusions in the ALH84001 meteorite show no fractionation in $^{15}N/^{14}N$ or $^{38}Ar/^{36}Ar$ (Marti & Mathew 2000; Mathew & Marti 2001). This is consistent with hydrodynamic escape or impact loss mechanisms, which do not induce isotope fractionations, dominating atmospheric loss prior to 3.9 Ga (O'Neill *et al.* 2007c). Subsequent to this epoch, solar wind interactions such as sputtering or hydrodynamic collisions would dominate, resulting in the isotopic fractionation currently observed in the atmosphere (Jakosky & Phillips 2001). The D/H ratio of the atmosphere suggests around 60–90% of the water present in Noachian has subsequently been lost to space (Yung *et al.* 1988; Jakosky & Phillips 2001).

Volcanic degassing is the primary mechanism by which the Martian atmospheric volatile budget is replenished, and Mars' atmospheric and surface evolution is strongly tied to the evolution of volcanic activity through time. Mars is still volcanically active, as demonstrated by the identification of young (>2 Ma) lava flows, observed cinder cones near the North Pole (Garvin *et al.* 2000), identification of atmospheric methane (Formisano *et al.* 2004) and the young ages of many Martian meteorites (180–160 Ma; Borg *et al.* 2005). The declining volcanism of Mars has been mitigating the steadily diminishing atmosphere of the planet since the Noachian. Jakosky & Phillips (2001) calculated that the emplacement of the giant Tharsis province may have produced enough volatiles—primarily CO_2 and H_2O—to drive the atmospheric pressure to several bars and transiently allow liquid water on the surface. O'Neill *et al.* (2007c) calculated the evolution of the rate of volcanism and atmospheric CO_2 through time, from stagnant lid convection models, which were constrained by thermal, geological, geophysical and petrological constraints. The simulations used an Arrhenius-style viscosity, and Ra appropriate for Mars, to calculate melting rates under Martian mantle conditions for different mantle heat production values, basal temperatures, solidii and crustal enrichment values. Their models demonstrate a decline in the rate of volcanism from >0.17 km^3/yr in the early Hesperian, to ~1e–4 km^3/yr at present, which is consistent with geological constraints on volcanism rates. This implies a decline in volcanic CO_2 flux from 8.8e7 kg/yr to 6.7e6 kg/yr over the same period. Assuming that the early Hesperian supported an atmosphere >1 bar, then the average loss rate of CO_2 was around 15 times greater than the influx during this period. A similar mismatch likely existed between H_2O loss and replenishment rates since the Noachian.

The fundamental imbalance between large atmospheric loss rates for Mars—primarily owing to its small size and non-existent magnetic field over much of its evolution—and the very low volatile replenishing rates in a sluggishly convecting stagnant lid regime, highlight the importance of surface tectonics for a planet's evolution. In all likelihood Mars was once a clement, perhaps habitable, planet but in the absence of prolific volcanism associated with surface plate creation, it could not sustain its surface atmospheric pressure.

It should be noted, however, that stagnant lid convection does not preclude extensive volcanism. In strongly convecting planets with hot interiors, the stagnant lid melt production rate can exceed that of an equivalent mobile lid system (e.g. Solomatov & Moresi 2000). A case in point is Io—a nominally stagnant lid moon with the highest internal heat production, eruption temperatures, or rate of volcanism of any body on the solar system (McKinnon *et al.* 2000). It primarily loses heat via volcanic resurfacing, demonstrating the efficiency of this mechanism, which has implications for both the early Earth, and for exosolar superearths.

EARTH AND VENUS: VISCOSITY AND THERMAL EVOLUTION

As the internal temperatures of a planet decline through time, not only do the internal velocities decline—as previously modelled in Figure 2 and by O'Neill *et al.* (2007a), but the internal viscosities also rise. Given the exponential, Arrhenius-type relationship between temperature and viscosity, this effect should in fact be dominant. This leads to the non-intuitive prediction that as we go back in time to a hotter early Earth, and the interior viscosities decrease, so do the convective stresses, as demonstrated by O'Neill *et al.* (2007b).

Decreasing convective stresses in the past have rather large implications for the Precambrian tectonic regime: it predicts a transit from mobile-lid to episodic convection at some point in Earth's past (O'Neill *et al.* 2007b). This differs from most previous work (e.g. Figure 2)—the primary difference being the change in viscosities through time, and their effect on convective stress—a factor not incorporated in O'Neill *et al.* (2007a; Figure 2). A large number of studies have highlighted the episodic nature of the Precambrian geological record, and while these will not be enumerated here the interested reader is referred to the compilations of Condie (1998), O'Neill *et al.* (2007b), Condie & Pease (2008) and Condie *et al.* (2009). Episodic Precambrian subduction events have a number of important implications. While the heat-flux in this regime is highly variable, with long periods of low heat flux interspersed with periods of very high heat flux during plate subduction and rapid seafloor spreading events, the average heat flux in this regime is similar to that of mobile lid regimes. However, mantle mixing in this regime is diminished compared with equivalent mobile lid models. The mixing problem is exacerbated in stagnant lid models, which are very inefficient (Van Keken & Zhong 1999). Neodymium isotopes suggest the survival of long-lived heterogeneities in the early mantle (Bennett *et al.* 2007), an observation that is difficult to reconcile with efficient mixing of the mantle in a vigorous plate tectonic regime. However, the mixing times in an episodic or even stagnant regime are consistent with the longevity of these mantle source regions.

In the case of Venus, such a thermal evolution implies that Venus may have once been in a stagnant lid regime, and subsequently evolved into an episodic regime. While geological observations of Venus' surface are insufficient to constrain its early evolution, its current atmosphere contains important clues. Radiogenic ^{40}Ar is present in Venus's atmosphere at concentrations of 25% of what should be there were Venus completely degassed. A similar situation exists for Earth, which possesses only 50% of its calculated ^{40}Ar budget. While other explanations have been proposed for Earth, including non-chrondritic ^{40}K concentrations, or the sequestering of K in the core, a primary factor in ^{40}Ar degassing is the dependence of volcanic degassing efficiency on tectonic regime. Kaula (1999) and O'Neill & Lenardic (2008) suggested that the transition from a stagnant-lid to an episodic regime on Venus, and an episodic to plate tectonic regime on Earth, could explain

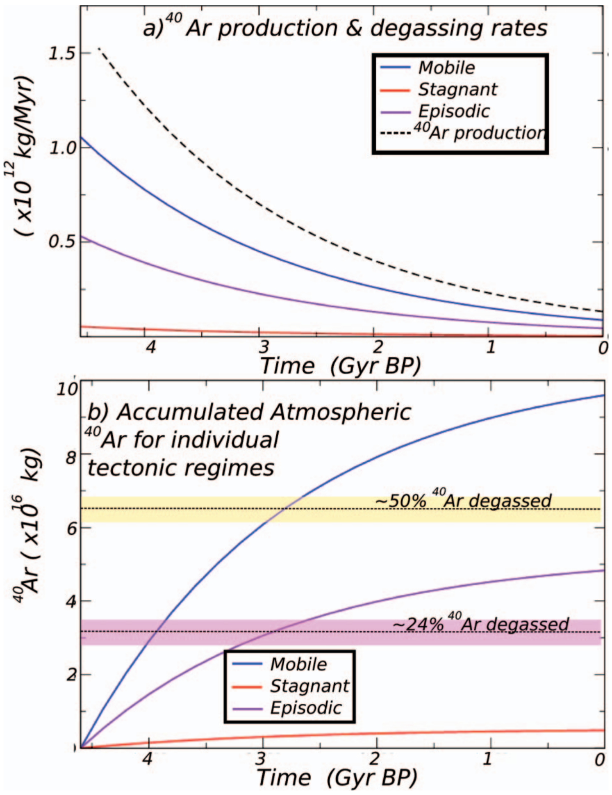

Figure 4 (a) ^{40}Ar production and degassing rates for the mantle (modified from O'Neill & Lenardic 2009). ^{40}Ar production in the mantle is calculated from the ^{40}K decay over time. Modelled using the parameters listed in Turcotte & Schubert (2002). Volcanic degassing is calculated from 2D viscoplastic convection models with basal Ra of 3e7, otherwise similar to Figure 1, and using a melting model described in O'Neill *et al.* (2005). Also shown are the relative degassing rates of ^{40}Ar, calculated using the tectonic efficiencies of numerical results. O'Neill & Lenardic (2009) found mobile-lid convection 1.8 times more efficient at degassing than episodic convection, and 20 times more effective than stagnant-lid for Earth-like parameters. Volcanism declines over time by a factor of $\sim$12 (from O'Neill *et al.* 2007c), and this is scaled to give a current rate similar to Earth. (b) Accumulated atmospheric ^{40}Ar over time for individual tectonic regimes, based on the degassing rates in Figure 4a. Also shown are atmospheric ^{40}Ar abundances of $\sim$50% that produced in the crust–mantle reservoir (applicable to Earth), and $\sim$24% of all ^{40}Ar produced (which is applicable to Venus, though the value is different when scaled to Venus).

the finite, diminished ^{40}Ar degassing efficiencies of both these planets (Figure 4).

The concept of planets evolving from hot, stagnant lid worlds, through to episodic subduction then finally cool mobile lid planets is a departure from the commonly viewed evolution of terrestrial planets. The common conceptual model posits that planets begin life hot and with active lid resurfacing. As they cool, they eventually become stagnant, as Mars is today. This is still the most probable fate of Earth as it continues to cool, but it has been little appreciated that stagnant lid convection is a natural consequence of hotter, higher Ra planets also. Outside computer simulations, Io again stands as an example of an extremely hot terrestrial body, in a stagnant lid regime, and is arguably a potential early Earth or Venus analogue.

ICY MOONS, FREEZING SUBSURFACE OCEANS AND TINY POWERHOUSES

Tidal interaction of closely orbiting moons in eccentric orbits around large gas giants, such as Jupiter or Saturn, imparts significant energy into the interior of these moons. This tidal heat production acts to power convection in the interior of these bodies, enabling active volcanism in the case of Io, or cryovolcanism and subsurface oceans in the case of Europa and Ganymede (McKinnon *et al.* 2000; Nimmo *et al.* 2002; Figueredo & Greeley 2004). Convection within these moons can potentially power surface tectonics, and many moons, notably Europa, Ganymede and Enceladus, exhibit evidence of active resurfacing not associated with cryovolcanic processes.

The Saturnian satellite Enceladus is particularly notable. This small satellite, 500 km in diameter, possesses a thin icy shell 92 km thick over a rocky interior (O'Neill & Nimmo 2010). The Cassini mission documented high heat flows and active ice geysers in the geologically active south pole region (Porco *et al.* 2006; Spencer *et al.* 2006; see Figure 5). The south pole terrain appears geologically young, certainly less than 100 Ma from cratering statistics, but potentially much less, with estimates down to 0.5 Ma—essentially reflecting ongoing geological activity (Porco *et al.* 2006). In contrast, the heavily cratered plains exhibit ages of *ca* 4.2 Ga, while the plains of Sarandib Planitia and Samarkand Sulcus record ages of 3750–170 Ma and 980–10 Ma respectively, depending on cratering rate (Porco *et al.* 2006). Intriguingly, the Cassini thermal imager measured heat flows of 5.8 ± 1.9 GW in the south pole terrain (Spencer *et al.* 2006). This is far in excess of the global heat production rate, of around 1.1 GW. The average heat flux in the south pole terrain is around 80 mW/m^2—far beyond what could be sustained by stagnant lid convection (O'Neill & Nimmo 2010). The heat flux is localised around four long subparallel fissures in the south pole terrain, dubbed the tiger stripes, which are also the locus of active geyserism (Porco *et al.* 2006; Figure 5).

O'Neill & Nimmo (2010) suggested Enceladus is in an episodic overturn regime. This would explain the heat flux disparity—the long term heat flux would be significantly lower than that currently observed, which is transiently enhanced owing to the current surface activity (Figure 6). It also explains the observed episodic age distribution, which is an intrinsic result of episodic partial lid recycling. For plausible Ra, internal heat generation rates and yield strengths, the recycling episodes would last approximately 10 Myr, and recur with a predicted periodicity of 0.1–1 Ga, consistent with the current pulse of activity and age distribution.

The story of Enceladus is an important one for two reasons. First, given Enceladus's 92 km thin convecting ice shell, it demonstrates that size is no obstacle to active tectonics, provided that the driving stress/resistive strength criterion is met. Second, it suggests

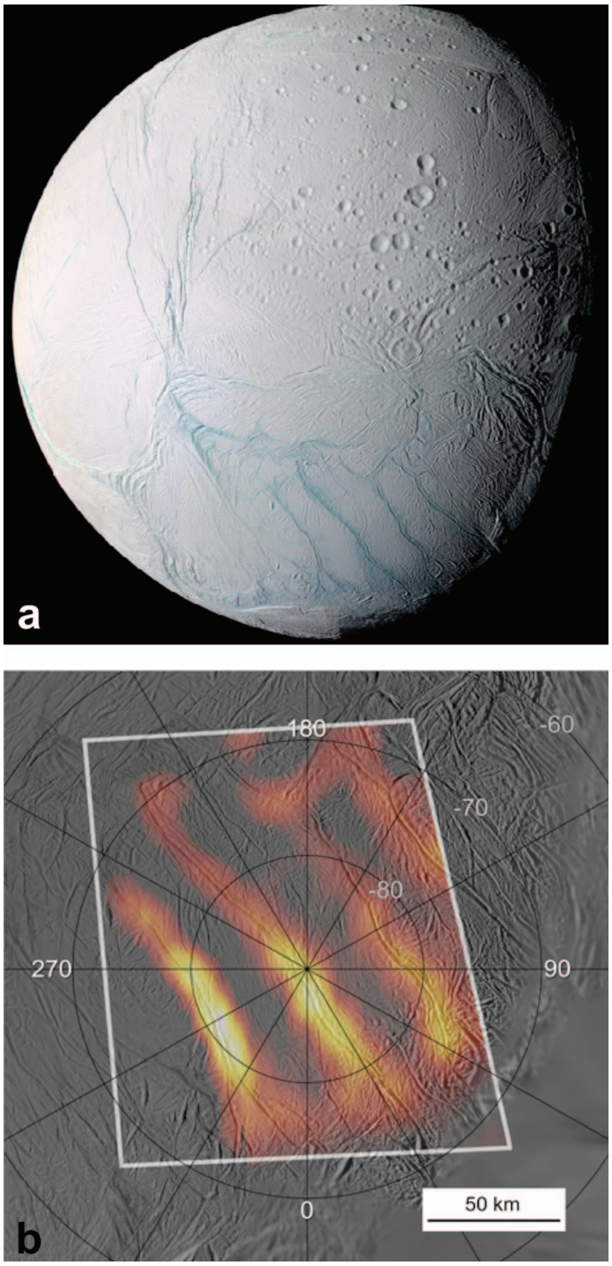

Figure 5 (a) Global image of Enceladus from the Cassini mission (Credits: NASA Image Gallery), showing the geologically and sparsely cratered south-pole terrane. (b) A view of the south-pole terrane from Cassini, overlayed with thermal imaging data showing heat flow and high temperatures are focussed on 4 sub-linear fractures known as the 'Tiger Stripes' (Credits: Nasa Image Gallery).

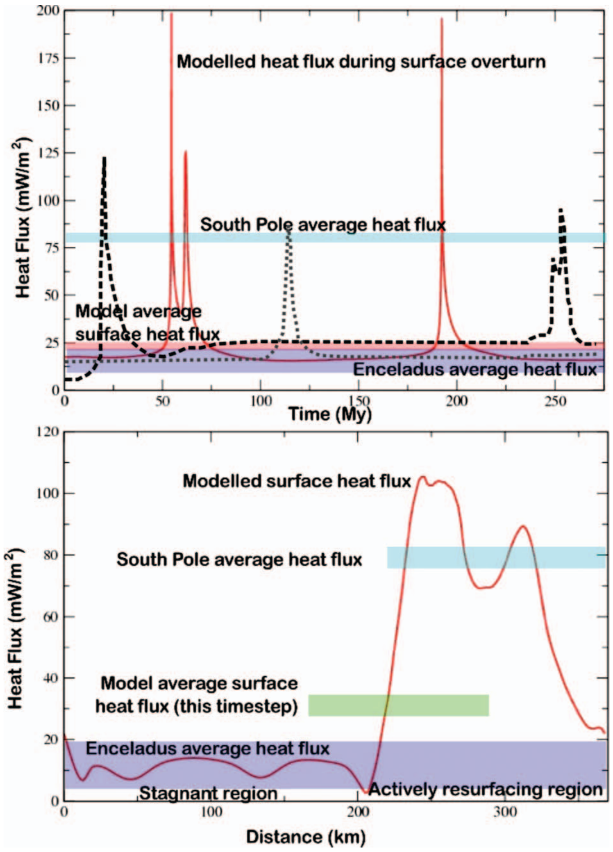

Figure 6 (a) Time evolution of episodic convection models at different aspect ratios (solid line, 1×1; dark dashed, 4×1; light dashed, 8×1; modified from O'Neill & Nimmo 2010). Active tectonics during the episodic subduction events results in pulses of heat flow far in excess of the background rate. The 1×1 model's average surface heat flux over the simulation is also shown (light rectangle). Enceladus's estimated globally averaged heat flux (7 mWm^{-2}, dark rectangle, labelled) and the present heat flux of the SPT (54–106 mWm^{-2}, labelled rectangle). (b) Transect of heat flux along the surface of a 4×1 numerical model, illustrating the regionally elevated heat flux of an actively overturning surface.

that internal temperatures on icy moons may fluctuate quite significantly, and the long-term existence of a subsurface ocean or water reservoir on Enceladus is not guaranteed. This has implications for other icy moons, too. Given both Europa and Ganymede exhibit evidence of resurfacing, the thickness of their subsurface oceans could fluctuate quite significantly in response to dynamics within the overlying icy shell. While it may seem fortuitous that Enceladus has been observed during an active phase, many other moons in the outer solar system, including not just Europa and Ganymede but also Rhea and Miranda, exhibit episodic age distributions or evidence of recent resurfacing, suggesting that they too are in an episodic regime, but happen to currently be in a quiescent phase. Active ice tectonics may be more common in the outer solar system than hitherto imagined.

SUPEREARTHS AND THE SIZE PARADOX

The success of exoplanet detection programs has led to the discovery of smaller, very probably terrestrial-like worlds, of less than 10 Earth masses (Beaulieau *et al.* 2006; Boss 2006). Too small to develop into gas or ice giants, these planets most likely have a rocky, iron-rich composition similar to the terrestrial planets, potentially with a large water budget (Valencia *et al.* 2006). More recently, observations of Corot-7b (Rouan *et al.* 2009) and Kepler-10b (Batalha *et al.* 2011), discovered by the Corot and Kepler missions respectively, have allowed derivations of their mass and radius, and thus also density. Kepler-10b's mass has been estimated at $\sim 4.56\ M_E$, and its radius at $1.416\ R_E$, giving it a density of $8.8 + 3.1/-2.9$ g/cm^3 (Batalha *et al.* 2011). Such a high density suggests the planet is predominantly of iron-silicate composition. Estimates of Corot-7b's density are hindered by stellar activity, which complicate calculations of its mass. Initial estimates placed it's density at 5.6 ± 1.3 g/cm^3 (Queloz *et al.* 2009)—similar to Earth's density—but this has been later revised both down (mass of $2.3 \pm 1.8\ M_E$;, Pont *et al.* 2011), and up (density of 10.4 ± 1.8 g/cm^3; Hatzes *et al.* 2011), placing its composition anywhere between a waterworld (water ice, rock and iron mix), and a predominantly iron planet. Collectively these planets confirm the existence of Earth-like rocky exoplanets. Speculation as to their surface conditions and potential habitability, though, hinges strongly on the question of their tectonic regime.

Early work based on simple scaling theory for stress variation with Rayleigh number suggested plate tectonics on these planets was inevitable (Valencia *et al.* 2007). However, this work did not consider variation in lithospheric strength with a planet's size, or adequately address basal/internal heating variations and thus internal viscosity constraints. O'Neill & Lenardic (2007) presented a suite of numerical models that showed that while convective stress does scale with gravity, the effect is modulated by increasing internal temperatures on larger worlds—provided the ratio of internal to basal heating is maintained. Additionally, they showed that the resistive strength of the lithosphere would increase for higher gravity, owing to pressure dependence of the brittle yield strength along faults. As a result, they presented a suite of superearths, which would exhibit stagnant lid behaviour at large planet sizes. As stressed in O'Neill & Lenardic (2007), the tectonothermal state of a planet depends not just on its size but also on its evolution, and a simple scaled up version of a modelled Earth is not at the point in its tectonothermal evolution as Earth is. Mobile lid convection is certainly permissible on superearths, but so are other tectonic regimes depending on the thermal state of these planets.

Van Heck & Tackley (2009) investigated the problem in 2D, using simple buoyancy variations to simulate larger planets, and found, similar to Valencia *et al.* (2007), that mobile lid convection is possible on these planets, but again the same caveats on the thermal state apply. Further work by Hansen *et al.* (2010), using fully dynamic convection simulations, demonstrated the importance of internal heating on superearths, and suggested that stagnant lid convection is the rule rather than the exception on such worlds. Kite *et al.* (2009) examined the effect of crustal production on such planets, and suggested that thick crust could in fact clog up the subduction system and also lead to stagnation of the lithosphere on these planets.

In the absence of observational constraints, the issue of the tectonothermal evolution of superearths is unlikely to be solved soon. Furthermore, variations in composition, phase change effects and pressure dependent physical parameters—particularly thermal expansivity and thermal conductivity (van den Berg *et al.* 2010)—are all likely to be critical to the discussion. The important point is the complexity of the tectonothermal evolution of terrestrial planets, and a reassessment of the assumption that large Earth-like compositions necessarily mean Earth-like behaviours.

CONCLUSIONS AND DIRECTIONS

While the tectonic and volcanic evolution of terrestrial planets play a pivotal role in their surface evolution and habitability, these processes are still inadequately understood. Ultimately, the planetary community needs an equivalent of the Hertzbrung-Russell diagram for stellar evolution. Stresses generated by interior convection within solid planets or moons are sensitive to viscosity variations, internal thermal state and pressure-dependent physical properties—as well as the fundamental buoyancy forces within the system. The effect of geometry (spherical *vs* Cartesian, 2 *vs* 3D), on stress scalings and tectonic regime has not been fully addressed. Future work ultimately needs to explore these factors systematically. While recently attention has been drawn to the effect of variable material properties under ultra-high pressures (van den Berg *et al.* 2010), their significance remains under-explored. Additionally, much greater emphasis needs to be placed on problem formulation, as recent controversy has largely revolved around inconsistencies with how a problem is cast (e.g. O'Neill & Lenardic 2007; Valencia *et al.* 2007). A case in point is Rayleigh number—a convecting system's induced stresses may either increase or decrease with increasing Rayleigh number, depending on whether the buoyancy terms or viscosity are altered, respectively. This non-uniqueness suggests some caution in the use of Ra as a unique system descriptor for terrestrial planets.

While numeric simulations have the potential to systematically map out a parameter space of tectonic behaviour, the combination of modelling results with observational constraints has the potential to vastly advance our understanding of planetary evolution. The isolated examples presented here demonstrate the capacity of numerical models to inform and consolidate geochemical and cosmochemical constraints, and vice versa. In the cases presented here simulations were used to calculate degassing rates, atmospheric evolution, crustal geochronology, timing and degree of melting, and tectonic responses. Observational constraints on these processes are conversely able to minimise the parameter space of the modelling—and this workflow has clear potential for bodies of which we have return samples, such as the Moon and Mars. The ability of simulations to simultaneously satisfy multiple observational constraints, and present a physically quantifiable framework for forwarding geochemical/cosmochemical models is a future direction with clear and attainable rewards.

Through recent outer solar system missions and the success of exoplanet detection programs, our observational base for planetary tectonics is expanding. However, observations alone are unlikely to fill the gaps in our understanding of planetary tectonothermal evolution. What is needed is an integrated approach combining our sparse observational database with extensive modelling suites to tackle the problem systematically—a daunting undertaking, but one that would ultimately overhaul our understanding of planetary evolution.

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