International Journal of Astrobiology 11 (4): 257–268 (2012)
doi:10.1017/S1473550412000122

Maintenance of permeable habitable subsurface environments by earthquakes and tidal stresses

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Abstract: Life inhabits the subsurface of the Earth down to depths where temperature precludes it. Similar conditions are likely to exist within the traditional habitable zone for objects between 0.1 Earth mass (Mars) and 10 Earth masses (superearth). Long-term cooling and internal radioactivity maintain surface heat flow on the Earth. These heat sources are comparable and likely to be comparable in general within old rocky planets. Surface heat flow scales with mass divided by surface area and hence with surface gravity. The average absolute habitable subsurface thickness scales inversely with heat flow and gravity. Surface gravity varies by only 0.4 g for Mars to 3.15 g for a superearth. This range is less than the regional variation of heat flow on the Earth. Still ocean-boiling asteroid impacts (if they occur) are more likely to sterilize the thin habitable subsurface of large objects than thick habitable subsurface of small ones. Tectonics self-organizes to maintain subsurface permeability and habitability within both stable and active regions on the Earth. Small earthquakes within stable regions allow sudden mixing of water masses. Large earthquakes at plate boundaries allow surface water to descend to great habitable depths. Seismic shaking near major faults cracks shallow rock forming permeable regolith. Strong tidal strains form a similar porous regolith on small bodies such as Enceladus with weak stellar heating. This regolith may be water-saturated within rocky bodies and thus habitable.

Received 8 February 2012, accepted 14 March 2012, first published online 16 April 2012

Key words: seismic regolith, Enceladus, crustal fault zones, superearth, Mars, tidally stress planets.

Introduction

The subsurface of the Earth is inhabited likely depths to where high temperatures preclude life. Sherwood Lollar *et al.* (2006) and Sleep & Zoback (2007) discussed the ecological problems facing such deep subsurface organisms, which utilize chemical energy sources provided by inorganic processes within the Earth and sources ultimately owing to photosynthesis at the surface. In reductionist terms, subsurface life harvests energy from chemical disequilibria to drive biologically beneficial chemical reactions in the direction of disequilibrium. The substrate must be sufficiently out of equilibrium to drive these beneficial reactions including additional energy lost to inefficiency (Hoehler *et al.* 1998). Life thus competes with abiotic processes that would eventually produce chemical equilibrium, especially at high temperatures where kinetics are rapid.

Sherwood Lollar *et al.* (2006) and Sleep & Zoback (2007) noted that a finite concentration of substrate (which depends linearly on bulk solid plus fluid concentration, as opposed to formal disequilibrium, which depends on the logarithms of fluid concentrations) is necessary to sustain organisms. Subsurface chemical energy sources, including the reducing power of serpentinite, and photosynthetic sources from the disequilibrium of buried organic matter and oxidants, are locally exhaustible. The 'right' range of fluid flow rates and hence the right rock permeability must exist for subsurface ecology to persist for geological time. Rapid fluid flow keeps the subsurface well mixed so the organisms quickly consume the available substrate such as yeast in a wine vat. Conversely, very slow fluid flow brings negligible substrate to an organism.

Furthermore, organisms need to reproduce to evolve. An environment that evenly supplies substrate in space and time, for example H_2 from radiolysis (Lin *et al.* 2005), is unfavourable, as the biomass will stagnate at its carrying capacity unless fluctuations as from rock fracture occur. Earthquakes locally produce suitable cracks within the shallow hard crustal rocks, allowing the mixing of water masses that are out of chemical equilibrium (Sherwood Lollar *et al.* 2006; Sleep & Zoback 2007). The biomass of organisms within the mixing zone quickly transiently increases from the bounty.

I address related issues in this paper to show that one can extrapolate from the habitable subsurface of the Earth to other rocky planets within a reasonable mass range of habitability. I begin by reviewing failure criteria for the formation of cracks in the subsurface and then proceed with scaling relationships with regard to planetary size for the deep biosphere. I then discuss rapid fluid movements within shallow fault zones and how geobiology provides information on this process. I finally discuss recent findings on the shallow subsurface porosity at shallow clement depths and apply them to the subsurface of small tidally heated objects. I concentrate discussion to water-based life within rocky planets for brevity. The formalism is easily extended to water-filled cracks within ice by replacing rock properties with ice ones. I do not attempt to deduce formalism for the mode of convection including the nature and existence of plate tectonics on the Earth through time or on other objects.

As a matter of semantics, the habitable upper few kilometres of the Earth are deep for a biologist but shallow for a geologist interested in internal planetary dynamics. The temperatures at the base of this habitable region ~100 °C are much greater than those at the surface but much less than those ~1350 °C in the asthenospheric mantle beneath plates or even ~400 °C at the base of seismic activity within continents. I qualify deep, hot, cool and shallow in terms of biology or geology when it aids understanding.

Physics of crack formation in the geologically shallow crust

Cracks provide habitable space and connected permeability. I present simple phenomenological theory of frictional cracking for use in the remainder of the paper and discuss stress levels from thermal buoyancy within the lithosphere.

Frictional failure

I begin with the familiar construct of friction. Cracking occurs when

$$\tau > \mu(P_R - P_F), \tag{1}$$

where τ is the shear traction (force per area on surface) driving slip, μ is the coefficient of friction, P_R is the rock pressure (force per area) closing the surface and P_F is the fluid pressure propping cracks open. Typical rock pressure is approximately the lithostatic pressure from the weight of overlying rocks. To compact notation, I let shallow rock density ρ_R be independent of depth. Lithostatic pressure is then

$$P_R \approx P_{lith} \equiv \rho_R g Z, \tag{2}$$

where g is the acceleration of gravity and Z is the depth. Fluid pressure P_F at habitable depth on the Earth is often close to the hydrostatic pressure from the weight of the overlying water,

$$P_{hydro} = \rho_W g Z, \tag{3}$$

where ρ_W is the density of water.

Intraplate forces and stresses

The upper part of the lithosphere fails in friction where (1) is applicable. The deeper depths within the lithosphere creep under ambient stresses. I review well-known dimensional scaling relationships for intraplate forces in the frictional lid that are applicable one-plate planets, such as Mars, and multi-plate planets such as the Earth.

I consider only thermal buoyancy on the grounds that chemical buoyancy associated with continents, including their crust and magnesium-rich lithosphere is ultimately a result of driving forces from thermal buoyancy (England & Bickle 1984). That is, forces from chemical buoyancy scale dimensionally to those from thermal buoyancy. I do not consider tidal stresses within this section, but later discuss tidal stresses in the section on regolith formation.

For brevity, I let density vary linearly with temperature and consider only the mantle as in an oceanic plate on the Earth. Lateral variations in weight (force per volume from buoyancy) depend on lateral density contrasts

$$\Delta B = g\Delta\rho = \rho_{ref} g\alpha(T - T_{ref}), \tag{4}$$

where $\Delta\rho$ is the lateral density contrast, ρ is the reference density of the mantle, α is the volume thermal expansion coefficient, T is the temperature at depth and T_{ref} is a reference temperature; T_{asth} of hot upwelling lithosphere at ridge axes is convenient for the Earth. The average temperature contrast across the lithosphere with the ridge-axis of the scales is the difference between the surface temperature T_{surf} and the asthenosphere temperature

$$\Delta B = \langle g\Delta\rho\rangle \approx g\alpha\rho_{ref}(T_{asth} - T_{surf}), \tag{5}$$

where French brackets indicate averaging over the lithosphere and the temperature contrast is redefined to be positive in the final equality as the sign of intraplate stresses is known.

To the first order, the mantle beneath the lithosphere behaves as a fluid so that isostasy applies. Strictly, the fluid pressure along an equipotential surface is considered to be constant. The elevation of ridge-axis relative to old oceanic crust is

$$\Delta E = \frac{g\langle\Delta\rho\rangle Z_{lith}}{g\rho_{comp}} \approx \alpha(T_{asth} - T_{surf})Z_{lith}, \tag{6}$$

where Z_{lith} is the old lithosphere thickness, ρ_{comp} is the density of the fluid isostatically compensating mantle essentially the mantle reference density, and the weight of ocean water is ignored for brevity to yield the final dimensional result. The difference in weight between the ridge-axis and the old lithosphere exerts a lateral force within the lithosphere that causes the ridge to spread and puts old ocean lithosphere under compression. The force acts over the thickness of the lithosphere Z_{lith} and scales with the lateral pressure difference $\Delta P = \langle\Delta\rho\rangle g Z_{lith}$,

$$S \approx \Delta P Z_{lith} \approx \rho_{ref} g\alpha(T_{asth} - T_{surf})Z_{lith}^2, \tag{7}$$

Only the uppermost strong lithosphere of thickness Z_{up} significantly supports this force. So the stress (force per area) in the strong layer is dimensionally

$$\tau_B \approx \frac{\Delta P Z_{lith}}{Z_{up}} \approx \frac{\rho_{ref} g\alpha(T_{asth} - T_{surf})Z_{lith}^2}{Z_{up}}. \tag{8}$$

The Coulomb ratio of intraplate stress to lithostatic stress at the base of the strong lid, where frictional strength from (1) to (3) and (8) is greatest, is

$$\frac{\tau_B}{(\rho_R - \rho_W)gZ_{up}} \approx \frac{\rho_{ref}\alpha(T_{asth} - T_{surf})Z_{lith}^2}{(\rho_R - \rho_W)Z_{up}^2}. \tag{9}$$

This quantity is not explicitly dependent on the planetary size. In particular, the ratio Z_{up}/Z_{lith} depends on the mechanical properties of rock with the temperature as discussed in the next full section on planetary size.

A difference between stable and active regions

Stable continental crust self-organizes differently from plate boundaries discussed in the section on the San Andreas Fault (e.g. Zoback & Townend 2001). The lithospheric lid beneath stable regions fails under friction from (1) with a coefficient of friction of ~0.6–0.7 under essentially hydrostatic fluid pressure. Within continents, the base of the frictional lid is 400 °C and ~16 km deep in the models of Zoback & Townend (2001). The frictional lid supports most of the available intraplate force, as significant stresses are required for failure. For example, letting rock and water density be 2700 and 1000 kg m^{-3}, the failure stress at 10 km depth for $\mu = 0.7$ is 120 MPa. The plate frictionally supports this stress essentially independent of the rate of strain. Only a small fraction of the available intraplate force drives creep in the ductile region beneath the frictional lid. The rate of ductile creep depends on the small stress in the ductile region, crudely to the third power. This slow ductile creep controls the macroscopic rate of lithospheric extension or compression.

Furthermore, Townend & Zoback (2000) show that a frictional failure within the lithosphere maintains permeability, and hence a near hydrostatic fluid pressure gradient. The net effect is that stable regions self-organize so that the lithosphere is strong or equivalently deforms very slowly under the available stresses. With regard to life, Sleep & Zoback (2007) showed that the rate of earthquakes at shallow depths associated with this process is enough to maintain permeability so that fluid flow is neither too slow nor too fast. Fine tuning is not necessary; two order of magnitude ranges of strain rate and permeability suffice. In addition, permeability continues below habitable depths in hard rocks, implies that high temperatures limit the depth of habitability. Similar reasoning with much less data applies to the crust of Mars.

Plate boundaries differ from stable regions in that the effective strength of the frictional lid is quite low <20 MPa (Fulton & Saffer 2009) and the ductile region beneath the frictional lid supports much of the available intraplate force. The ductile material beneath the surface plate boundary thus deforms rapidly. The mechanism of this self-organization is not fully understood. I defer discussion of mechanics and the habitability of plate boundaries to the section on the San Andreas Fault.

Deep biosphere scaling for rocky planetary size

Astronomers are likely to continue to find planets of various sizes within the habitable zone. I have already shown that the Coulomb ratio of intraplate stress to lithostatic stress in (9) is weakly dependent on planetary size. I apply simple scaling arguments to show that several properties of the shallow habitable surface also do not vary explicitly with planetary size and that the absolute thickness of the habitable region decreases inversely with planetary size expressed surface gravity.

I confine discussion to rocky objects, as the formalism carries through to icy ones. Astronomers infer planetary mass M from gravitational perturbations on nearby objects and measure radius R during transits. Considerable notational simplification for the shallow subsurface accrues by using surface gravity g as the parameter for planetary size. It is straightforward to apply rock physics to obtain density at high pressures and hence the relationship of mass to radius (Valencia *et al.* 2006; 2007a, b; Fortney *et al.* 2007; Seager *et al.* 2007; Sotin *et al.* 2007; Sasselov *et al.* 2008). For reasonably habitable rocky planets, gravity normalized to the Earth's value, varies from ~0.4 for Mars to 3.15 for a hypothetical 10 Earth-mass object (Valencia *et al.* 2006). Smaller objects cannot gravitationally hold a clement ocean and atmosphere for geological time; larger objects accrete too much gas and become like Neptune.

Beginning simple derivations with Gauss's law, surface gravity is proportional to planetary mass and inversely proportional to surface area

$$g = \frac{4\pi GM}{4\pi R^2}, \tag{10}$$

where G is the gravitation constant. Surface heat flow is similarly obtained from Gauss's law by conserving energy

$$q = \frac{Q_{\mathrm{total}} M}{4\pi R^2}. \tag{11}$$

where Q_{total} is the total effective heat source per mass including transient cooling of the planetary interior. Thus, the ratio of heat flow to gravity is

$$\frac{q}{g} = \frac{Q_{\mathrm{total}}}{4\pi G}, \tag{12}$$

which has the convenient property that it is not explicitly dependent on planetary size.

With regard to habitable conditions, the shallow thermal gradient is proportional to heat flow

$$\frac{\partial T}{\partial Z} = \frac{q}{k}, \tag{13}$$

where k is the thermal conductivity. The thermal gradient in terms of lithostatic pressure from (2) and (10)–(12) is similarly not explicitly dependent on planetary size

$$\frac{\partial T}{\partial P_{\mathrm{lith}}} = \frac{Q_{\mathrm{total}}}{4\pi G\rho k}. \tag{14}$$

where Q_{total} is again the sum of virtual heat sources per mass within planet.

The relationship between heat flow and gravity (12) indicates that planetary size has only a moderate effect on the former. The thickness of the strong lid of the lithosphere $Z_{\mathrm{up}} = k(T_{\mathrm{up}} - T_{\mathrm{surf}})/q$ (where T_{up} is the temperature at the base of the strong region) and the lithosphere thickness $Z_{\mathrm{lith}} = k(T_{\mathrm{asth}} - T_{\mathrm{surf}})/q$ vary inversely with gravity. So does the thickness of the habitable subsurface $Z_{\mathrm{hab}} = k(T_{\mathrm{max}} - T_{\mathrm{min}})/q$, where T_{max} is the upper limit for life and T_{min} is the minimum temperature for life or the surface temperature if the surface is warm enough to be habitable. The ratios of these depths from (9) do not explicitly depend on planetary size. As already noted, gravity varies but less than a factor of 10 over the range of plausible habitable planetary sizes from Mars to a 10 Earth mass object. Local heat-flow variations offset the significance of the systematic variation to some extent. An active planet such as the Earth has both stable regions with low heat flow and regions such as ridge axes with high heat flow. For example, the heat flow through 100-million-year-old oceanic crust is $\sim 0.05\ \mathrm{W\,m^{-2}}$, while the heat flow including hydrothermal circulation above the axial magma lens of very fast mid-oceanic ridges is $\sim 30\ \mathrm{W\,m^{-2}}$ (Fontaine *et al.* 2011).

The thermal gradient (14) depends on rock density that does not vary much. The thermal conductivity in water-saturated rock including sediments varies moderately. However, as discussed in the section on shallow porosity generation by tides, thermal conductivity in regolith in a vacuum can be quite low causing large shallow thermal gradients.

Effective heat source

The effective heat source per mass is more problematic in (14). It is formally the sum of contributions from radioactive heat generation, tidal dissipation and transient cooling of the planetary interior including the effects of phase changes

$$Q_{\mathrm{total}} = Q_{\mathrm{rad}} + Q_{\mathrm{tide}} + \frac{C\Delta T}{\Delta t_{\mathrm{planet}}}, \qquad (15)$$

where C is the specific heat per mass, which does not vary much for rock. The transient cooling term implies that planetary interior cools by ΔT over geological time $\Delta t_{\mathrm{planet}}$. The tidal term is negligible for the modern Earth. Astronomers are able to identify bodies with strong tidal heating. The radioactive and cooling terms are comparable, but their ratio is not precisely constrained (Korenaga 2008). The sum of mantle sources expressed as surface heat flow $Q_{\mathrm{total}}/4\pi R^2$ is $\sim 0.070\ \mathrm{W\,m^{-2}}$. For example, the widely used Bulk Silicate Earth model gives $Q_{\mathrm{rad}}/4\pi R^2$ of $0.024\ \mathrm{W\,m^{-2}}$ (McDonough & Sun 1995). The available temperature change of the interior from the early Earth with very high heat flow to an eventually dead planet with a frozen interior is only a few 100 K (e.g. Sleep 2007; Korenaga 2008). The estimates of the current cooling rate are compatible: ~ 50 K per billion years (Abbott *et al.* 1994; Galer & Mezger 1998) and ~ 100 K per billion years (Korenaga 2008). There is not enough available energy to maintain heat flows of a few times the current value for more than a few 100 million years even early in the Earth's history (Sleep 2007). Measurements of the flux of anti-neutrinos from the Earth's interior may eventually provide precise measures of current radioactivity (e.g. Araki *et al.* 2005).

That is, overall (14) is a guide to older planets. High temperature at depth precludes life and aids chemical reactions that close pore space. Pressure that directly closes pore space increases proportionally with temperature. Furthermore, the relationship for the Earth that radioactivity and transient cooling are comparable grossly applies to old extrasolar rocky planets. Radioactive heat production is potentially inferred from stellar spectra with the caveat that the method does not yield a precisely accepted value from the Sun for the Earth.

Conversely, the persistence and formation of pore space in terrestrial environments constrain extrasolar analogues. I discuss the San Andreas Fault as an active plate boundary in this regard in the next full section and regolith formation in the following section. I first discuss the effect of the absolute thickness of the habitable surface on survival of biota following large asteroid impacts.

Scaling of subsurface heating for large asteroid impacts

Planets are likely to be hit with large asteroids early in their history where their heat flows are quite high (Ryder 2002; Zahnle & Sleep 2006; Zahnle *et al.* 2007). A projectile with a diameter greater than 300 km significantly vaporizes the ocean leaving hot sterilizing surface conditions in its wake. The expected number of such impacts for the Earth is 0–4. Estimates of the actual number are subject to the vagaries of the statistics of small numbers, given the lack of Hadean record.

The deep subsurface is a potential refugium in the aftermath of surface-sterilizing impacts (Zahnle & Sleep 2006; Zahnle *et al.* 2007). The surface remained hot for the time required for the evaporated ocean to rain out. Heat escaped to space at the runaway greenhouse limit that is comparable to present solar flux. The net rate of rainfall was comparable to the current global average, ~ 1 m per year. Thus it took thousands of years for the Earth where the surface to return to clement conditions.

High temperatures penetrated by conduction into the subsurface during the period t_{hot} where the surface was hot to a depth Z_{hot} that scaled as $t_{\mathrm{hot}}^{1/2}\kappa^{1/2}$, where κ is the thermal diffusivity. This depth is ~ 1 km for the Earth. Life at shallower depths perished. Deeper regions with hot ambient temperatures were uninhabited to begin with. The current record for thermophiles 122 °C (Takai *et al.* 2008) provides guidance. That is, the subsurface formed a refugium only in regions where the thermal gradient was low enough and that the temperature at ~ 1 km depth was less than ~ 122 °C.

A survivable deep refugium may well have existed on the early Earth during late heavy asteroid bombardment (Zahnle & Sleep 2006; Zahnle *et al.* 2007). Analysis of the chemistry of Hadean zircon crystals indicates that local regions with low thermal gradients existed by 4.19 Ga (Hopkins *et al.* 2008). The computed heat flow through older oceanic crust became low enough for habitability at 1 km depth by ~ 4.37 Ga in thermal models (Sleep 2007). Still any deep refugium was likely hot so that only thermophile organisms survived.

However, only a few (if any) impacts of the required size occurred over several 100 million years. There was thus time of life to originate in a clement environment and to evolve colonizing the deep surface. (The detailed rate of evolution is beyond the scope of this paper.) Decimation of life in an impact left only thermophile survivors. This sequence is an attractive explanation for 'hot cross' molecular phylogenies were life originates cold but major clades including bacteria and archaea have thermophile roots (e.g. Mat *et al.* 2008). Given the lack of geological record, it is also conceivable that thermophile organisms evolved and later globally outcom-peted their low-temperature relatives, leaving an apparent

LUCA bottleneck without a sudden mass extinction (Miller & Lazcano 1995). Origin of life in a hot ambient climate that gradually cooled so that it was still above 70 °C in the early Archean is another viable interpretation of molecular phylogenies (Gaucher *et al.* 2008).

Equations (12) and (13) illustrate that ocean boiling impacts (if they occur) are more likely to sterilize large planets than small ones (Zahnle & Sleep 2006). Specifically, the thickness of the subsurface zone with habitable ambient temperatures scales inversely with planetary gravity. The time to heat the habitable thickness scales with $1/g^2$. The subsurface on Mars likely remained clement and was continuously habitable before the Earth's surface cooled. The hypothesis that life originated on Mars and then was transported to Earth on rocks ejected by modest asteroid impacts is thus attractive. Perhaps, ironically, it is more testable than life starting on the Earth, as the ancient geological record is well preserved on Mars, just hard to access.

Lessons from San Andreas Fault

Active fault zones are attractive habitable environments on rocky planets. The association of fluid circulation, hot springs and hydrothermal mineral deposits with crustal fault zones is well known. Furthermore extrapolating from Mars, faulting is likely even on relative small planets. In this section, I examine general biological implications of results from a borehole (San Andreas Fault Observatory at Depth, SAFOD) near Parkfield California that penetrated the active fault zone at ~2700 m depth. Data from the borehole on the bears on rapid fluid flow following earthquakes, crustal permeability and the recognition of inhabited and sterile environments.

This fault segment near Parkfield is on the boundary between much longer segments. To the north, the fault creeps peacefully, apparently without any large earthquakes. The fault segment to the south is currently locked. It failed most recently with ~5 m of slip in the 1857 magnitude 7.8 Fort Tajon event. The Parkfield segment experiences both creeping and earthquake behaviour. Earthquakes of magnitude ~6 occur every ~30 years, most recently in 2004 along with creep (e.g. Bakun *et al.* 2005).

Available subsurface substrates and environments

It is not clear whether microbes inhabit rock at ~2700 m depth of core recovery from the fault zone. The ambient temperature is about 110–114 °C (Schleicher *et al.* 2009). Possible substrates include reaction of oxidants in surface-derived water with ambient reductants in the rock and reaction of locally generated disequilibrium fluids during mixing. Conditions have not been static; potentially habitable rock is continually produced and destroyed. Fission track data from the upper 800 m of the section indicates ~50 K of heating associated with basin subsidence followed by 30 K of cooling due to exhumation in the last 4–8 Ma (Blyth *et al.* 2004).

Core samples collected from the fault zone at ~2700 m depth indicate reducing conditions. Photosynthetically derived organic matter is present in deformed sediments including shale, siltstone and sandstone. Fe(II)-rich serpentine and talc were also recovered (Bradbury *et al.* 2011). Potential oxidants include O_2 from surface waters and sulphate. Anhydrite $CaSO_4$ associated with calcite was obtained from the core (Holdsworth *et al.* 2011; Mittempergher *et al.* 2011). It is not clear whether sulphuric acid in water reacted with calcite, making anhydrite. In any case, this vein deposit crystallized from sulphate-bearing water. Given its fine-grained nature and recent formation, the fluid near the crystals is likely still saturated with respect to anhydrite. The straightforward explanation is that sulphate-reducing organic matter-oxidizing organisms do not inhabit zone of the recovered core.

Gas samples indicate the presence of local disconnected domains. Some of the samples contain H_2, possibly from serpentinization, and CO_2 (Wiersberg & Erzinger 2008, 2011). This mixture is a potential substrate from methanogens (e.g. Hoehler 2005) and acteogens (e.g. Hoehler *et al.* 1998). Again the straightforward explanation is that methanogens and acetogens are not present.

The occurrence of sulphate, however, is likely to be biologically moderated and requires rapid fluid flow at great habitable depths. I discuss a least astonishing hypothesis, which differs from the one where the observed veins were injected at high fluid pressure (Holdsworth *et al.* 2011; Mittempergher *et al.* 2011). Pressurized veins tend to ascend upward as water is less dense than rock. However, a source of deep sulphate-bearing fluids is unlikely. The solubility of anhydrite and thus sulphate decreases with increasing temperature and hence depth. The kinetics of abiotic sulphate reduction also increases with depth.

In contrast, near-surface groundwater does passively move downward into cracks suddenly opened by faulting. Still, there is no significant local source of sulphate in surface rainwater. An indirect sulphate source is plausible. Dissolved O_2 exists in the shallow groundwater. These waters descended rapidly in the aftermath of cracking during an earthquake. The water then encountered sulphide in sediments and hard rock. Sulphide constitutes a small fraction of the total available reductants, but a biological process could have well preferentially reacted dissolved O_2 with sulphide. I do not have a useful way to obtain the depth of sulphide oxidation. In contrast, slow infiltration of water is unlikely to form sulphate at great habitable depths. Dioxygen is out of equilibrium with organic matter-rich sediments from the surface downward and would soon be consumed. Hydrocarbon gases in these rocks (Wiersberg & Erzinger 2008, 2011) would also tend to react with O_2.

A second episode of rapid fluid flow allowed sulphate-bearing fluid produced in this manner penetrate to the apparently abiotic region of the recovered core. Sulphate is out of equilibrium with organic matter and would like O_2 react biologically given the time during slow infiltration at the inhabited depth of sulphide oxidation. Moreover, sulphuric acid from sulphide oxidation would be neutralized abiotically by divalent cations, as in serpentine and talc, hindering calcite dissolution.

Evidence for fluid flow and permeability

Heat flow studies along the San Andreas Fault relate to both slow and rapid fluid flow. Physically, the fault plane slides under stress generating 'frictional' heat. The excess heat flow from this process, however, has proven very difficult to resolve. Rather, heat flow measurements show that the San Andreas Fault is weak in shear when slip actually occurs whether by creep or during earthquakes.

Mathematically, the long-term rate of heat generation per area with the fault zone is slip rate $V_F \sim 30\,\mathrm{mm\,a^{-1}}$ ($10^{-9}\,\mathrm{m\,s^{-1}}$) times the shear traction τ_F. The expected excess heat flow above the fault $\Delta q_F \approx V_F \tau_F$ scales with this product. The predicted excess heat flow for fault strength 100 MPa at hydrostatic pressure and a typical rock coefficient of friction of 0.7 in (1) is $0.1\,\mathrm{W\,m^{-2}}$ which exceeds the typical local heat flow of $0.08\,\mathrm{W\,m^{-2}}$ and can be thus excluded. More sophisticated analyses have resolved little if any heat flow anomaly associated with the fault (d'Alessio *et al.* 2006; Fulton & Saffer 2009). Scatter in heat flow data is likely associated with lateral variations in thermal conductivity. Fulton & Saffer (2009) gave an upper limit of 20 MPa for the depth and slip averaged stress on the fault. Both creeping and non-creeping segments of the fault are weak (d'Alessio *et al.* 2006).

Measurements from the SAFOD borehole indicate near hydrostatic fluid pressure and hence finite permeability (Zoback *et al.* 2010). Still, fluid flow in the subsurface is unlikely to mask a real anomaly from frictional heat on the fault plane. Modest rates of fluid flow driven by topographic relief are permitted. The allowed Darcy flow ($\mathrm{m^3}$ of water crossing a $1\,\mathrm{m^2}$ surface) is $\sim 0.01\,\mathrm{m\,a^{-1}}$ (Fulton *et al.* 2004). The physical flow rate (relevant to solute transport) is the quantity divided by the porosity or $\sim 0.1\,\mathrm{m\,a^{-1}}$. It would take thousands of years for water to reach the depth of the recovered core by distributed infiltration. In this regard, Wang (2011) cautions that it is quite difficult to measure the pore pressure and permeability with borehole instruments if the permeability is extremely low. Wang (2011) in addition noted that disconnected gas-bearing domains indicate locally low permeability.

Rupture-tip cracking

The modern theory of fault rupture bears both on fault weakness and sudden opening of connected cracks during seismic events (e.g. Noda *et al.* 2009). Fault rupture in a major earthquake that nucleates on a highly stressed part of the fault as expected from (1) with a coefficient of friction $\mu > 0.7$. High stresses in accord with (1) with $\mu \approx 1$ exist at the rupture tip. Strong heating from friction at the rupture tip briefly expands the pore fluid increasing fluid pressure or melts rock on the fault plane. Both processes decrease frictional sliding stress to a low value during the majority of slip. Dynamic rupture may break through previously creeping material. Sterilizing temperatures are localized to within ~ 0.1 m of the fault plane over minutes and are not a meaningful hindrance to habitability.

The starting coefficient of friction at the rupture tip is ~ 1 so that the absolute tension exists on planes aligned 45° to the fault plane. Vagaries in the stress and kinematics of the rupture tip open macroscopic cracks. The SAFOD core included numerous generations of mineralized veins within the active fault zone (Bradbury *et al.* 2011).

Thus, the sudden flow of water to great habitable depths is reasonable in the aftermath of large earthquakes that break the surface. The surface fluids provide O_2 for sulphide oxidizers. Microbes may react with some of the sulphates thus produced with organic matter. The examination of the chemistry of sulphates and sulphides would shed light on the biotic aspects of this process.

On a more local scale in the fault zone, sudden cracking connects previously separated fluid domains. As in stable regions, there is likely disequilibrium in the mixed region. In addition, abiotic fluids containing sulphate or H_2 and CO_2 may suddenly upwell into cooler inhabited regions providing transient bountiful conditions.

Maintenance of shallow regolith

Clement regolith is a habitable environment that is produced by strong earthquakes and likely by strong tides on small extraterrestrial objects. Terrestrial data relate mainly to earthquakes. Tidal strains are easily detectable on the Earth but a tiny compared with strains near the elastic limit of rocks. I begin with cracking during strong earthquakes to show the very shallow subsurface self-organizes so it is typically near its elastic limit during large seismic events. I then extend this concept to regolith on bodies with strong tides.

Seismic regolith

Frequent earthquakes damage shallow rocks producing cracked regolith. I review some observations before discussing physics, following Sleep (2011).

Seismically damaged regolith exists near major faults. Macroscopically, Brune (2001), Wechsler *et al.* (2009) and Dor *et al.* (2008) studied the tendency of repeatedly damaged rock near faults to easily erode more easily by rain and runoff than distal intact rock. At outcrop scale, hard rock along ridge tops in the San Gabriel Mountains of California resembles rock fractured by blasting (McCalpin & Hart 2003). Granitic rock became friable saprock near the Elsinore fault in California (Girty *et al.* 2008). Macroscopic cracks and volumetric strain allow groundwater to percolate through this rock mass increasing chemical weathering (Replogle 2011). This process gives access of both microbes and fluids to the rock.

Direct evidence links seismic waves to immediate cracking in the upper tens of metres. Shaking triggers numerous shallow very small earthquakes recorded by Fischer *et al.* (2008a, b) and Fischer & Sammis (2009). The largest of these events produce brief extreme >1 g ground accelerations at the surface (Aoi *et al.* 2008; Sleep & Ma 2008).

With regard to fluid flow, strong shaking during earthquakes is well known to change groundwater pressure and permeability in shallow wells (e.g. Elkhoury *et al.* 2006). This effect is often attributed to shaking disrupting debris that clog choke points in groundwater flow paths and to transient groundwater pressure variations due to shaking (Elkhoury *et al.* 2006). Some such changes are conceivably related to rock damage on fractures (Chao & Peng 2009). The extent to which each process operates is not obvious. I concentrate on cracking as my data bear on it and because cracks open permeability in exhumed hard rocks.

Mechanically, exhumed originally stiff rocks self-organize so that the regolith just reaches failure in typical strong events (Sleep 2011). The free surface acts as a zero stress boundary condition. Mathematically the components of the stress tensor that produce shear on horizontal planes and compression and tension on vertical planes are zero. Non-zero horizontal shear stresses exist on vertical planes. Extensional and compressive stresses also exist on vertical planes. These stresses resolve to shear on inclined planes. Except for the special case of a nearly vertically propagating body wave, it is necessary only to consider components of the stress tensor that are non-zero at the free surface.

Further simplification accrues because damaging seismic waves typically have wavelengths much greater than the tens of metre thickness of regolith. Stresses in stiff rock at depth thus control the strain ε in the more compliant regolith. The stress in the regolith is dimensionally proportional to this strain

$$\tau_{\text{rego}} \approx \Omega\varepsilon, \tag{16}$$

where Ω is the shear modulus of regolith. (The traditional symbols for shear modulus G and μ are already in use.) The dimensional expression is readily modified into a calibrated equation if the wave-type and period causing the strain and the ground structure are known.

From (1) and (3) in analogy with (9), failure occurs when the ratio of regolith stress to effective stress exceeds the short-term coefficient of friction

$$\frac{\tau_{\text{rego}}}{(\rho_R - \rho_W)gZ} = \frac{\varepsilon\Omega}{(\rho_R - \rho_W)gZ} > \mu. \tag{17}$$

In terms of remotely measurable parameters on the Earth, the shear wave velocity is $\beta = \sqrt{\Omega/\rho_R}$. Equations (16) and (17) then become

$$\frac{\varepsilon\beta^2\rho_R}{(\rho_R - \rho_W)gZ} > \mu. \tag{18}$$

In practice, the density term varies much less than the shear wave velocity. Then the shear wave velocity at the failure criterion increases with the square root of depth. Sleep (2011) showed that limited available field data do satisfy the failure criterion in (18) for reasonable seismic strains.

Furthermore, initially stiff exhumed rock self-organizes so that it satisfies (18). The initially stiff rock (high Ω in (16, 17)) has high stresses under the imposed strain and fails by cracking. This cracking reduces the shear modulus somewhat so that the dynamic stresses are less in the next event. Failure with cracking occurs in each event until the stiffness at each depth just satisfies the criterion in (18). Subsequent events just reach failure. A few new cracks may replace those that healed in the interseismic interval. A caveat is in order for depths beneath the regolith where the imposed strains do not produce failure in (16) and (17) with the original hard-rock stress modulus. It also needs to be remembered that the shear modulus cannot go to zero at the free surface. Loose material held down by gravity has a finite stress modulus. Rock at the free surface may well have finite short-term strength under zero lithostatic pressure, that is, cohesion.

Tidal regolith

The gradient of gravity produces tides that transiently deform orbiting objects. As already noted stresses, strains and internal heating from tides are mild within the present Earth. Tidal heating is significantly larger than radioactive heating and transient cooling on other objects. The small (500 km diameter) Saturn moon Enceladus is an example (McKay *et al.* 2008). The heat flow in the active south-polar region is comparable to that of the Earth. O'Neill & Nimmo (2010) assumed the mean regional heat flow is 0.08 W m^{-2} in their models. Howett *et al.* (2011) reported $\sim$3 times the heat flow from measurements.

With regard to internal processes, the scaling relationship for thermal gradient in terms of pressure (14) is formally correct. Surface gravity on Enceladus is $\sim$0.01 g so internal heat sources from radioactivity and cooling over the age of the solar system provide only a negligible $\sim$1% of the heat flow and tides essentially the total amount. Conversely, the effective heat source per mass from tides can be orders of magnitude greater than radioactive heating within an old planet.

Tidal heating is also significant on the inner satellites of giant planets in our solar system. In the Jovian system, the innermost large moon Io is continuously volcanically active; tidal heating maintains a liquid ocean beneath the icy surface of Europa. Farther away, high tidal heating is likely on planets within the habitable zones of dim M-class stars.

I concentrate on Enceladus since it appears likely to have tidal regolith that affects habitability. My intent is to extrapolate biologically favourable conditions within rocky bodies of similar size, tides and stellar heat flux. Lithostatic pressure increases slowly with depth given the low surface gravity. Modest tidal stresses (assuming intact ice) in Enceladus are over 0.1 MPa (Collins *et al.* 2009), comparable to the lithostatic pressure ('cyrostatic' is pre-empted with a very different meaning) 1 km down. Tidal heating within Enceladus has changed as its orbit evolved. On occasions including now, its heat flow regionally approached Earth-like values, periods of episodic freezing of the interior occurred in between (O'Neill & Nimmo 2010).

Conversely, other classes of tidally heated objects are less interesting to biologists. Moderately small objects such as Io with tidal heat flows much greater than the Earth become uninhabitable, as their water eventually escapes to space. Small objects with enough stellar heating to have clement surfaces suffer the same fate. Mars size and larger objects can retain volatiles. However, ones with Io-like tidal heating have very rapid tectonics and volcanism that rapidly renews porous surface environments. There is no need to specifically consider tidal regolith independently of seismic regolith and porous volcanic rocks. Even more vigorous tidal heating could conceivably raise the surface temperature of a Mars size or larger object to clement conditions in the absence of significant stellar heating. Terrestrial tectonics then provide little insight, as internal planet-wide heat flow would have a few times greater than the terrestrial maximum, 30 mW m^{-2} observed above the magma lens of fast ridge axes (Fontaine *et al.* 2011). This clement 'optimum' would likely be unstable, as modest increase in tidal heating would drive the planet into a runaway greenhouse. It is not evident how orbital and tectonic processes could buffer a planet at this optimum over geological time.

Returning to Enceladus, the cyclic process is self-regulating so the cracked regolith forms if it does not already exist. It is not necessary to know the precise value of the coefficient of friction, failure occurs only when tidal stresses exceed the limit imposed by the actual coefficient. Starting with a cold object that did not dissipate tides by ductile creep, eccentricity and tidal stresses built up to where ice failed in friction. If stiff ice extended to the surface, failure began near the top where there is little lithostatic stress. Eccentricity continued to build up until cracking extended down to ~10 km depth at ~0.1 MPa tidal stress and an effective coefficient of friction of ~0.1. Tidal dissipation from the frictional failure warmed the interior ice. The warm interior of the south-polar region became buoyant enough to convect; I do not attempt to model ice tectonics. At about the same time, tidal dissipation in the warm interior became intense enough to damp orbital eccentricity eventually to a low value. The heat generated by tides in the active period escaped to space and the object again froze. The cold object did not dissipate tidal energy until its eccentricity grew again to the value producing stresses for the frictional failure.

Tidal heating produces gas venting and seafloor spreading-like activity near its South Pole (McKay *et al.* 2008). These active regions thus experience both large tide strains and likely occasional seismic waves from tectonic earthquakes in the icy shell. A key criterion of seismic regolith discussion applies to these strong internal tides; the wavelength of tidal deformation is essentially the size of planet, giving strain boundary conditions within thin regolith. Tides occur periodically over Earth days much more frequently than the seismic cycle that produces regolith on the Earth. In addition, seismic rupture once started may even cause large earthquakes within a similarly stressed quadrant of an object.

The mechanical conditions that form rock regolith apply within the icy near surface of Enceladus. Cold ice in the outer solar system, including Enceladus with a surface temperature of 70 K, is effectively a rock. Its starting coefficient of friction is ~0.6 and its thermal conductivity is 1.4 W m^{-1} K^{-1} similar to crustal hard rock with 2.4 W m^{-1} K^{-1} (e.g., O'Neill & Nimmo 2010). As with rock (Noda *et al.* 2009), fault-plane melting once rupture starts greatly reduces the effective coefficient of friction of active ice to ~0.1 (O'Neill & Nimmo 2010). In particular, regolith just reaches the frictional failure over its full depth range over each tidal cycle.

The self-organization of regolith within cold airless objects such as Enceladus is more complicated than beneath an atmosphere (or an interior ocean), because the thermal conductivity of highly cracked regolith in a vacuum is quite low. At constant heat flow q from below, the thermal gradient q/k in the regolith may be quite large. The object as with young asteroids heated by short-lived radioactivity evolves towards a cold low-conductivity lid and a more isothermal interior (e.g. McSween *et al.* 2002).

In practice, it is difficult to constrain the precise thermal conductivity structure of deep regolith. Mathematically, the thermal conductivity of a mixture is between the volume linear (Voight bound) average of the conductivities $k_{\mathrm{lin}} \equiv \langle k_i \rangle$ and the inverse (Ruess bound) average $1/k_{\mathrm{inv}} \equiv \langle 1/k_i \rangle$, where i indicates individual components. The limiting averages do not differ a lot if the components have similar conductivities. This is the case with ice, water and rock. The conductivity of a near vacuum in equilibrium with the vapour pressure of ice is not exactly zero as some molecules and blackbody radiation are present, but it is extremely small. The inverse average conductivity then can be quite small. Geometrically, the case corresponds to open cracks intersecting the direction of heat conduction. This result applies with a more complicated formula even if the material is isotropic (Hashin & Shtrikman 1963). Conversely, the linear average conductivity is only modestly below the intact ice or rock conductivity. This case corresponds to thoroughgoing solid structures in the direction of conduction and again applies in a more complicated way to an isotropic material (Hashin & Shtrikman 1963). The calculation of actual conductivities in vacuum significantly below the upper limit is not feasible unless the detailed geometry is known. Note that the thermal models of Enceladus history by Prialnik & Merk (2008) assume linear average conductivity, which precludes very low thermal conductivity regolith.

Data are available for very shallow regolith. Scientists report such results from lunar probes and the laboratory as thermal diffusivity $\kappa \equiv k/\rho C$, where C is the specific heat per mass and results from thermal radiation emission as thermal inertia $\Phi \equiv \sqrt{k\rho C}$. Their measured values of these quantities in vacuum regolith are orders of magnitude below the values for intact ice and rock and hence far below linearly averaged conductivities. I seek the order of magnitude of conductivity from such measurements. There is little intrinsic difference in volume-specific heat ρC between ice and rock. Furthermore, this quantity depends linearly on density that cannot vary a lot in the presence of gravity. I hence do not need to constrain volume-specific heat precisely to express results as conductivity. Lunar regolith (Langseth *et al.* 1976) and near-earth asteroid regolith (Delbo *et al.* 2007) with a conductivity k of 10^{-2} W m^{-1} K^{-1} provide some guidance for very shallow Enceladus regolith. Remote thermal emission measurements are available for Enceladus. Abramov & Spencer (2009) used thermal inertia measurement to infer a very shallow thermal conductivity of $10^{-3}\,W\,m^{-1}\,K^{-1}$. Howett *et al.* (2011) obtained even lower conductivities for the extreme near surface and noted that thermal conductivity increases with depth.

The conductivity of deeper regolith is less constrained by data. For guidance, conductivity in vacuum of meteorites that resided at some depth within asteroids is ~ 0.1 that intact rock and still well below a linear average (Yomogida & Matsui 1983).

Unreasonable results are obtained by assuming that very low thermal conductivity inferred for very shallow regolith continues to depth within Encleadus (Abramov & Spencer 2009). I provide a numerical example with thermal conductivity of $10^{-2}\,W\,m^{-1}\,K^{-1}$ and an Earth-like heat flow of $0.08\,W\,m^{-2}$. (It is not necessary here to speculate about the effect of solutes on the liquidus and solidus of Enceladus ice or on their detailed effect on the behaviour of warm ice.) The surface is ~ 200 K cooler than liquid water. Liquid water resides at 25 m computed depth. It would reside at 250 m depth in a lid with the asteroid-interior conductivity of $10^{-1}\,W\,m^{-1}\,K^{-1}$ and at 2500 m depth for the nearly intact ice conductivity of $1\,W\,m^{-1}\,K^{-1}$. Overall, tidal heat flows comparable to terrestrial heat flow prevent oceans from freezing to the bottom and allow liquid water within rock.

The physical properties of ice likely preclude very shallow regolith with extremely low thermal conductivity from continuing into warm ice (Abramov & Spencer 2009). First starting from the bottom near the melting point, warm ice in analogy with ice in glaciers creeps, which closes broadens grain–grain contact points increasing thermal conductivity. (Enceladus regolith at 100 and 1000 m depth is at the familiar pressure of terrestrial ice at 1 and 10 m depth, respectively.) Farther up, moderately cold ice has a finite vapour pressure. For reference, the vapour pressure of ice is still 7×10^{-4} Pa at 170 K (Bryson *et al.* 1974). Ice with finite vapour pressure sinters increasing its thermal conductivity.

The sintering is more efficient near vapour plumes from slush. These features have vapour pressures around 600 Pa (near the triple point of water) before they vent. This lithostatic pressure exists at ~ 60 m depth within Enceladus. Vapour pressure can then force shallow cracks open and maintain already open cracks. In such regions with high thermal gradients, ascending water vapour is likely to sinter the cold ice it ascends. In this regard, Abramov & Spencer (2009) state that the thermal conductivity even at the surface is much higher within the very active regions with vapour plumes than elsewhere.

Overall, cracking, ice deformation and sintering self-organize Enceladus regolith structure. Outside of active plume regions from top-down, low-thermal conductivity regolith exists at cold temperatures where cracks heal very slowly; moderate thermal conductive regolith forms where sintering occurs; and warm ice creeps rapidly preventing cracks from building up. From (18), cracking maintains the S-wave velocity near a critical value. By analogy to the Earth, cracks even in warm creeping ice persist for the days of the tidal cycle. This reasoning does not yield precise thermal conductivities and thermal gradients, but it is likely that the depth to warm ice is much greater than 25 m obtained from very shallow thermal conductivity and much less than 2500 m obtained for nearly intact ice.

With regard to life, the low conductivity of the regolith allows liquid water to exist near the surface within cracked ice. Vapour plumes may originate near the surface and entrain microbes and their chemical products. However, much of the cracked regolith on Enceladus occurs at temperatures too low to sustain terrestrial life.

Extrapolating from Enceladus to rocky bodies with similar stellar heating, small wet objects with strong tides are likely to have liquid-water filled cracks at modest depths. The key differences are that rock unlike ice does not thermally convect and rapidly creep to close cracks as its temperature approaches the freezing point of water. The temperature in water-saturated regolith thus increases downward as within seismic regolith on the Earth. Ice-saturated regolith (as in a rock glacier) and water-saturated regolith have thermal conductivities similar to wet terrestrial sediments that are only modestly below those of hard rock. Tidal cracking is still likely to produce very cold low conductivity regolith as on Enceladus, allowing the habitable zone to be closer to the surface. Ice within moderately warm pore space is likely to cause some sintering.

Returning to (14), habitable temperatures occur at much lower pressures in a strongly tidally heated small body than in an Earth-like object where tidal heating is now negligible. Strains from strong tides are very likely to cause the frictional failure at these shallow depths. Cracking in water-saturated rock is likely. Hydrothermal circulation then continually recycles substrate as on the Earth.

Discussion and conclusions

The Earth is logarithmically in the middle of the size range between 0.1 and 10 Earth masses where clement oceans and atmosphere are possible with the traditional habitable zone. The Earth's subsurface is inhabited, likely down to where temperatures preclude life. Stable regions on the Earth provide analogy with Mars-like bodies. Active terrestrial regions are analogous with those on extrasolar objects and on early Mars.

Physically, it is convenient to express planetary size in terms of surface gravity. This parameter varies from the Earth by only a factor of a few upward to the largest and downward to the smallest traditionally habitable objects. In turn, heat flow is usefully proportional to surface gravity in (12) for objects without tidal heating. Lithostatic pressure closes cracks at depth and makes it more difficult for earthquakes and seismic waves to open new cracks. The thermal gradient in terms of lithostatic pressure in (14) is not explicitly dependent on planetary size.

The absolute thickness of the habitable subsurface is inversely proportional to heat flow and hence surface gravity. The planetary average of the thickness of the habitable subsurface for large objects is only a factor of a few less than that of the Earth and the thickness for Mars-like objects a factor of a few greater. Regional variations of heat flow on the Earth are greater than this. The habitability of the Earth's surface thus provides extrapolation over the traditionally habitable size range. However, ocean-boiling asteroid impacts, if they occur, are more likely to sterilize thin habitable subsurfaces.

Weak tectonic processes within stable regions on the Earth self-organize in a way that maintains habitability. Substrates from disequilibrium of fluids residing in different rock types and substrates from surface fluids reacting with rocks at depth are both available and not rapidly exhausted. Fluids circulate slowly at near hydrostatic pressure. Small earthquakes transiently release batches of fluid that mix, providing transient bounty and the opportunity for evolution.

Buried organic matter is a product of past photosynthesis. This substrate will persist along with surface oxidants even if photosynthesis ceases. It was present on the Earth when photosynthesis generated sulphate and Fe(II), not O_2. Photosynthesis may well have existed long ago while Mars was wet (Hartman & McKay 1995). Its surface is now oxidized and out of equilibrium with its interior.

Fault zones within active regions rapidly move rock with potential substrate in and out of habitable environments. Sudden cracking during large earthquakes lets water penetrate deeply where surface O_2 is out of equilibrium with buried organic matter, sulphide and Fe(II). Sulphate-bearing water similarly moves quickly. As in stable regions, transient bountiful conditions recur. Relatively modest earthquakes (magnitude $\sim$6 in Parkfield) suffice to open cracks into the deep habitable zone.

Shaking near active faults maintains permeable seismic regolith on the Earth. The regolith self-organizes so that it exceeds elastic limits during episodes of strong shaking. Cracks sometimes allow rapid movement of water. Chemical weathering combined with shaking opens porosity within hard rocks.

Tidal regolith is likely within bodies with strong tidal heating. The thermal gradient in terms of pressure may be orders of magnitude greater than that within non-tidal objects. Small cold airless objects such as Enceladus self-organize so that the uppermost regolith has low thermal conductivity. Small rocky bodies with little stellar heating are favourable as water-saturated regolith can form.

Acknowledgements

This research was supported by the Southern California Earthquake Center. SCEC is funded by NSF Cooperative Agreement EAR-0106924 and USGS Cooperative Agreement 02HQAG0008. The SCEC contribution number for this paper is 1629. Francis Nimmo promptly answered questions on Enceladus. I benefited from discussions at the SPASA meeting including those with Britney Schmidt on Enceladus.

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