Geodynamics and Rate of Volcanism on Massive Earth-like Planets

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ABSTRACT

We provide estimates of volcanism versus time for planets with Earth-like composition and masses 0.25 - 25 $M_\oplus$, as a step toward predicting atmospheric mass on extrasolar rocky planets. Volcanism requires melting of the silicate mantle. We use a thermal evolution model, calibrated against Earth, in combination with standard melting models, to explore the dependence of convection-driven decompression mantle melting on planet mass. Here we show that (1) volcanism is likely to proceed on massive planets with plate tectonics over the main-sequence lifetime of the parent star; (2) crustal thickness (and melting rate normalized to planet mass) is weakly dependent on planet mass; (3) stagnant lid planets can have higher rates of melting than their plate tectonic counterparts early in their thermal evolution, but melting shuts down after a few Gyr; (4) plate tectonics may not operate on high mass planets because of the production of buoyant crust which is difficult to subduct; and (5) melting is necessary but insufficient for efficient volcanic degassing — volatiles partition into the earliest, deepest melts, which may be denser than the residue and sink to the base of the mantle on young, massive planets. Magma must also crystallize at or near the surface, and the pressure of overlying volatiles must be fairly low, if volatiles are to reach the surface.

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1. Introduction

Theory predicts rock-dominated planets in the range 1-10 Earth masses (e.g. Ida & Lin 2004). Planets in this mass range are now being detected (Rivera et al. 2005), but using mass and radius measurements to infer their bulk composition will be difficult (Adams et al. 2008). However, next-decade observatories, such as JWST, GMT and TMT may constrain atmospheric mass and perhaps composition.

Such worlds, if massive enough, may retain a nebular (H-He rich) atmosphere. Degassing of an accretion-heated magma ocean probably envelops the growing planet in a steam atmosphere. But what if this initial volatile envelope is lost, or never accumulates? Then atmospheric composition will reflect the balance between volcanic degassing and atmospheric loss, as on Venus, Earth and Mars – perhaps modified by biology. Loss rates vary between gases, so planetary atmospheres could be a mixture of gases left over from the initial atmosphere, and those replenished by volcanism. Arguably, we have a nearby example in Titan, whose nitrogen (∼95% of atmospheric mass) was released early, but whose methane (∼5% of atmospheric mass) may have been supplied by cryovolcanoes over geological time (Tobie et al. 2006).

Thermal emission phase-curves gathered from extrasolar planets can set bounds on atmospheric mass – which, if negligible for an Earth-like world, would require that the current rate of volcanic degassing be less than or equal to the current rate of atmospheric escape. On the other hand, spectral observations of atmospheric constituents with short photochemical lifetimes, such as SO$_2$, would require an ongoing source – most likely, volcanic degassing. Such observations would also rule out a thick ice or gas envelope on the planet, because high overburden pressure suppresses degassing.

Volcanism results from partial melting of the upper mantle. (Planetary mantles can cool convectively without volcanism - present-day Mercury is almost certainly an example). The fundamental requirement for partial melting to occur is that the adiabat must cross the solidus somewhere. Assuming that the adiabat is steeper than the solidus, this requires at a minimum that the potential temperature of the mantle $T_p$ exceeds the zero-pressure solidus of mantle rock (peridotite):

$$T_p = T_{m,r} - P_r \frac{\partial V}{\partial S} \geq T_{sol}(0),$$

where $T_{m,r}$ is the mantle temperature evaluated at some reference pressure $P_r$, $\frac{\partial V}{\partial S}$ is the adiabat, and $T_{sol}$ is the solidus, evaluated here at zero pressure.

On planets undergoing plate tectonics, the thickness of the crust (the crystallized melt layer) is a good measure of the intensity of volcanism. The rate of volcanism is the product of crustal

\[1\text{James Webb Space Telescope.}\] http://www.jwst.nasa.gov

\[2\text{Giant Magellan Telescope.}\] http://www.gmto.org

\[3\text{Thirty Meter Telescope.}\] http://www.tmt.org
thickness, plate spreading rate, and mid-ocean ridge length. The pressure at the base of the crust is the product of crustal thickness, the planet’s surface gravity, and crustal density (which is taken in this paper to be constant). A reasonable approximation to the pressure at the base of the crust is the integral of the fractional-melting curve from great depth to the surface (Figure [1]).

Venus and Mars lack plate tectonics: their mantles are capped by largely immobile, so-called ‘stagnant lid’ lithospheres, which cool conductively. Because mantle cannot rise far into the stagnant lid, melting can only occur if the temperature at the base of the stagnant lid exceeds the local solidus of mantle rock:

\[ T_{m,r} - (P_r - P_{\text{lith}}) \frac{\partial V}{\partial S} \geq T_{\text{sol}}(P_{\text{lith}}), \]

where \( P_{\text{lith}} = \rho_{\text{lith}} g Z_{\text{lith}} \) is the pressure at the base of the stagnant lid, \( \rho_{\text{lith}} \) is lithospheric density, \( g \) is gravity, and \( Z_{\text{lith}} \) is stagnant lid thickness. This is a more stringent condition than (1) if the adiabat is steeper than the solidus (Figure [1b]).

Here we use a range of melting models to compute the crustal thickness, and rate of volcanism, for massive Earth-like planets whose temperature evolves with time (§2). We use ‘Earth-like’ to mean ‘compositionally akin to Earth, and has an ocean’. We present results for both the plate tectonic (§3.2) and stagnant lid (§3.3) modes of mantle convection. We find that results differ greatly depending on the mode of convection. Massive Earths with plate tectonics will produce melt for at least as long as the age of the Galaxy. However, plate tectonics may not operate on massive planets for a variety of reasons. Plate tectonics involves breaking cold lithosphere and subducting it. Scaling from Earth shows that higher gravity favors subduction (Valencia et al. 2007). But subduction might never begin if the yield stress of old plate exceeds the stresses imposed by mantle convection, which fall steeply with increasing planet mass (O’Neill & Lenardic 2007). Melting sets up more hurdles for plate tectonics, which we rank in §4. We show that plate buoyancy is likely to be a severe problem, and may be limiting for plate tectonics. In §5, we relate our results to the soon-to-be constrained property of atmospheric thickness, and discuss the possible suppression of degassing (and, perhaps, melting) by the higher ocean pressures expected on massive Earth-like planets. Finally, in §6, we summarize our results; justify our approximations and model limitations; trace the implications of galactic cosmochemical evolution for heat production and planetary thermal evolution; and place our results in astrobiological context.

2. Model description and inputs

We use a model of internal structure (§2.1) to set boundary conditions for a simple, parameterized model of mantle temperature evolution (§2.2), which in turn forces a melting model (§2.3). Feedbacks from temperature to internal structure are weak and safely neglected. Via greenhouse-gas regulation of surface temperature, melting and degassing might feed back to mantle thermal evolution (e.g. Lenardic et al. 2008), but we neglect this. Throughout, we assume whole-mantle convection.
Rather than attempt to predict exoplanet properties solely from basic physics and chemistry, we tune our parameterized models to reproduce the thickness of oceanic crust (7 km) on present-day Earth (White et al. 2001).

2.1. Radius and mantle depth

Given our assumption of whole-mantle convection, we need to know only the mantle’s outer and inner radii. The crust is thin, so the top of the mantle is \( R \approx R_{\oplus} \). Constant-density scaling, \( R/R_{\oplus} = (M/M_{\oplus})^{1/3} \), is not an adequate approximation for massive Earth-like planets. Valencia et al. (2006) suggest the scaling \( R/R_{\oplus} = (M/M_{\oplus})^{-0.27} \). Seager et al. (2007) provide an alternative functional form for the mass-radius relationship. We use their equation (23) to determine planet radius (Figure 2). However, we take \( R/R_{\oplus} = (M/M_{\oplus})^{-0.25} \) as a reasonable scaling in our order-of-magnitude equations (13) – (14) and (21). To find the Core-Mantle Boundary (CMB) radius for the Seager et al. (2007) scaling, we take the core mass fraction \( f_{\text{core}} \) to be 0.325 and numerically integrate inward using a pure magnesioperovskite mantle composition, a 4th-order Burch-Murnaghan equation-of-state, and material properties following Seager et al. (2007). As expected, and similarly to the findings of Valencia et al. (2006), we find that core radius fraction (\( \approx 0.55R \) on Earth) decreases slightly with increasing mass, because of the high pressures to which the core is subjected. Core-mantle boundary pressure is calculated to be 1.5 Mbar for 1 \( M_{\oplus} \) (14% greater than the more refined estimate of Dziewonski and Anderson, 1981), and 2.9 (6.9, 14, 40) Mbar for 2 (5, 10, 25) \( M_{\oplus} \).

2.2. Thermal model

For a convecting planet with a mobile lithosphere, in thermal equilibrium, and heated solely by mantle radioactivity,

\[
Q = \frac{M_{\text{mantle}}}{A} \sum_{i=1}^{4} H_{0}(i)e^{-\lambda_{i}t} = Nu \frac{k(T_{m} - T_{s})}{d} \tag{3}
\]

\[
Nu \approx \left( \frac{g \alpha(T_{m} - T_{s})d^{3}}{\kappa \nu(T)Ra_{cr}} \right)^{\beta} \tag{4}
\]

\[
\nu(T) = \nu_{0}e^{(A_{0}/T_{m})} = \nu_{1}e^{(T_{m}/T_{\nu})}, \tag{5}
\]

where \( Q \) is lithospheric heat flux, \( M_{\text{mantle}} = M_{\text{planet}}(1 - f_{\text{core}}) \) is the mass of the mantle, \( A \) is the planet’s surface area, \( H \) is power per unit mass, \( i = 1-4 \) are the principal long-lived radioisotopes (\( {}^{40}\text{K}, {}^{232}\text{Th}, {}^{235}\text{U}, {}^{238}\text{U} \)), \( \lambda \) is the decay constant, \( t \) is time in seconds, \( Nu \) (Nusselt number) is the dimensionless ratio of total heat flow to conductive heat flow, \( k \) is thermal conductivity, \( T_{m} \) is mantle temperature, \( T_{s} \) is surface temperature, \( d \) is the depth to the core-mantle boundary, \( g \) is
gravitational acceleration, $\alpha$ is thermal expansivity, $\kappa$ is thermal diffusivity, $\nu$ is viscosity, $Ra_{cr}$ is the critical Raleigh number with value $\sim 10^3$, $\beta$ is 0.3, $A_0$ is activation temperature, and $T_\nu$ is the temperature change associated with a factor-e viscosity change (Schubert et al. 2001). In equation (5), the subscripts 0 and 1 are different reference viscosities for the otherwise equivalent expressions for $\nu(T)$.

We neglect the pressure dependence of viscosity (Papuc & Davies 2008), which cannot be fully captured by parameterized models. This is equivalent to assuming that the viscosity beneath the upper boundary layer determines the properties of the flow.

For the $H_i$ we use the canonical values given by Turcotte & Schubert (2002), who estimate that 80% of Earth’s current mantle heat flux is supplied by radioactive decay. Although Earth’s surface heat flux is well constrained, the fraction of the flux out of the mantle that is due to radiogenic heat production is not. Literature values vary from $\leq 0.2$ (Lyubetskaya & Korenaga 2007) to 0.8 (Turcotte & Schubert 2002), with low values increasingly favored (Jaupart et al. 2007; Loyd et al. 2007). All values $< 1$ require a cooling Earth. Neutrino observatories can in principle directly constrain mantle [U] and [Th], but existing observatories are poorly located for this task (Dye & Guillian 2008). For comparison, we show results obtained using radioisotope complements appropriate to ‘undepleted’ Earth (Ringwood 1991), CI chondrites (Anders & Grevasse 1989), and EH chondrites (Newsom 1995) (Table 1; Figure 6). Since we will use Earth’s observed oceanic crust thickness to tune mantle temperature, it is not particularly important to get the absolute values right. However, variability in $H_i$ could swamp any size signal in rates of volcanism. A useful rule of thumb is that a doubling a planet’s concentration of radiogenic elements makes it behave like a planet with double the radius (Stevenson 2003). Variability in $H_i$ could arise from galactic cosmochemical evolution, or stochastic cosmochemical variability. We discuss these possibilities in §6.3.

Our model does not consider tidal heating, which may be important for Earthlike planets on close eccentric orbits about low-mass stars (Jackson et al. 2008), nor does it take into account the energetics of the core (Nimmo 2007). Because the magnitude of core cooling is set by mantle cooling, core energetics are only important for mantle thermal evolution in Earthlike planets as $t \to \infty$ and $H \to 0$. Surface temperature is assumed to be constant, which is likely to be a good approximation except for close-in exoplanets with thin atmospheres in a 1:1 spin-orbit resonance (Ganesan et al. 2008).

For the other parameters in equations (3) – (5) we pick values used in classical thermal evolution models (e.g., Schubert et al. 2001) (Table 2). Equations (3) – (5) can be solved directly for $T$. It is then easy to find the mass dependence of temperature (§3.1; Figure 7).

Real planets are not like this; secular cooling significantly contributes to the heat flux at the bottom of the lithosphere, so at a given time a planet will have a higher internal temperature and a higher heat flow than thermal equilibrium calculations would suggest. Planets of different masses follow parallel cooling tracks (Stevenson 2003), and internal temperature is regulated by the
sensitive temperature dependence of mantle viscosity on temperature (Tozer 1970). A very simple model for mantle thermal evolution with temperature-dependent viscosity in plate tectonic mode is (Schubert et al. 2001):

\[
\frac{\partial T}{\partial t} = \frac{H}{c} - k_1(T_m - T_s)^{(1+\beta)} \exp \left( -\frac{\beta A_0}{T_m} \right),
\]

where

\[
k_1 = \frac{A k}{c d M_{mantle}} \left( \frac{\alpha g d^3}{\kappa \nu_0 R_{cr}} \right)
\]

and \(c\) is the specific heat capacity of mantle rock. As before, we neglect core cooling and the pressure dependence of viscosity. We integrate this model forward in time from a hot start \((T_m = 3273\, \text{K})\), using a fourth-order Runge-Kutta scheme.

Because of the exponential temperature dependence of convective velocity, the transient associated with the initial conditions is brief (Figure 9). The property of insensitivity to initial conditions requires both a hot start and whole-mantle convection. The thermal evolution model is not reliable for planets \(\ll 1\, \text{Gyr old}\).

The degree to which the mantle cools over geological time depends on the activation energy for deformation of mantle rock, which is poorly known. We take \(T_\nu \approx 43\, \text{K}\), for which \(A_0 = 7 \times 10^4\, \text{K}\), giving an order-of-magnitude viscosity change for each 100K temperature increment (although \(T_\nu\) could be as high as 100K) (Sleep 2007). This is conservative in terms of the effect of mass on temperature, because small values of \(T_\nu\) allow mass (heat flux) variations to be accommodated by small changes in temperature (Figure 3).

For planets in stagnant lid mode, we implement an analogous model, following the scaling in Grasset & Parmentier (1998):

\[
T_c = T_m - 2.23 \frac{T_m^2}{A_0},
\]

where \(T_c\) is the temperature at the base of the stagnant lid; plausible values lead to \((T_m - T_c) \ll T_m\). Then:

\[
Nu \approx \left( \frac{g \alpha (T_m - T_c) d^3}{\kappa \nu (T) R_{cr}} \right) \beta.
\]

Since \(T_c > T_s\), stagnant lid convection is less efficient at transporting heat than plate tectonics.

The thermal model is linked to the melting model through the offset \(T_m - T_p\), which is adjusted to obtain 7 km thick crust on today’s Earth. The required offsets are 741 K, 707 K, and 642 K, for the MB88, K03, and pMELTS models, respectively (Figure 5).

### 2.3. Melting model

Our interest is in the potential for volcanic degassing, because the atmosphere is the near-term observable. Therefore we equate shallow decompression melting to volcanism. In addition to lavas...
that are extruded at the surface, magmas that crystallize below the surface as intrusions — such as the sheeted dykes and gabbros of Earth’s oceanic crust — are assumed to degas fully and are counted as ‘volcanism’.

Deep within Earth’s mantle, high pressure suppresses melting. Because the solidus is steeper than the adiabat, an ascending parcel of mantle may suffer partial melting, with the extent of partial melting increasing as pressure decreases (Figure 1c). Beneath mid-ocean ridges, the mantle undergoes corner flow. Melt is generated in a prism with triangular cross-section, ascends buoyantly, and is focused to a narrow magma lens beneath the ridge (Figure 4).

Earth generates 34 km$^3$ yr$^{-1}$ crust of which 63% (21 km$^3$ yr$^{-1}$) is by isoentropic decompression melting at mid-ocean ridges (Best & Christiansen 2001). After four decades of intensive study, this is also the best-understood melting process (Juteau & Maury 1999). Petrological systematics require (Langmuir et al., 1992), and most melting models assume (Ghiorso et al. 2002), that the source magmas for mid-ocean ridge basalt melt fractionally or with small residual porosity, separate quickly, and suffer relatively little re-equilibration during ascent. For more massive planets, these remain robust assumptions. Buoyancy forces driving segregation are stronger and, because the pressure at which the solidus and adiabat intersect is at a shallower absolute depth, the ascent pathways are also shorter. Because of these attractive simplifications and because mid-ocean ridge melting dominates Earth’s crust production budget, we focus on mid-ocean ridge melting in this paper.

Isoentropic decompression melting pathways are distinguished by their values of $T_p$. Actual temperatures of near-surface magmas are lower because of the latent heat of melting, the greater compressibility of melts with respect to solids, and, usually less important, near-surface conductive cooling. All decompression-melting schemes are very sensitive to $T_p$ just above the zero-pressure solidus. That is because in this regime, increasing $T_p$ both increases the pressure at which melting first occurs (lengthening the ‘melting column’) and also increases the fraction of melting ($X$) suffered by the top of the melt column (Figure 1c; Figure 4).

With the above assumptions,

$$P_{crust} = - \int_{P_o}^{P_f} X(T, P) dP$$  \hspace{1cm} (10)

$$T(P) = T(P + \delta P) - \left( \frac{\partial V}{\partial S} \right) \delta P + \left( \frac{\partial X}{\partial P} \right) L$$  \hspace{1cm} (11)

$$T(P_o) = T_p + P_o \frac{\partial V}{\partial S},$$  \hspace{1cm} (12)

where $P_f = 0$ in the case of plate tectonics or $P_f = P_{lith}$ (the pressure at the base of the lithosphere).

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4This percentage understates the contribution of mid-ocean ridges to overall mantle degassing, because most of the volatile flux at arcs is just recycled from subducting crust.
in the case of stagnant lid convection; \( P_o \) is given by the intersection of the adiabat with the solidus, which is the point on the adiabat where \( X = 0 \); \( L \) is the latent heat of melting.

The simplest possible model for \( X(T, P) \) has a solidus of fixed slope, with contours of equal partial-melt fraction parallel to the solidus and equally spaced (Sleep 2007; Papuc & Davies 2008). This is not very realistic because the solidus is strongly curved at low pressure, and because \( \frac{\partial X}{\partial T} \) decreases sharply at moderate melt fractions. Instead, we use three popular (and more realistic) models for \( X(T, P) \). In order of increasing complexity, they are those of McKenzie (1984) as extended in McKenzie & Bickle (1988) (henceforth MB88), Katz et al. (2003) (henceforth K03), and (for plate tectonic models only) Ghiorso et al. (2002) (henceforth pMELTS). MB88 and K03 are similar in that they fit simple functional forms to experimental data, with MB88 more widely used although it uses fewer data. pMELTS is a state-of-the-art model of phase equilibria for compositions similar to Earth’s mantle (Ghiorso et al. 2002; Smith & Asimow 2005). We use the depleted-mantle composition estimated by Workman & Hart (2005) with 500 ppm water.

We also carried out calculations with the model of Langmuir et al. (1992) (henceforth L92), which is almost as simple as those used by Sleep (2007) and Papuc & Davies (2008); for brevity, we do not figure results obtained using this model. Representative results with these models are shown in Figure 5.

Each model is required to produce 7 km thick basaltic crust, which is the observed value on Earth (White et al. 2001), after 4.5 Gyr on an Earth-mass planet undergoing plate tectonics. We adjust the offset between potential temperature in the melting models, and the characteristic mantle temperature used in the thermal model, to obtain the observed crustal thickness.

### 3. Model output

Time-independent thermal equilibrium calculations, using Equations (3) – (5), are presented first (§3.1). Then we present time-dependent calculations which couple Equations (6) – (9) to our melting models (§3.2, 3.3).

#### 3.1. Thermal equilibrium

The simplest possible rocky planet model assumes thermal equilibrium. Radiogenic heat production equals lithospheric heat flux. The ratio of convective to conductive heat flux, \( N_u \), uniquely specifies the vigor of convection. With temperature-dependent viscosity, this increases in a defined way with increasing temperature.

For our thermal equilibrium model, we take the radiogenic-element concentrations given by Turcotte & Schubert (2002) (Table 1). It is then straightforward to find the required mantle temperature. This is greater for more massive planets because their decreased surface area/volume
ratio requires higher heat fluxes (and more vigorous convection) to dispose of the same heat flux. From equations (3)–(5), but neglecting variations in \( T_m \):

\[
\frac{Ra}{Ra_\oplus} = \left( \frac{(M/M_\oplus)}{(A/A_\oplus)} \frac{d}{d_\oplus} \right)^{\frac{1}{7}} \tag{13}
\]

which with \( R \propto M^{0.25-0.28} \) (Valencia et al. 2006) gives \( Ra \propto M^{\approx 2.45} \); using this simplification leads to

\[
\nu(T) \propto M^{-5/4} \tag{14}
\]

— which when inserted into Equation (5) gives a good fit to the results shown in Figure 7. Similar scaling arguments with \( R \propto M^{1/3} \) give \( \nu(T) \propto M^{-8/9} \), i.e. an approximately straight line on a linear-log graph of temperature versus mass (Figure 7).

### 3.2. Plate tectonics

As anticipated, for example by Stevenson (2003), mantle temperatures for planets of different masses follow parallel cooling curves—the difference in potential temperature due to increased mass changes little with time. Planets with \( M = M_\oplus = 2 \) (5, 10, 25) have potential temperatures 39 K (97 K, 146 K, 221 K) greater than Earth after 4.5 Gyr (Figure 8). The variation in calculated potential temperatures between the internal structure models of Valencia et al. (2006) and Seager et al. (2007) is always \(< 15 \) K, but the constant-density planet runs significantly (up to 100K) colder (Figure 7). This is unsurprising because it has a much larger surface area, but the same radiogenic-element complement.

Because in plate tectonics mode melting is controlled by \( T_p \) alone, we can relate the behaviour of large planets (of comparable age to the Earth) to that of earlier versions of the Earth. For example, a 4.5 Gyr-old 10 \( M_\oplus \) planet behaves like a 1.5 Gyr-old 1 \( M_\oplus \) planet, apart from corrections for gravity (which thins the melting zone and thus the crust) and surface area (the more massive planet will have a larger gross rate of plate production). This means that early Earth data is useful in understanding the geodynamics of massive Earths (§6.2).

Using thermal evolution calculations for the Seager et al. (2007) internal-structure model, we calculate melt production using the MB88, K03 and pMELTS models. Potential temperature increases monotonically with mass, so the pressure at the base of the crust also increases monotonically (Figure 5). However, the absolute thickness of the crust also scales as the inverse of gravity. In other words, although bigger planets run hotter, higher surface gravity moves the solidus and suppresses melting. For temperatures close to the solidus, the first effect dominates, and increasing planet mass increases crustal thickness (Figure 10). Young and/or large planets show the opposite trend, with crustal thickness decreasing as planet mass increases. Crustal thicknesses are within a factor of two of each other for 1 \( M_\oplus \) and 25 \( M_\oplus \) until 8.6 Gyr. After that, the ratio of crustal thickness diverges, as melting begins to shut down on the lowest-mass planets.
In the L92, MB88 and K03 models, and for planets of intermediate mass and with ages slightly greater than the Solar System, increasing mass has only a small (and negative) effect on crustal thickness. However, crust production per unit time increases with increasing mass, because more massive planets have more rapid plate spreading: see §5.1. (Figure 10b & c).

Finally we use the Smith & Asimow (2005) front-end to pMELTS, in 2-D (corner flow) mode. We use pMELTS throughout the predicted melting range, even though the model is only calibrated for use in the range 1-3 GPa. We assume continuous melting with a residual porosity of 0.5% (that is, melt fractions greater than 0.5% are evacuated from the melting zone), and the mantle composition inferred to underly Earth’s mid-ocean ridge system (Workman & Hart 2005) (Figure 10d).

The pMELTS model predicts crustal thickness will increase sharply with increasing planet mass for massive planets with ages comparable to the Solar System. Potential temperatures for these planets are > 1500 °C. Therefore, the crustal-thickness result should be treated with suspicion, because when $T_p > 1460$ °C, pMELTS predicts melting at pressures greater than those for which it has been experimentally calibrated (< 3 GPa). Melting is still fairly robust after 13 Gyr in the pMELTS model, even for masses of 1 $M_\oplus$. This is because pMELTS ramps up slowly to 7 km crustal thickness as temperature increases (Figure 5), so isopachs are spaced more widely in temperature (equivalently, time) than with the other models. As with the other melting models, the planet mass that produces the thickest crust at a given time increases as the planets age.

In summary, mass dependence increases with time as planets cool toward the solidus, and the sign of mass dependence changes with time. This is because of the strongly nonlinear behaviour of melt production near the solidus; melting goes from zero to significant over a small range in potential temperature. Low-mass planets approach this temperature by 7-8 Gyr. Their crustal thickness declines more rapidly than on high-mass planets.

With a constant-density internal model (not shown) crustal thickness is very insensitive to increasing mass, as anticipated by Stevenson (2003). With our more realistic internal models, crustal thickness tends to decrease with increasing mass. This is because the increasing melt fraction is more than compensated by increased gravity — the absolute depth of the melt column is greatly reduced.

### 3.3. Stagnant lid

Because stagnant lid convection is less efficient at transferring heat than plate tectonics, a planet in which plate tectonics is suddenly halted will heat up, as shown by the thin solid line in Figure 11. This temperature rise reduces mantle viscosity, so $Ra$ increases. Temperature converges on the evolutionary track of a planet that has always been in stagnant lid mode, with a characteristic convergence timescale of $(H \Delta T_{mode})/cM_{mantle} \sim 1$ Ga. A similar argument explains the temperature changes associated with going from stagnant lid mode to plate tectonics. The
temperature difference between the tracks is $\Delta T_{\text{mode}} \approx 160$ K for all masses. This is roughly $T_0 \ln \left( \frac{(T_m - T_s)}{(T_m - T_c)} \right)$, which can be understood by equating the RHS of Equations (4) and (9).

As beneath Earth’s continents, melt production within an ascending column of mantle in stagnant lid mode is truncated at the base of the lithosphere. Consequently, the predicted erupted thickness is much smaller for the same mantle temperatures (Figure 12).

To find crustal thickness, we use the same mantle-potential temperature offsets as in plate tectonic mode (so the model is still ‘tuned to Earth’). We calculate crustal thickness as the integral of melting fraction from the surface to great depth. In plate tectonic mode this corresponds to the geophysical crustal thickness. But in stagnant lid mode, the geophysical crustal thickness need not be the same everywhere (because melting will only occur above mantle upwellings). Also, the geophysical crustal thickness may be everywhere greater than the integral of melting fraction from the surface to great depth, because crust generated in previous melting events is located directly underneath ‘fresh’ crust. (This situation is described in the context of Io by Moore (2001)). Therefore, the rate of volcanism (§5) is a more meaningful quantity than crustal thickness for comparing plate-tectonic and stagnant-lid planets.

Most planets run hotter in stagnant-lid mode to the extent that pMELTS cannot be used, as $T_p$ exceeds the range over which it is calibrated. The MB88 and K03 models show roughly the same behavior in stagnant lid mode (Figure 13). For young (<2-3 Gyr) planets, an ascending column of mantle produces more melt in a stagnant lid mode than in plate tectonic mode – the higher temperature matters more than the (small) lithospheric thickness. For somewhat older planets, an ascending column of mantle produces less melt in stagnant lid mode than in plate tectonic mode. The temperature difference is much the same, but the growing lithosphere increasingly truncates the melting column. At a mass-dependent age much less than the age of the Galaxy, melting ceases. With the MB88 model, our calculations show that volcanism on a planet in stagnant lid mode and with Venus’ mass ($M_\oplus = 0.85$) will cease after 3.9 Gyr, which is consistent with observations.

### 3.4. Cessation of melting

For a given melt production function, all planets in plate tectonics mode will cease volcanism at the same potential temperature. Consequently, a planet’s volcanic lifetime is delineated by an isotherm ($T_p = 1080 - 1193$ °C, depending on the melting model). For planets in stagnant lid mode, this is not the case. There is still a one-to-one relationship between temperature and the absolute thickness of the lithosphere. However, the absolute thickness of the melt zone at fixed temperature decreases with increasing gravity, so there is a temperature range for which low-mass planets can sustain melting in stagnant lid mode, whereas high–mass planets cannot. Over the mass range we consider, this temperature range is 180K. Consequently, in stagnant lid mode, more massive planets run much hotter but cease melting only moderately later than smaller planets (Figure 13).
Approximating stellar main-sequence lifetime as $M_*^{-3}$, we can compare stellar main-sequence lifetime to planet volcanic lifetime (Figure 14). Also shown are the curves for plate tectonics, though none of our massive-Earth models predict that plate tectonics should have ceased yet, even for planets that formed early in the Galaxy’s history.

When decompression melting of mantle material at background temperatures cannot sustain volcanism on Earth-like planets undergoing whole-mantle convection, other mechanisms may take over. Plumes rising from the core-mantle boundary, or from a compositional interface within the mantle, can have temperatures up to several hundred degrees greater than that of passively upwelling mantle. This requires a more detailed treatment of thermal evolution than we have taken in this paper, and could be the basis of a more detailed study.

Transitions between plate tectonics and stagnant lid mode have an impact on thermal evolution comparable to that of mass over the range of masses considered in this paper. A planet in stagnant lid mode may melt a smaller fraction of each ascending column of mantle, but the faster convection associated with its higher internal temperature may permit a greater rate of volcanism to that on a similar planet in plate tectonic mode (§5.1). With this motivation, we now assess whether plate tectonics is viable on more massive Earth-like planets.

4. Limits to plate tectonics

Transitions between stagnant lid mode and plate tectonics mode are reviewed in Sleep (2000). Contrasting predictions have been made about the existence of plate tectonics on massive earth-like planets; that plate tectonics is inevitable (Valencia et al. 2007), and that plate tectonics is unlikely (O’Neill & Lenardic 2007). Both studies considered the initiation of subduction, a necessary condition for the initiation of plate tectonics, but one that is not well understood (Hansen 2007; Sleep 2008). The range of mantle states that permit plate tectonics to continue will differ from the range of states permitting plate tectonics to initiate, and ongoing plate tectonics will alter the state of the mantle. Some studies (e.g., Hansen, 2007) suggest that plate tectonics must initiate early on, or not at all.

Here we exploit our comparatively robust understanding of melting to discuss some necessary, but perhaps more easily satisfied, requirements for maintaining Earthlike plate tectonics (assuming that it has somehow begun). Vertical (heat-pipe) tectonics, as seen on Io (Moore 2003; Lopes & Spencer 2007) may be thought to take over from horizontal (plate) tectonics when either:– (1) the thickness of the crust becomes comparable to that of the lithosphere; (2) the lithosphere is underlain by a layer with a significant melt fraction; (3) most of the planet’s energy is lost by magma transport rather than conduction, or (4) the crust delaminates. We also assess the likelihood that (5) vigorous plate tectonics on high-mass planets self-limits through more vigorous continental growth, and that (6) greater plate buoyancy prevents subduction. Figure 15 is a sketch of these processes. We present results only for our fiducial calculation with MB88 melting and
TS02 heating.

We assume throughout that an ocean is present on the planet. Weakening the lithosphere by hydration is thought to be a prerequisite for plate tectonics.

4.1. Crust thicker than lithosphere

If the crustal thickness $Z_{\text{crust}}$ is comparable to the lithospheric thickness $Z_{\text{lith}}$, the lower crust is likely to melt and form buoyant diapirs. Widespread intracrustal diapirism (vertical tectonics) within the oceanic crust is not known on Earth and, if it were a major heat sink for the mantle, would be incompatible with plate tectonics. $Z_{\text{lith}}$ scales as $Q^{-1}$, and on Earth the equilibrium value of $Z_{\text{lith}}$ is computed to be 110 km (McKenzie et al. 2005). Therefore,

$$
\left( \frac{Z_{\text{crust}}}{Z_{\text{lith}}} \right) = \frac{7}{110} \left( \frac{Q}{Q_{\text{Earth}}} \right) \left( \frac{Z_{\text{crust}}}{Z_{\text{crust,Earth}}} \right).
$$

We find $Z_{\text{crust}}/Z_{\text{lith}} < 1$ for all planets > 2 Gya, so intracrustal diapirism is unlikely within equilibrium lithosphere. What about younger lithosphere, where $Z_{\text{lith}}$ is smaller? If only half-space cooling is considered, then we would expect the crust to be partially molten for > 1 Myr for 15 km thick crust. However, hydrothermal cooling of young oceanic crust (Maclennan et al. 2005) along cracks whose penetration depth scales as the inverse of temperature (Vance et al. 2007) is efficient on Earth, and probably comparably efficient on massive Earth-like planets, if we assume that an ocean is present. Intracrustal diapirism is unlikely to be the limiting factor for super-Earth plate tectonics.

4.2. Melting at depth

Earth’s oceanic lithosphere may be underlain by a layer with a small fraction of partial melt, but the melt fraction is not high enough to erupt (except possibly under special circumstances). Significant melting at depths greater than the equilibrium lithosphere would lead to widespread intraplate volcanism, in contrast to volcanism on Earth which is largely (> 95% by volume) collimated along plate boundaries. We define ‘significant’ melting as $X > 0.1$. Significant melting at depth is predicted to have ceased after about 2 Gyr (2.5 Gya) on Earth, but to continue for about 4 Gyr on a 25 $M_\oplus$ planet. There is rather little variation in this timing between different melting models.

For this criterion to make physical sense, depleted material beneath the crust must be replaced by more enriched material through sublithospheric convection (Korenaga & Jordan 2004). Only then would we expect to see volcanism beneath old plate. This requires that the plate reach a sufficient age for sublithospheric convection to refresh the mechanical boundary layer. The neccessary
age is uncertain; gravity and bathymetry data yield a value of 90-130 Myr \cite{Crosby2006}, but this analysis is contested \cite{Korenaga2008}. Whether this age is achieved depends on plate velocities and ridge-to-trench lengths, but it is broadly unlikely on more massive planets, which will refresh their lithosphere more quickly than does Earth. Melting at depth is unlikely to be a long-term limit to plate tectonics.

4.3. Magma-pipe transport energetically trumps conduction

On Earth, heat lost by conduction through thin lithosphere near mid-ocean ridges greatly exceeds heat lost by advection of magma. On Io, the opposite is true: most internally-deposited heat is lost by magma transport. Such a planet, although it may have limited plate spreading, is not in plate tectonic mode; vertical rather than horizontal motion of the material making up the lid is the more important process.

We introduce the dimensionless ‘Moore number’\footnote{In appreciation of the work of Moore \cite{Moore2001, Moore2003}.} which we define to be the ratio of magma-pipe heat transport, $Q_{\text{magma}}$, to heat lost by conduction across the lithosphere, $Q_{\text{cond}}$ (analogous to a Peclet number). To calculate how this scales with increased mass, we specify that total heat loss (magma pipe transport plus conduction across the boundary layer) adjusts to match the heat loss across the boundary layer prescribed by the thermal evolution model:

$$Q = Q_{\text{magma}} + Q_{\text{cond}}$$ (16)

An upper bound on magma-pipe transport is to assume that all melt crystallizes completely and cools to the surface temperature. In that case:

$$Q_{\text{magma}} = \rho_{\text{crust}}(c_b\Delta T_1 + E_{lc})Z_{\text{crust}}l_rS \sim c_1s.$$ (17)

Provided that half-space cooling is a good approximation to the thermal evolution of plates,

$$Q_{\text{cond}} = 2k\Delta T_2 \left( \sqrt{\frac{A_{oc}l_r}{\pi\kappa}} \right) \sqrt{s} \sim c_2\sqrt{s},$$ (18)

where $A_{oc}$ is the area of the ocean basins only, which we take to be $0.6 \times A$ as on Earth. We can now solve the quadratic in $s$ —

$$Q^2 = c_1^2s^2 + c_2^2S.$$ (19)
Finally we obtain an expression for $Mo$:

$$Mo = \frac{Q_{magma}}{Q_{cond}} = \left( \frac{\rho_{crust} (c_b \Delta T_1 + E_{lc}) Z_{crust} l_r}{2k \Delta T_2 \sqrt{Al} / (\sqrt{\pi \kappa})} \right) \sqrt{s}. \quad (20)$$

Here $l_r$ is ridge length and $s$ is spreading rate. We assume a latent heat of crystallisation $E_{lc} = 550 \text{ kJ/kg}$, a specific heat of basalt $c_b = 0.84 \text{ kJ/kg/K}$, crustal density $\rho_{crust}$ of 2860 kg/m$^3$ (Carlson & Herrick 1990), base lithosphere temperature of 1300 °C, and lava temperature of 1100 °C and surface temperature of 0 °C giving $\Delta T_1 = 1100 \text{ K}$ and $\Delta T_2 = 1300 \text{ K}$. This gives $4.2 \times 10^{18}$ J in melt/km$^3$ plate made. As an illustration, for Earth (mid-ocean-ridge crust production 21 km$^3$ yr$^{-1}$), $Q_{magma}$ is 2.8 TW. The total oceanic heat flux is 32 TW, so today’s Earth has a $Mo \sim 0.10$.

Therefore, given $T_p$, we can solve for spreading rate (Figure 17). As $Mo \to 0$, $s \to Q^2$ (Sleep & Zahnle 2001). Notice that the characteristic age at subduction $\tau = A/(2l_r s)$. As a consequence, the Moore number does not vary with different possible scalings of ridge length (equivalently, plate size) with increased planet mass. Because not all crustal material cools to the surface temperature nor crystallizes completely, the correct spreading rate is intermediate between that assuming $Mo = 0$ and that we obtain using (18). For Earth, (18) gives $s \sim 5 \text{ cm yr}^{-1}$, a good match to present-day observations (minimum < 1 cm yr$^{-1}$ in the Arctic Ocean, maximum 15 cm yr$^{-1}$ at the East Pacific Rise, mean $s \sim 4 \text{ cm yr}^{-1}$).

Reaching $Mo \sim 1$ requires temperatures near or beyond the limit of validity of our melting models. Coincidentally, $Mo = 1$ plots close to $Z_{crust}/Z_{lith} = 1$ on a mass-time graph. If massive Earth-like planets are in heat-pipe mode for $> 1 \text{ Gya}$, a substantial fraction of the census of nearby massive Earth-like planets would be in heat-pipe mode. And if Io is any guide (e.g., Spencer et al. 2007), dramatic spatial and temporal variations in thermal emission should be present.

However, the relatively slow cooling in our thermal evolution model is inappropriate for magma oceans (Sleep 2007), because our parameterization of viscosity does not include the fifteen-order-of-magnitude decrease associated with melting. Planets with $Mo > 1$ are likely to cool quickly to $Mo < 1$; we suspect that Io-type cooling is unsustainable for geologically significant periods without a nonradiogenic source of energy.

### 4.4. Phase transitions within crust

Basalt, which is less dense than mantle rock, undergoes a high-pressure exothermic phase transition to eclogite, which is more dense than mantle rock (Bucher & Frey 2002). Except during

\footnote{As planet mass increases, we hold plate area (rather than the number of plates) constant. If one instead holds the number of plates constant, $l_r$ falls and $s$ increases, but, because $\tau$ is unchanged, our buoyancy arguments in the following subsections are not affected.}
subduction, the pressures at the base of Earth’s basaltic crust are too low to make eclogite. However, the pressures at the base of the crust on massive Earth-like planets are far greater. If the crust of a massive earth-like planet includes an eclogite layer at its base, this may delaminate and founder, being refreshed by hotter mantle (Vlaar et al. 1994).

We tracked temperature and pressure at the base of the crust for planets in plate tectonics mode, and compared these with the phase boundaries for eclogite plotted in Fig. 9.9 of Bucher \& Frey (2002). We assumed uniform thermal conductivity within the thermal boundary layer. We found that eclogite is not stable for $M < 25 M_\oplus$. As planet mass increases, the ratio of crustal to lithospheric thickness increases, while mantle temperature also rises. Consequently, the temperature at the base of the crust rises, inhibiting the exothermic eclogite-forming reaction. Phase transitions within the crust should not inhibit ongoing plate tectonics on massive Earth-like planets.

4.5. A continental throttle?

Continental crust may sequester radiogenic elements, inhibiting plate tectonics and melting by imposing more rapid mantle cooling. The limit is a fully differentiated planet, in which successive melting events have distilled nearly all radiogenic elements into a thin shell near the surface. Independent of this effect, too great an area of nonsubductible, insulating continents (Lenardic et al. 2005) may itself be enough to choke off plate tectonics - a limit of 50% of total area has been suggested. Which limit is more restrictive to plate tectonics? On Earth, continents contain (26-77)% of the radiogenic complement of Bulk Silicate Earth (Korenaga 2008), and cover $\sim 40\%$ of the planet’s surface area. If the threshold value for a significant nonsubductible-cover effect on mantle dynamics is 50% (Lenardic et al. 2005), this limit will be passed before complete differentiation occurs. However, all radiogenic elements would be sequestered in continental crust before continental coverage reached 100%. We expect that the thickness of continental crust will scale as the inverse of gravity. This is because crustal thickness is limited by crustal flow (and brittle failure down gravitational-potential-energy gradients). If in addition the radiogenic-element content of continental crust does not vary, then we can relate the two limits by expressing the fraction of planet surface area covered by continents as

$$f_{\text{area}} = \frac{V_{\text{cont}}}{AZ_{\text{crust}}} \propto \frac{f_{\text{radio}} M^1}{M^{1/2} M^{-1/2}} \propto f_{\text{radio}} M. \quad (21)$$

This is linear in mass, implying that continental coverage will be the more severe limit for massive planets as well. It is much easier to enshroud the surface of a massive planet with nonsubductible material than to sequester a substantial fraction of that planet’s radiogenic elements into nonsubductible material.

Provided that crustal flow limits continental thickness, a representative calculation shows that
continents will quickly ooze out to coat the surface of an Earth-like planet \( > 3 \, M_\oplus \) (Figure 18); this is because continental production rate scales roughly as planet mass, but planet area increases only as the square root of mass. To produce this figure, we set continental growth to nil for the first 1 Gyr of each planet’s evolution (guided by the age of the oldest sizeable continental blocks on Earth). From 1 Gyr forward, we set continental growth to be proportional to crustal production rate, with a proportionality constant picked to obtain 40% coverage on Earth today.

These rather artificial choices reflect our poor understanding of continental growth and survival. Hydration of the oceanic crust at mid-ocean ridges, and subduction of this H\(_2\)O, is probably required for continental growth (Campbell & Taylor 1983). There is geological and isotopic evidence for episodic continental growth (Condie & Benn 2006). Durable continents may require photosynthetic life both in the oceans (Rosing 2006) and on land (Lee et al. 2008). None of these effects are straightforward to model. So, though we note that excessive continental surface area and radiogenic-element sequestration both have the potential to throttle our predicted increase in mantle melting on high-mass Earths, we consider this only a tentative prediction.

What would a planet look like as the fractional continental area approached 1? As the horizontal length scale of convection associated with the ocean basins dwindled and heat flow became concentrated into a smaller area, plate rates would accelerate and \( \tau \) would fall.

4.6. Will trenches jam?

Crust is less dense than mantle, which tends to retard subduction. However, there is another contribution to the density of the plate: that of the underlying lithospheric mantle, which is colder and denser than convecting mantle. As plate material moves away from the ridge, the lithospheric mantle thickens and the initially positive buoyancy of the plate becomes negative, aiding subduction. Provided that thermal conductivity and crustal density are constant,

\[
\Delta \rho \approx -\rho_{lith} + \frac{1}{Z_{lith}(\tau)} \left( \rho_{lith}(1 + \alpha \Delta T_3)(Z_{lith}(\tau) - Z_{crust}) + (\rho_{crust}Z_{crust}) \right),
\]

where

\[
\Delta T_3 = \frac{1}{2} \left( 1 - \frac{Z_{crust}}{Z_{lith}(\tau)} \right) \Delta T_2 \tag{23}
\]

and \( \Delta \rho \) is the density difference favoring subduction, \( \rho_{lith} \) is the reference density of mantle underlying the plate, \( Z_{lith} \) is lithospheric thickness, \( \Delta T_3 \) is the average cooling of mantle lithosphere, and \( \tau \) the characteristic age at subduction.

Hotter – that is, bigger or younger – planets must recycle plate faster, so plate has less time to cool. In addition, higher potential temperatures produce a thicker crust. Both factors tend to
produce positively buoyant plate, which is harder to subduct. This effect is more severe for massive planets because of their greater gravity.

Very young (and positively buoyant) oceanic plate can subduct on today’s Earth, provided that it is attached to a negatively buoyant slab. The associated ‘slab pull’ force depends, in part, on the basalt–eclogite transformation, which will occur at shallower depths on more massive Earths. However, it is not clear that subduction is possible if plate is positively buoyant everywhere, as likely for Early Earth (Davies 1992, Sleep 2000). The geological record is neither mute nor decisive. All but the last Gyr of Earth’s tectonic history is disputed, but evidence is accumulating that subduction first occurred at least 2.7 Gya (1.8 Ga after formation), and perhaps earlier than 3.2 Gya (1.3 Ga after formation) (Rollinson 2007). In some way Early Earth must have overcome buoyancy stress after this point. Taking buoyancy stress per unit length of trench to be the appropriate metric for buoyancy, we can relate Archean observations to high mass planets (Figure 20). For example, the buoyancy stress that had to be overcome on Earth after 1.8 Ga is the same as that on a 16 $M_\oplus$ planet after 4.5 Ga. Using buoyancy force per unit volume leads to basically the same conclusions (Figure 19). Here we have assumed that all plate reaches the subduction zone at a characteristic age, $\tau$, that the temperature distribution within the plate is described by half-space cooling (so $Z_{lith} = 2.32 \kappa^{0.5}r^{0.5}$), and that $k$ and $\rho_{crust}$ (2860 kg m$^{-3}$; Carlson & Herrick 1990) are constant. This gives a negative (subduction-favoring) buoyancy stress of 38 MPa at the characteristic age of subduction on the present Earth. The much more sophisticated model of Afonso et al. (2007) yields 21 MPa, so our estimate of forces retarding subduction is probably conservative.

The lower parts of a thick, water-rich crust may be in a temperature-pressure regime which permits amphibolite-grade metamorphism, leading to a significant increase in density. For this reason, we also plot results for a constant crustal density of 3000 kg m$^{-3}$, intermediate between the densities of amphibolite (3000 – 3300 kg m$^{-3}$; Cloos 1993) and unmetamorphosed crust. Because high surface temperatures favor amphibolite-forming metamorphic reactions, subduction should be easier on hot, water-rich (‘sauna’) worlds.

Our treatment is crude, not least because the crust-mantle density contrast declines with increasing melt fraction as the olivine content of the crust increases. And hydrothermal cooling may be ineffective beyond a fixed temperature (Vance et al. 2007), supplementing conductive cooling to a lesser extent on more massive planets than on Earth. Nevertheless, our assessment is that plate buoyancy raises a severe hurdle for plate tectonics on massive Earths, and may well be limiting.
5. Implications for degassing and atmospheric mass

5.1. Production rate

Rate of crust production is the product of crustal thickness (§3), spreading rate, and mid-ocean-ridge length. In the absence of continents, with constant plate geometry, and neglecting heat loss through magma pipes, spreading rate should rise as the square of lithospheric heat flux if heat lost through the top of the lithosphere is to balance that supplied at the top of the mantle (Sleep & Zahnle 2001).

In plate tectonics mode, for ‘Earth-like’ planets with oceans and some land, we calculate the rate of volcanism by multiplying spreading rate, mid-ocean ridge length, crustal thickness and crustal density, then dividing by planet mass (Figure 22). Our model gives $1.2 \times 10^{-11}$ yr$^{-1}$ by mass for Earth, observations give $1.1 \times 10^{-11}$ yr$^{-1}$. The discrepancy is mainly due to our model’s higher-than-observed spreading rate (§4.3). For comparison, Earth’s total observed present-day rate of volcanism is $1.7 \times 10^{-11}$ yr$^{-1}$ by mass, and $3.1 \times 10^{-11}$ yr$^{-1}$ by volume (Best & Christiansen 2001). Rate per unit mass increases monotonically with increasing mass, for all times. Per unit mass, rates of volcanism vary only by a factor of three of each other on planets < 3 Gyr old and > 1 $M_\oplus$, but a stronger mass dependence develops for older planets (Figure 22). Melting in plate-tectonic mode ceases when potential temperature falls below the zero-pressure solidus, but this only occurs in our modeling for small (0.25 $M_\oplus$) planets with ages comparable to the Galaxy.

In plate tectonics, the flux of mantle into the melting zone balances spreading rate (Figure 4). To calculate the rate of volcanism on a planet in stagnant lid mode (Figure 23), we must first find the flux of material into the upper boundary layer as a function of $Ra$. We obtain the radial convective velocity from the Nusselt number, noting that all heat flow in excess of the conductive heat flow must be advected:

\[ Nu - 1 = \frac{u(T_m - T_s)\rho c}{2} \]
\[ u = 2 \frac{(Nu - 1)}{(T_m - T_s)\rho c}, \]

where the factor of 2 takes account of the need for downwelling material to balance upwelling. Next we modify an expression in Breuer & Spohn (2006) to obtain

\[ R = \frac{A u \rho}{M_{\text{planet}} \rho_{\text{crust}}} \left( \frac{V_a}{V_m} \right) \left( \frac{\rho_{\text{crust}} Z_{\text{crust}}}{P_f - P_o} \right), \]

\[ \text{(25)} \]

\(^7\text{Including arc and ocean-island volcanoes, not modelled in this paper.}\)
where $R$ is the rate of melt generation per unit mass, $V_a$ and $V_m$ are the melt zone and total mantle volumes, and the last term on the right is the mean melt fraction.

This sets an upper limit to the rate of volcanism (Figure 23) because it assumes that all melt generated reaches the surface, and that all ascending mantle parcels reach $P_o^8$. With these assumptions, we find that the rate of volcanism on stagnant lid planets is initially more than double that on their plate tectonic counterparts (Figure 24), but this contrast soon reverses as the stagnant lid thickens. The predicted shutdown of non-plume melting on stagnant lid planets (Figure 24) is remarkably independent of mass and melting model, as explained in §3.3. For these reasons, and because the prediction tallies with Solar System observations, we consider it to be a robust result.

Crustal production is a proxy for degassing to within a factor of a few; therefore, scaling rates of degassing using our rate of volcanism plots (Figures 22–24) should be a good guide to degassing on extrasolar planets with a complement of mantle volatiles similar to that of Earth. But because volatiles are incompatible, they partition almost quantitatively into even small fractions of melt. Therefore, a more accurate statement is that degassing should be proportional to the flux of mantle processed through melting zones (Papuc & Davies 2008). Because of volatile incompatibility, order-unity differences in the volatile concentrations of planetary mantles should not make much of a difference to our melting curves. Wet parcels of ascending mantle soon ‘dry out’.

The implications for atmospheric mass of our results depend on the water and CO$_2$ contents of extrasolar planet mantles. These will depend on the mantle’s oxidation state, the range of semimajor axes from which the growing planet draws material, and the extent of volatile loss on the planetesimals which collide to form the planet. The water and CO$_2$ contents of terrestrial magmas are our only empirical guide. At mid-ocean ridges, these range from 0.1 % — 1.5 % for H$_2$O, and 50-400 ppm for CO$_2$ (Oppenheimer 2003). Taking the upper-limit values$^9$ assuming that all mantle fluxing through the melting zone is completely degassed, and ignoring overburden pressure, we obtain for $6 \times 10^{13}$ mol a$^{-1}$ for H$_2$O and $7 \times 10^{11}$ mol a$^{-1}$ for CO$_2$.

This is within the range of uncertainty of observational estimates. Because our model is not ‘tuned’ to observed rates of volcanism, but only to crustal thickness, this lends some credence to our results for other Earth-like planets.

5.2 Overburden pressure

If we relax the requirement that ocean pressure is small, then the efficiency of degassing from seafloor volcanoes will be regulated by the thickness of the volatile overburden. A volatile envelope

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$^8$ However, the first assumption probably introduces only a small error because of the large forces driving magma ascent on massive planets (gravity), and the short absolute distances they must ascend (thin lithospheres).

$^9$ Which is appropriate, given that additional volatiles are almost certainly present in an unmeasured fluid phase.
can have two effects:

(1) through a greenhouse effect, a volatile envelope can raise surface temperatures, and increase partial melting. This is shown for our thermal-equilibrium model in Figure 7. Here, we show the internal temperature needed to drive convection when the surface temperature is 647K, the critical point of water (q.v. 273 K for the baseline model). In thermal equilibrium, the reduced mantle-surface temperature difference demands more vigorous convection to drive the same heat flux across the upper boundary layer. The increase in mantle temperature is 50-55K (Figure 7), which has a significant feedback effect on melting (Figure 5). Such high surface temperatures are classically considered to be encountered only for brief intervals on the path to a runaway greenhouse (Ingersoll, 1969), but if this catastrophe is suppressed for massive Earths (Pierrehumbert, 2007), massive Earths will experience these temperatures for geologically significant intervals. The sign of this feedback is positive; higher mantle temperatures increase crustal thickness, and the associated degassing would enhance the greenhouse effect. We do not consider still higher temperatures, relevant for close-in rocky exoplanets (Gaidos et al. 2007), because plate tectonics is thought to require liquid water.

(2) through overburden pressure, a volatile envelope can suppress degassing (and melting for sufficiently thick volatile layers). It is not necessary to invoke an ‘ocean-planet’ (Kuchner 2003; Léger et al. 2004), or even a high-pressure ice layer, to suppress degassing (though either would suffice). Water degassing is readily suppressed by 0.1-0.2 GPa of overburden (Papale 1997), although joint-solubility effects allow some water to degas at higher pressures in association with other gases (Papale 1999). CO$_2$, an important regulator of climate on the known terrestrial planets, has a solubility of $0.005\% \text{ GPa}^{-1}$. Therefore, an overburden pressure of 1 GPa is enough to suppress CO$_2$ degassing from a magma containing 0.5 wt % CO$_2$.

A planet with a massive ocean cannot degas. However, it can regas, provided that hydrous/carbonated minerals pass the line of arcs at subduction zones. This suggests a steady state ocean mass over geodynamic time.

Deterministic scalings relating volatile mass, $M_{vol}$, to planet mass, $M$, should be treated with caution, because numerical simulations of late-stage accretion show a large range of volatile fractions. Reasonable limits are $M_{vol} \propto R^2 \sim M^{1/2}$ (if all planetary volatiles are delivered in small bodies with high crossing velocities) and $M_{vol} \propto M$ (if planetesimal composition is uniform and no water is lost). Best-fits to published N-body simulation output (Raymond et al. 2004, 2006) are intermediate between these values, but with water mass fractions scattered from 0.1% to 2% for planets $\sim 1 M_{\oplus}$. Assuming that volatiles are partitioned between surface and mantle reservoirs in the same proportions as on Earth, the lower limit gives an ocean of constant depth, and the upper limit an ocean whose depth scales as $M^{1/2}$. We plot the resulting mid-ocean-ridge pressures in figure 25.

Increased pressure will inhibit melting, though rather high pressures (e.g., a rather deep ocean) are needed to shut down melting at the mantle temperatures we consider. As an illustration, we
show the effect of an ocean whose mass scales as $M$ on crustal thickness for the K03 melting model (Figure 26). Volatile overburden leads to a reduction in crustal thickness that is always $\leq 45\%$, and $\leq 20\%$ for planets $\leq 10 M_{\oplus}$ and $\leq 10$ Gya old.

Note that even for an ocean of constant depth, land is unlikely. Gravity defeats hypsometry. For example, doubling Earth’s gravity would reduce the land area by a factor of eight. This is important for climate-stabilizing feedback loops involving greenhouse-gas drawdown, because only subaerial weathering is strongly temperature-dependent. Submarine weathering (carbonitization of seafloor basalt) is instead CO$_2$ concentration dependent.

The largest planets we consider may accrete and retain a significant H/He atmosphere [Ikoma et al. 2001], which could also frustrate melting. Apart from Uranus and Neptune, this has as yet only been confirmed for the 23 $M_{\oplus}$ ice giant GJ 436 b, most recently by Bean et al. (2008).

### 5.3. Melt-residue density inversion: decoupling melting from mantle degassing?

Melts are more compressible than mantle minerals, so at sufficiently high pressures melt will be denser than its residue. As a representative value, mid-ocean ridge basalt becomes denser than liquidus garnet at 12.5 – 19.5 GPa [Agee 1998]. If the sinking rate exceeds mantle velocities, these denser melts may accumulate at the base of the mantle. Meanwhile, the residue will continue to rise to shallow pressures, where it will eventually generate melt less dense than itself. That melt will segregate, ascend, and form a crust. However, because atmosphere-forming volatiles are highly incompatible, they will partition into the early-stage (sinking) melt. Because erupted melts will be volatile-poor, planets subject to this process will be unable to balance atmospheric losses to space. If the planet loses its initial (magma ocean phase) atmosphere due to high FUV/EUV flux from the young star [Guinan et al. 2003; Manning et al. 2006], it will be unable to rejuvenate that atmosphere.

Generating melt at sufficiently high pressures for the density inversion to come into play requires very high potential temperatures. In our modelling, these are not encountered for planets in plate tectonic mode (except during the first 1 Gyr, which is a transient associated with cooling from our high initial temperature; initial conditions are thus very important). But the density crossover is encountered for $> 5 M_{\oplus}$ and $< 3$ Gyr in stagnant lid mode. We speculate that in these planets, melting is decoupled from mantle degassing. Melting models calibrated to higher pressures and temperatures, and based on a high-pressure equation-of-state for silicate liquids [Ghiorso 2004], will soon become available [Ghiorso et al. 2007], allowing this hypothesis to be developed further.
6. Discussion and conclusions

6.1. Summary of results

Because of the exponential dependence of viscosity on temperature, modest variation in $T_m$ can accommodate the range of heat flows generated by planets of varying masses. If plate tectonics is assumed to be possible on massive Earth-like planets, our model confirms the expected, relatively weak dependencies: crustal thickness decreases slightly with increasing planet mass, and vertical tectonics is slightly favored over plate tectonics with increasing planet mass. For real planets, these effects are convolved with the effects of varying cosmochemistry and initial radiogenic element complement.

Our current understanding indicates that the first generation of Earth-like planets in the Galaxy would have formed 10-12 Gya. All our melting models predict that the viability of present-day melting on those planets depends strongly on mass and tectonic mode. In plate tectonic mode, larger planets are much more likely to be presently volcanically active and degassing. Our simple models indicate that all first-generation Earth-like planets in stagnant lid mode should have ceased melting. This prediction is robust with respect to variation of parameters such as $f_{\text{core}}$, radiogenic-element complement, and $T_i$. However, Mars’ continuing — although fitful and low-rate — volcanic activity (Borg & Drake 2005) indicates that volcanic activity can continue on planets in stagnant-lid mode for longer than simple models predict, perhaps as the result of plumes in a compositionally-layered mantle (Wenzel et al. 2004) or a thick, thermally insulating crust (Schumacher & Breuer 2007). On Venus, mantle fluxing by sinking delaminated lithosphere has been proposed as a mechanism that would allow volcanism to continue through to the present day (Elkins-Tanton et al. 2007), although there is no robust evidence of ongoing volcanism. Continued Solar System exploration is needed if we are to fully exploit nearby data points to understand the geodynamic window for life.

6.2. Overview of approximations and model limitations

6.2.1. Nature of mantle convection

In this paper we assume whole-mantle convection. Layered mantle convection can alter the volcanic history of a planet by introducing long-term sensitivity to initial thermal conditions. Butler & Peltier (2002) recover Earth’s thermal history using a parameterization of $Ra$ dependent mantle layering from (2D) numerical models of mantle convection. However, more complete (3D) models show that rare mantle avalanches overcome the apparently high degree of layering in high-$Ra$ models, which invalidates the proposed lower-mantle buffering of upper-mantle temperature.

We have not considered metalized silicates (Umemoto et al. 2006), nor the possibility that lower-mantle convection is extremely sluggish (even isoviscous; Fowler 1983) due to low homol-
ogous temperatures, or pressure dependent viscosity (Papuc & Davies 2008). Our parameterized treatment of mantle convection assumes the solid state and is inappropriate for very high temperatures, when the greater part of the lithosphere is underlain by a magma ocean. For almost all cases, however, the transition to a magma ocean takes place at temperatures beyond the range of validity of our melting models, so magma ocean development is not the limiting consideration in interpreting our results.

6.2.2. Thermal evolution and evidence from Early Earth

We have restricted our attention to planets with an Earth-like bulk composition and $f_{\text{core}}$, and with $H_i$ similar to chondritic. Although these may turn out to be rather uncommon (Stevenson 2004), reproducing the thermal history of Earth’s mantle should be a minimum criterion for a successful rocky planet model (Loyd et al. 2007). We do not want to give the reader an optimistic picture of progress. Embarrassingly, only one physically-based model (Korenaga 2006) can currently reproduce Earth’s thermal history as inferred from geology and petrology, and its assumptions might not apply to Earth’s past. Empirical models may approximate Earth’s thermal history (Labrosse & Jaupart 2007), but defy scaling to higher mass.

Given these problems, we have used a very simple thermal evolution model (Schubert et al. 2001) that, although it has known problems (Sleep 2007), at least has well-characterized behavior. The virtues of this approach include ease of interpretation, a common method for plate tectonics and stagnant lid mode, wide use, insensitivity to initial conditions (given a ‘hot start’), and negligible computational expense. A key assumption with this approach is that plate spreading rate adjusts to balance the heat flow at the top of the convecting mantle, which need not be even approximately true (Sleep 2000).

Earth-like behavior, our point of reference in this paper, may be atypical (Bostrom 2002). Cooling times for rocky planets are far greater than for stars, and their thermal state carries a correspondingly greater imprint of past conditions. This means that thermal history (Grove & Parman 2004) matters a lot, but we do not understand plate tectonics well enough to model thermal evolution in this mode accurately. The simple thermal models outlined here should be considered as a placeholder for future, more detailed models. Improved melting models will also be very welcome, but cannot overcome the limiting factor in our present-day understanding – how does plate tectonics couple to mantle thermal evolution (Sleep 2007)?

6.2.3. Other caveats

Volcanism is not the only possible source of degassing. Metamorphic decarbonation (Bickle 1996), and the episodic release of deep-seated volatiles as on the Moon (Gorenstein & Bjorkholm 1973), could permit (low) rates of degassing on non-volcanic worlds.
On stagnant lid planets, volcanism may be episodic (as on Mars; Wilson et al. 2001) and absent for long periods. This means that an observed lack of compounds with a short photochemical lifetime would not preclude a high level of volcanic activity averaged over a sufficiently long period ( \( \geq 10^{-8} \) yr).

By assuming that all melt reaches the surface, we have ignored the distinction between extrusion and intrusion. However, extrusion can alter the reflectance properties of the planet without sustaining an atmosphere and it may eventually be possible to discriminate between the two possible fates for melt.

### 6.3. Initial bulk-chemistry and initial radiogenic-power variations

The long-term thermal evolution of rocky planets depends on the abundance of the long-lived radiosotopes \(^{232}\text{Th},^{235}\text{U},\) and \(^{238}\text{U}\) at the time of planet formation. These are produced only by the rapid neutron capture process (\(r\)-process) acting on the iron-peak isotopes. This is thought to occur only during explosive nucleosynthesis in stars with \(10^{-20}M_\odot\) (Chen et al. 2006). In contrast, Si is produced during \(\alpha\)-chain process by the whole range of massive stars. Th, and especially U, are difficult to detect in stars but europium (Eu), another exclusively \(r\)-process element, can be readily measured. The average observed stellar abundance of Eu to silicon decreases by a factor of 0.63 as the abundance of heavy elements or metallicity (represented by iron Fe) increases by a factor of 100 to the solar value (Cescutti 2008). The \(r\)-process appears to be universal and all \(r\)-process elements scale closely with solar values (Frebel 2008). Therefore the average abundance of \(^{232}\text{Th},^{235}\text{U}\) and \(^{238}\text{U}\) isotopes can be predicted using the trend of Eu with abundance or metallicity, the age-metallicity relationship of the Galaxy, the star formation history of the Galaxy, and the half life of each isotope. We adopt a simple linear age-metallicity relationship with an increase of 1 dex (factor of 2.5) over the age of the Galaxy, with solar metallicity occurring 4.6 Gyr ago (e.g. Pont & Eyer 2004). Figure 27 plots the predicted abundance of the three isotopes using the observed trend of Eu and the Prantzos & Silk (1998) parameterization of the star formation history of the Galaxy. (The predictions are only weakly sensitive to the model of star formation). The age of the Galaxy is taken to be 13.6 Gyr. All abundances are normalized to the value at the formation of the Sun.

Planets forming early in the history of the Galaxy would have 50% more \(^{238}\text{U}\), but 6 times more \(^{235}\text{U}\), than Earth. The higher abundance is because the amount of radioisotopes in the interstellar medium only reflects massive star formation over a few half-lives, whereas \(^{28}\text{Si}\) and other stable isotopes accumulate over the history of the Galaxy. Therefore these systems are not U- and Th-rich, they are Si-poor. The high abundance of \(^{235}\text{U}\) could have an important role in the early thermal history of such planets.

\(^{10}\)Consider the Moon.
The effect of these trends on present-day planet temperatures, while still significant, is more modest. Figure 28 plots planet temperature against the age of the host star. Comparison with Figure 10a shows that inclusion of cosmochemical trends in $H_l$ lowers $T_m$ by up to 50 K for young planets, while raising $T_m$ by up to 40 K for old stars, compared to their present-day temperature had they formed with an Earthlike inventory of radiogenic elements.

We have assumed that the major-element composition of planetary mantles is similar everywhere and at all times. This might not be true: for example, it has been proposed that on early Earth the mid-ocean-ridge-basalt source was more depleted than at the present day (Davies 2007b). Earth’s continent mass fraction may be higher (Rosing 2006), or lower than is typical. More severe variations in major-element composition, with correspondingly major shifts in rheology and in the solidus, can be imagined (e.g., Gaidos 2000; Kuchner & Seager 2005). Even highly oxidized ‘coreless’ planets have been modelled (Elkins-Tanton & Seager in press). We will address the geodynamic consequences of such variations in future work, with the objective of using geodynamic observables to constrain internal structure and bulk composition.

6.4. Volcanism and consequences for chemical cycles and climate stability

Most of the partitioning of mass between a rocky planet’s chemical reservoirs occurs early in that planet’s history (Stevenson 2008), but persistent exchange between near-surface and deep-subsurface reservoirs is needed to maintain the disequilibrium chemistry that permits life to endure (Chyba 2000; Parkinson et al. 2007; Rosing 2006). Planets may close their heat budgets without resorting to volcanism (Michael et al. 2003; Reese et al. 2007). So orbital stability and a Goldilocks quotient of metal, rock, and volatiles are insufficient conditions for the long-term habitability that the fossil record teaches us is a prerequisite for advanced life (Butterfield 2007). Geodynamics also matters (e.g. von Bloh et al. 2007).

On the Earth, the relative abundances of the rare gases and their isotopes indicate the loss of an earlier (although not necessarily primordial) atmosphere and its replacement by an atmosphere sustained by slow degassing of the mantle over time. The persistence of this atmosphere and its modification by chemical reaction with surface rocks and fluids are considered critical to the maintenance of “habitable” conditions on the Earth over a significant fraction of the age of the Sun. Since the appearance of free molecular oxygen in the atmosphere 2.4 Ga it is thought that carbon dioxide has been the leading greenhouse gas. A negative feedback involving weathering of silicate rocks and sequestration of carbon dioxide in carbonate rocks has been invoked to explain the (almost?) continuous stability of surface liquid water, which would otherwise be hard to maintain in the face of a brightening Sun (Walker et al. 1981). (The alternative to a negative feedback is that Earth’s apparent climate stability is the result of chance.) Partial melting is required for the silicate-weathering feedback both by providing the CO$_2$ that is degassing and by producing fresh volcanic crust for weathering.
On the Earth, plate tectonics, via degassing at active plate margins, subduction of carbon-bearing rocks, and exposure of crustal rocks in mountain belts, is intimately associated with the long-term geologic carbon cycle (Hayes & Waldbauer 2006). However, the absence of plate tectonics on other, otherwise Earth-like planets need not preclude the operation of a carbonate-silicate feedback nor habitable conditions, i.e. stable liquid water at the surface. At the minimum, such a condition requires that the rate of silicate weathering and production of alkalinity at surface temperatures above freezing balances outgassing of CO$_2$ from all sources, and that the temperature dependence of weathering is positive. On a planet with plate tectonics, the first requirement is expressed as:

\[ f_{\text{mantle}} + f_{\text{arc}} + f_{\text{meta}} = f_{\text{weathering}} \] (26)

where the terms are, respectively:– the flux of carbon dioxide to/from the atmosphere/ocean reservoir due to mantle melting and degassing; arc volcanism, partially recycling subducted carbon; metasomatism of carbonate rocks; and carbonate deposition due to the weathering of silicate rocks and production of alkalinity. In the absence of plate tectonics and any mountain-building, the only non-zero terms are the right hand side, and the first term on the LHS. A stagnant lid planet is less capable of recycling carbon dioxide back into the atmosphere from its crustal reservoir. We also predict that SL planets the same mass and age of the Earth will have less mantle melting - or none at all. (We have also shown that younger SL planets will have higher rates of mantle melting.)

For example, estimated CO$_2$ fluxes on the present Earth are 2.2, 2.5, and < 3 for mid-ocean ridge, arc, and plume volcanism, respectively, in units of 10$^{12}$ mol a$^{-1}$ (Marty & Tolstikhin 1998). On an SL planet of identical age and mass, mantle degassing should either have ceased entirely (using the MB88 model) or be 1/3 that of the present Earth (using the K03 model; Figure 24). (However, earlier in their evolution, stagnant lid planets have rates of volcanism/degassing greater than that of contemporary plate tectonic planets; Figure 24).

We schematically show the balance between degassing of CO$_2$ and weathering of silicates in Figure [29] Weathering rates in the laboratory have an exponential temperature dependence. The mean surface temperature is set by the balance of degassing and weathering; if temperatures increase (decrease), weathering exceeds (falls below) degassing and the CO$_2$ content of the atmosphere and oceans decreases (increases). Ultimately, silicate weathering requires the production and exposure of igneous rocks. Rocks formed from the weathering and alteration of igneous rocks are less fertile sources of alkalinity (e.g., quartz sandstone) during subsequent weathering. At sufficiently high rates of degassing and sufficiently high surface temperatures on a wet planet, weathering may be limited by the availability of fresh igneous rock, rather than chemical rates of reaction. In this case, the rate of weathering becomes temperature-independent, the silicate weathering “thermostat” fails, and climate may become unstable and enter a runaway wet greenhouse state. However, seafloor weathering may provide a base level of weathering that is relatively independent of surface temperature, but proportional to the rate of mid-ocean ridge volcanism (Brady & Gislason 1997).

Therefore, all else being equal, massive stagnant-lid planets of comparable age to Earth will experience lower rates of silicate weathering (Figure [29]. Because of a lower rate of igneous rock
production, the weathering curve saturates at a lower temperature (Figure 29). These considerations suggest that massive stagnant-lid planets that are Earth’s contemporaries will have a colder climate than those with plate tectonics, but that the silicate weathering feedback can still operate. Once volcanic degassing ceases, however, no process can balance long-term CO$_2$ drawdown, and plants would fast become extinct (Caldeira & Kasting 1992).

Stagnant-lid planet climate systems operating at a lower temperature will necessarily have less CO$_2$ in their atmosphere. As stellar luminosity rises the point at which there is no CO$_2$, and at which surface temperatures begin to rise, is reached earlier. However, the evolutionary trajectories of the atmospheres of SL and PT planets will converge once both planets have no atmospheric CO$_2$.

Many of the ideas discussed here are drawn from the conversation of Dave Stevenson, whose support for the early stages of this project was invaluable. We thank Nick Butterfield, Rhea Workman and Brook Peterson for useful suggestions. We acknowledge support from the NASA Astrobiology Institute. E.K. is supported by the Berkeley Fellowship and the Harkness Scholarship.

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Table 1. Radioisotope data: half lives, specific power \( W \), and concentrations \([X](i)\) (ppb) after 4.5 Gyr.

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<tr>
<th></th>
<th>(^{40}\text{K})</th>
<th>(^{232}\text{Th})</th>
<th>(^{235}\text{U})</th>
<th>(^{238}\text{U})</th>
<th>Ref.</th>
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<tr>
<td>( t^{1/2} ) (Gyr)</td>
<td>1.26</td>
<td>14.0</td>
<td>0.704</td>
<td>4.47</td>
<td></td>
</tr>
<tr>
<td>Specific power (x ( 10^{-5} ) W/kg)</td>
<td>2.92</td>
<td>2.64</td>
<td>56.9</td>
<td>9.46</td>
<td></td>
</tr>
<tr>
<td>Concentrations:</td>
<td>36.9</td>
<td>124</td>
<td>0.22</td>
<td>30.8</td>
<td></td>
</tr>
<tr>
<td>‘Mantle’</td>
<td>30.7</td>
<td>84.1</td>
<td>0.15</td>
<td>21.0</td>
<td></td>
</tr>
<tr>
<td>‘Undepleted Earth’</td>
<td>71.4</td>
<td>29.4</td>
<td>0.058</td>
<td>8.1</td>
<td></td>
</tr>
<tr>
<td>CI chondrites</td>
<td>147.8</td>
<td>2.8</td>
<td>5.8</td>
<td>13.0</td>
<td></td>
</tr>
<tr>
<td>EH chondrites</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
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</table>

Note. — \( H_i = [X](i)W_i \)

Table 2. Parameters used in interior and thermal models.

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<th>Parameter</th>
<th>Symbol</th>
<th>Value</th>
<th>Units</th>
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<tbody>
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<td>Thermal expansivity, mantle</td>
<td>( \alpha )</td>
<td>( 3 \times 10^{-5} )</td>
<td>K(^{-1} )</td>
</tr>
<tr>
<td>Thermal conductivity</td>
<td>( k )</td>
<td>4.18</td>
<td>W m(^{-1} ) K(^{-1} )</td>
</tr>
<tr>
<td>Thermal diffusivity</td>
<td>( \kappa )</td>
<td>0.3</td>
<td>m(^2) s(^{-1} )</td>
</tr>
<tr>
<td>Critical Raleigh number</td>
<td>( Ra_{cr} )</td>
<td>1100</td>
<td></td>
</tr>
<tr>
<td>Gas constant</td>
<td>( R )</td>
<td>8.31</td>
<td>J K(^{-1} ) mol(^{-1} )</td>
</tr>
<tr>
<td>Specific heat capacity, mantle</td>
<td>( c )</td>
<td>914.3636</td>
<td>J K(^{-1} ) kg(^{-1} )</td>
</tr>
<tr>
<td>Density, mantle</td>
<td>( \rho_{mantle} )</td>
<td>3400</td>
<td>Kg m(^{-3} )</td>
</tr>
<tr>
<td>Density, crust</td>
<td>( \rho_{crust} )</td>
<td>2860</td>
<td>Kg m(^{-3} )</td>
</tr>
<tr>
<td>Reference viscosity</td>
<td>( \nu_0 )</td>
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<td>m(^2) s(^{-1} )</td>
</tr>
<tr>
<td>Gravitational constant</td>
<td>( G )</td>
<td>( 6.673 \times 10^{-11} )</td>
<td>m(^3) kg(^{-1} ) s(^{-2} )</td>
</tr>
<tr>
<td>Core mass fraction</td>
<td>( f_{core} )</td>
<td>0.325</td>
<td></td>
</tr>
<tr>
<td>Earth radius</td>
<td>( R_{\oplus} )</td>
<td>6.372 \times 10(^6)</td>
<td>m</td>
</tr>
<tr>
<td>Earth mass</td>
<td>( M_{\oplus} )</td>
<td>5.9742 \times 10(^{24})</td>
<td>kg</td>
</tr>
<tr>
<td>Mantle temperature, initial</td>
<td>( T_m )</td>
<td>3273</td>
<td>K</td>
</tr>
<tr>
<td>Temperature change causing e-folding in viscosity</td>
<td>( T_\nu )</td>
<td>43 or 100</td>
<td>K</td>
</tr>
</tbody>
</table>
Plate tectonics

- 35 -

pressure
temperature
$T_p$
$T_lava$
solidus
liquidus

Plate tectonics

- 35 -
Fig. 1.— Sketches of pressure-temperature paths for passively upwelling mantle. a) Plate tectonics. Thick solid line is the adiabatic decompression path for solid mantle. Actual path taken by upwelling mantle, traced by arrows, differs above the solidus because of latent heat of fusion. b) Effect of a stagnant lid, whose base corresponds to the dashed line. Ascending mantle tracks the conductive geotherm within the lid (arrowed path). c) Melt fraction versus pressure for plate tectonics (thick solid line) and stagnant lid mode (thin solid line). Melt generated at $P < P_{\text{lith}}$ in the stagnant lid case is a small fraction of total melt, and we ignore it in this paper.
Fig. 2.— Planet radius as a function of mass. The dotted line corresponds to the constant-density control case. The dashed line uses the scaling of Valencia et al. (2006). The solid line uses the scaling of Seager et al. (2007), which neglects thermal expansion and so predicts a radius for Earth 3% smaller than the observed value.
Fig. 3.— Effect of changing activation energy for temperature-dependent viscosity on thermal evolution. Turcotte & Schubert (2002) radiogenic-element complement. Mantle temperatures are adjusted to produce 7 km thick crust under MB88 at 4.5 Gyr. Dots correspond to results with $T_v = 43$K. Stars correspond to results with $T_v = 100$K. Note that the cooling rate at 4.5 Gyr is 71 K/Gyr for $T_v = 100$K, more than double that for $T_v = 43$K (33 K/Gyr). Cessation of melting (dashed line) occurs around 9 Gyr for $T_v = 100$K, but takes longer than the age of the Universe for $T_v = 43$K.
Fig. 4.— a) Sketch to illustrate corner flow beneath mid-ocean ridges. The extent of partial melting suffered by parcels of mantle material near the edge of the melting triangle is less than that suffered by parcels directly beneath the mid-ocean ridge. Melt is focussed from a broad melting zone to a narrow magma lens beneath the ridge (Modified after Langmuir, 1992).
Fig. 5.— Crustal thickness as a function of potential temperature for $M_\oplus = 1$. Thick line corresponds to L92 model, thin solid line to MB88 model, thick dotted line to K03 model, and dashed line to pMELTS model. Thin dotted line is a cubic extrapolation of MB88 beyond its range of validity. pMELTS results are only shown where first melt occurs at $< 3$ GPa. MB88 results are only shown where melt fraction is zero at 8 GPa. Horizontal dash-dot line is observed crustal thickness on today’s Earth.
Fig. 6.— Effect of different radiogenic-element complements on thermal evolution. Mantle temperatures are not adjusted; all runs are with $T_\nu = 43$ K. Thick lines correspond to various chondritic scenarios: thick dashed line is EH chondrite; thick dotted line: CI chondrite; thick solid line: Ringwood, 1991. The thin solid line is for Turcotte & Schubert (2002), and the thin dotted line is for no radiogenic elements in the mantle. This could correspond, for example, to early and complete differentiation of the mantle to produce a thick crust, which is then swiftly removed by impacts. The horizontal dash-dot line at 1510 K corresponds to $Nu = 1$ (convection is no more efficient than conduction).
Fig. 7.— Effect of mass on mantle temperature for a planet in thermal equilibrium (Ur = 1) with a specific radiogenic power appropriate for today’s Earth. Thick lines correspond to a surface temperature of 273 K, thin lines correspond to a surface temperature of 647K. The solid line uses the scaling of Seager et al. (2007), and the dashed lines use the scaling of Valencia et al. (2006) note that the latter is only valid for $M < 10$ Earths. The dotted lines use constant-density scaling.
Fig. 8.— Effect of increasing planet mass on thermal evolution. Mantle temperatures are adjusted to produce 7 km thick crust under MB88 at 4.5 Gyr for M = 1. Dashed line is for M = 1; solid lines are for M = 5, 10, 15, 20 and 25 Earths. $T_v = 43$ K.
Fig. 9.— Thermal evolution of mantle for initial conditions at 250 K intervals from 273 K to 5273 K. Above 2500K, thermal evolution tracks rapidly converge; thermal evolution is insensitive to initial conditions if the starting mantle temperature is greater than that required for thermal equilibrium - a ‘hot start’. Hot starts are overwhelmingly likely for differentiated massive planets.
Fig. 10.— (Uppermost panel) Evolution of mantle temperature (K) with time under plate tectonics. Initial temperature for all models is 3273 K. (Lower panel, top to bottom) Corresponding crustal thicknesses for MB88, K03 and pMELTS melting modes. Contour interval is 1000m from 0 to 15000m, and 5000m for larger values. Light grey regions are where melting models are extrapolated beyond their stated range of validity; contours are terminated where melting models output unphysical results (e.g., decreasing melt with increasing temperature).
Fig. 11.— Effect of mode of mantle convection on thermal evolution. Stars correspond to stagnant lid mode. Dots correspond to plate tectonics mode. The thin solid line shows thermal evolution when an instantaneous switch to stagnant lid mode is imposed, after 5 Gyr, on a planet undergoing plate tectonics. The dashed line shows thermal evolution for the converse switch. For equivalent radiogenic complements, a planet in plate tectonics mode will have a lower potential temperature than a planet in stagnant lid mode. The difference is comparable to the range in potential temperatures due to mass.
Fig. 12.— Crustal thickness dependency on stagnant lid thickness, MB88 melting model. Dotted line is melting to surface. Thick solid line is melting to base of crust, no stagnant lid. Thin solid lines are melting to base of lithosphere or base of crust for a stagnant lid with a thickness of 20 km, 40 km, 60 km, and so on.
Fig. 13.— Evolution of crustal thickness with time in stagnant lid mode, for MB88 (top) and K03 (bottom) melting models. Wiggles in the CT = 0 contour are interpolation artifacts.
Fig. 14.— Limits to geodynamic lifetime. The region in the lower left corresponds to planets whose geodynamic lifetime is limited by cessation of volcanic activity. The region in the upper right corresponds to planets whose geodynamic lifetime is limited by the main-sequence lifetime of the host star. The lines dividing the two regions are for stagnant-lid convection (MB88 melting model, thick solid line) and for plate tectonics (MB88 model, thin solid line; pMELTs model, thin dotted line; L92 model, thick dotted line). Climate instability, not shown here, sets Earth’s remaining biosphere lifetime to \( \sim 1 \) Gyr (Caldeira & Kasting 1992).
Fig. 15.— Sketch of possible volcanism- and melting-related limits to plate tectonics: (1) crust thicker than lithosphere leading to intracrustal diapirism; (2) melting at depth leading to widespread intraplate volcanism; (3) magma-pipe transport energetically trumps conduction leading to Io-type tectonics; (4) phase transitions within crust leading to crustal delamination; (5) growing continents spread to cover most of the planet; (6) trench jam due to excessive plate buoyancy.
Fig. 16.— Possible limits to plate tectonics. Thick solid lines correspond to crust–to–lithosphere thickness ratios of 0.5 and 1.0 for MB88 melting model; thin solid lines are the same, but for K03 melting model. Points to the left of these solid lines may be subject to vertical (Io-type) tectonics. Dotted line is the limit of validity of the MB88 melting model. Results not shown for pMELTS melting model because solid lines fall in the temperature region for which pMELTS is not valid.
Fig. 17.— Plate spreading rate (m/yr) as a function of time. Thick lines correspond to solution including heat transport through magma pipes. Thin lines correspond to conduction-only solution. MB88 melting model.
Fig. 18.— Fractional area covered by continents, versus time. MB88 melting model. The dark shaded region is that for which the melting model is invalid. The light shaded region is that for which buoyancy stresses probably prevent plate tectonics. The vertical dashed line is the time after which continental growth is permitted.
Fig. 19.— Buoyancy forces as a function of thermal evolution and planet mass. Positive values denote plate denser than underlying mantle, favoring subduction; negative values denote plate more dense than underlying mantle, retarding subduction. Solid lines connect buoyancy values for planets of different masses 2.5 Gyr, 5 Gyr 7.5 Gyr and 10 Gyr after planet formation, for constant crustal density of 2860 kg m$^{-3}$. Dash-dot lines are for a crustal density of 3000 kg m$^{-3}$, as might be the case for partial amphibolitization. Dotted lines are possible lower limits to plate tectonics based on Earth’s (disputed) geological record; arguably, subduction must be possible on planets whose buoyancy forces plot above these lines. The Earth symbol is the model calculation for present day conditions on Earth.
Fig. 20.— As previous figure, but plotting buoyancy stresses (buoyancy force multiplied by lithospheric thickness).
Fig. 21.— Intensity of buoyancy stresses retarding subduction as a function of planet mass and age, for MB88 (top), K03 (middle) and pMELTS (bottom) melting models. Subduction is progressively less likely for points plotting to the left of the lines. Thin solid lines correspond to buoyancy stresses of 0, −50 and −100 MPa (positive values favor subduction). Thick solid lines correspond to the values of buoyancy stress on Earth 2.7 Gya (after 1.8 Gyr) and 3.2 Gyr (after 1.3 Gyr).
Fig. 22.— Rate of volcanism per unit mass on massive Earth-like planets experiencing plate tectonics, normalized to calculated rate on the Earth \( (3.7516 \times 10^{19} \text{ s}^{-1}, \text{equivalent to } 24 \text{ km}^3 \text{ yr}^{-1}) \), for MB88 (top), K03 (middle) and pMELTS (bottom) melting models. Light grey shaded regions correspond to negative buoyancy stresses with magnitudes in excess of 50 MPa, which would markedly inhibit subduction. Dark grey shaded regions correspond to mantle temperatures too high for a reliable crustal thickness calculation.
Fig. 23.— Rate of volcanism on massive Earth-like planets undergoing stagnant lid convection, normalized to calculated rate on a plate-tectonic Earth, for MB88 (top) and K03 (bottom) melting models. Dark grey shaded regions correspond to mantle temperatures too high for a reliable crustal thickness calculation. Contours are at 0, 0.5, 1, 2, 5, and 10 times Earth’s rate and at intervals of 10 thereafter.
Fig. 24.— Ratio of rate of volcanism per unit mass on stagnant lid planets to that on plate tectonic planets. MB88 melting model. Difference in rate is modest early in planet’s evolution, but volcanism shuts down on stagnant lid planets early.
Fig. 25.— Pressure exerted by oceans, assuming flat topography. The three solid lines correspond to different scalings of ocean mass with planet mass. The largest pressures we obtain are just sufficient to stabilize ice VI at 273 K (the dotted line corresponds to this phase transition). More aggressive scalings of volatile complement with planet mass will robustly generate a high-pressure ice deck between silicate and liquid-water shells for planets in the habitable zone.
Fig. 26.— To show effect on melting of a volatile overburden whose mass scales with planet mass, $M$. Thick lines correspond to results with a volatile overburden; thin lines correspond to results without a volatile overburden. Wiggles in the 3000m contour are interpolation artifacts. We treat the effect of the volatile overburden as being equivalent to that of a stagnant lid with the same basal pressure. Strictly, this isn’t so: it introduces an error of order $(\text{adiabat} \times \text{overburden pressure})$, which is acceptable for our purposes because the overburden pressure is never particularly large (max. 12 K error).
Fig. 27.— Abundance, relative to silicon and normalized to conditions at the time of the protosolar nebula, of the principal long-lived radionuclides in rocky planet mantles.
Fig. 28.— Observable temperature, tracking the effect of galactic cosmochemical evolution on initial radioisotope complement. Compare with top panel of Figure 10. Note that the abscissa is not time, but the age of planet at the present day.
Fig. 29.— Relationship between rates of weathering and temperature for planets with high (blue) and low (red) rates of volcanism. Equilibrium temperature is set by the balance between weathering drawdown of CO$_2$ and degassing. On the red planet — assumed to be in stagnant lid mode — we have much less degassing (dashed lines) because of less melt production, no arc volcanism, and no orogenic metasomatism of carbonates due to orogeny. But we also have less weathering (solid lines) because of less melting. Weathering rates saturate at high temperature because they become limited by the production of igneous rock. That saturation is lower for the red planet because of its lower rate of crust production. The rates do not vanish at 0°C because of seafloor weathering. The climate feedback need not disappear because the system is still in the positive slope regime, but the red planet will be colder.