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Authors

Nimmo, Francis
Mackwell, Stephen

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Viscous relaxation as a probe of heat flux and crustal plateau composition on Venus

Francis Nimmo[a,1] and Stephen Mackwell[b]

It has recently been suggested that deformed crustal plateaus on Venus may be composed of felsic (silica-rich) rocks, possibly supporting the idea of an ancient ocean there. However, these plateaus have a tendency to collapse owing to flow of the viscous lower crust. Felsic minerals, especially water-bearing ones, are much weaker and thus lead to more rapid collapse, than more mafic minerals. We model plateau topographic evolution using a non-Newtonian viscous relaxation code. Despite uncertainties in the likely crustal thickness and surface heat flux, we find that quartz-dominated rheologies relax too rapidly to be plausible plateau-forming material. For plateaus dominated by a dry anorthite rheology, survival is possible only if the background crustal thickness is less than 29 km, unless the heat flux on Venus is less than the radiogenic lower bound of 34 mW m^{-2}. Future spacecraft determinations of plateau crustal thickness and mineralogy will place firmer constraints on Venus's heat flux.

relaxation | rheology | geodynamics | tessera

Crustal plateaus are elevated, generally deformed regions on Venus, perhaps analogous to continents on Earth, e.g., ref. 1. They are dominated by radar-bright, heavily deformed terrain termed tessera. Recently, it has been proposed on the basis of infrared emissivity measurements (2–4) that tesserae are more felsic than the basaltic plains of Venus. If so, this is an important conclusion: granites—common felsic rocks—are rare on bodies other than the Earth (5) and are thought to form (6) and ascend (7) most readily in the presence of water. The existence of a possible ancient ocean on Venus is currently under debate (8, 9), so confirming the presence of felsic rocks there, and granites in particular, would be of considerable interest.

The aim of this paper is to point out an additional constraint on the problem that has not yet been exploited. Crustal plateaus have a tendency to undergo viscous relaxation over time (10, 11). The rate of relaxation depends on the rheology of the crustal material, the crustal thickness, and the heat flux. As will be shown below, felsic materials relax more rapidly than basaltic materials, and wet mineralogies relax more rapidly than dry mineralogies. Thus, in principle, the survival of elevated plateaus can be used to rule out certain mineralogies and water contents. At present, our knowledge of important Venusian parameters, most notably the heat flux, is too imprecise to allow definitive conclusions. However, if forthcoming spacecraft missions confirm the existence of felsic plateaus, a strong constraint on the allowable heat flux can be derived.

Background

Crustal plateaus have roughly circular planforms, with diameters ranging from 1,000 to 3,000 km, and are steep-sided with elevations as much as 2 to 4 km above the surrounding plains, e.g., ref. 11. They are usually characterized by complex tectonic fabrics and appear to be toward the base of the Venusian stratigraphic column (12). The most recent episode of deformation on crustal plateaus appears to be extensional, suggesting that they have experienced at least limited topographic relaxation (13). However, significant relaxation leads to predicted features that are not consistent with the observations (11).

Based on Magellan gravity and topography, the majority of crustal plateaus are supported isostatically, e.g., ref. 14; the high elevations make Pratt isostasy (i.e., support by density contrasts) implausible (15), so Airy isostasy (crustal thickness variations) is inferred. A crustal plateau 2 km high implies a crustal root about 11 km thick, depending on the density assumed. However, as discussed below, the mean crustal thickness of Venus is not very well constrained.

Significance

Crustal plateaus are elevated, generally deformed regions on Venus that may be made of silica-rich rocks. On the Earth, such rocks typically form in the presence of water, so silica-rich plateaus would support arguments for an ancient ocean on Venus. But these plateaus collapse because the rocks flow, and silica-rich rocks flow particularly fast. We use a model of plateau collapse to demonstrate that some silica-rich rocks can be ruled out entirely, and others are only plausible if either the crust of Venus is thin or its heat flux is small.

Author affiliations: [a]Department of Earth and Planetary Sciences, University of California, Santa Cruz, CA 95064; and [b]Department of Earth, Environmental and Planetary Sciences, Rice University, Houston, TX 77005

[1]To whom correspondence may be addressed. Email: fnimmo@ucsc.edu.

An early study of plateau relaxation on Venus found great difficulty in maintaining topographic contrasts over geological time (10). This turned out to be because the rheology adopted (16) was that of a diabase containing sufficient water to make it quite weak. A later relaxation study using dry diabase (which is much stronger) (17) found that plateaus could maintain their topography over Gyr timescales (11). This example illustrates the crucial role of water in determining crustal rheology.

Petrology and Rheology. The most felsic rocks (rhyolites and granites) consist primarily of feldspars and quartz. Terrestrial granites contain sodium-rich feldspars (e.g., albite) and typically a few percent water; the lunar crust is dominated by calcium-rich feldspars (anorthite) and is dry. The Martian crust also appears to contain abundant anorthite (18). Typically, rock rheology is controlled by the weakest abundant mineral. Thus, a granite consisting of feldspars and quartz will have a rheology resembling that of quartz (the weaker mineral).

While the plains of Venus are basaltic (19), a few steep-sided domes on Venus imply such high viscosities that they must be silica-rich (20). In the absence of water, producing such felsic rocks typically requires extreme igneous fractionation (5). This process could give rise to anorthosites on Venus (21), but the melt volumes involved may be too large to be plausible (4). On the Earth, granite formation is aided by the transport of water into the mantle at subduction zones (22). If Venus possessed an ancient ocean, then putative areas of vertical exchange between the near-surface and the mantle (23, 24) could give rise to felsic melts and hydrated minerals. Feldspars can retain significant amounts of bound water until at least 600 to 800 °C (25). Quartz grains heated to 550 °C are water-poor, resulting in a stronger terrestrial lower crust (26). Since the surface temperature of Venus is 450 °C, hydrated minerals at the surface of Venus are likely to stay hydrated. At depth, however, dehydration reactions (and potentially melting) will occur.

Below, we use the following flow laws for materials of interest: dry and wet anorthite from ref. 27; dry Columbia diabase from ref. 17; dry quartz from ref. 28; wet quartz* from ref. 29; dry albite from ref. 30. In all cases, dislocation creep ($n = 3$ for anorthite and quartz) is assumed, which means that creep behavior is independent of grain size.

We note here that a surface felsic composition, if present, does not necessarily require the lower crust to be felsic. In the bulk of this paper, we assume that it does, but we briefly return to this issue in the *Future Progress* section.

Relevant Parameter Values. As discussed below, several parameters are particularly relevant to interpreting our results: the age of crustal plateaus, the local crustal thickness, and the surface heat flux. These parameters are listed in order of increasing uncertainty.

The surface of Venus hosts about a thousand impact craters which are approximately randomly distributed. The inferred mean surface age is in the range 0.3 to 1 Gyr (31). While early investigations argued that this mean age in fact represented a catastrophic resurfacing event (32), more recent studies have argued that gradual volcanic resurfacing is more likely (33). Tesserae appear to sit at the base of the stratigraphic column (12), at least locally, and appear to be more heavily cratered than the surrounding plains (34). It is thus likely that they have persisted for at least a billion years.

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*Assuming a water fugacity of 300 MPa.

Crustal thickness estimates of Venus are based primarily on analysis of gravity and topography. This approach is limited by the quality of the data and inherent tradeoffs, e.g., between crustal and elastic thickness (14). Magellan-era estimates of global crustal thickness typically fall in the 20 to 40 km range (14, 35, 36) and 40 km for large highland regions specifically (37). More recent estimates are slightly lower: 8 to 25 km globally (38), 10 to 40 km (39), and 15 to 34 km (15) for crustal plateaus. Jiménez-Díaz et al. (40) assume a global mean value of 25 km. In general, these estimates depend on an assumed crustal and mantle density; below we will assume a mean thickness of between 20 and 30 km.

The heat flux of Venus is a vexed issue. The Earth's radiogenic heat production rate is estimated at 21.5 ± 3 TW (41), which is about half its heat loss rate of 47 TW (91 mW m^{-2}) (42). Plate tectonics is apparently allowing the Earth's mantle to cool rapidly. For a stagnant lid body, the relationship between heat production and surface heat loss is less clear, not least because of the potential importance of advection of heat by melt transport, e.g., ref. 43. Surface measurements of K/U ratios (19) suggest that Venus's radiogenic heat budget should not be very different from that of Earth. Accounting for the difference in radii, Venus radiogenic heat production would generate a mean surface heat flux of about 34 to 45 mW m^{-2}.

The most common way to estimate heat flux on Venus is to determine the thickness (termed the effective elastic thickness, T_e) of a perfectly elastic plate which reproduces an observed characteristic of the lithosphere, such as the flexural response to a load. This elastic thickness can then be converted to a heat flux using a rheological model. Unfortunately, published T_e estimates are highly variable; in some cases, this may be due to the low resolution of the spherical harmonic gravity (44), but there are also likely real spatial variations in heat flux at the time of loading. For example, topographic flexural profiles result in T_e values of 12 to 34 km (45), 29 ± 6 (46) and about 30 km (47). Gravity-topography studies result in global median T_e values of 20 km (39), 22 km (48), and about 15 km at plateaus (15). Plateau T_e values are sufficiently small that isostatic compensation is a good assumption (15).

Converting T_e to heat flux is not straightforward because it typically requires assumptions regarding the strain rate, topographic curvature, and rheology (49). Roughly speaking, the base of the elastic layer occurs at the temperature where viscous creep starts to dominate; thus, a higher T_e implies that this temperature occurs at greater depth, in turn implying a lower heat flux. The heat flux recorded is usually the value when deformation took place, which is not necessarily the present-day value.

Borreli et al. (47) infer a heat flux of 60 mW m^{-2} for $T_e = 30$ km; Maia and Wieczorek (15) find a range of 18 to 203 mW m^{-2} for six crustal plateaus. Coronae flexure implies heat fluxes in excess of 95 mW m^{-2} (50), but these regions may not be representative. Other heat flux estimation techniques have been applied locally. Bjonnes et al. (51) used impact models to place an upper bound of 28 mW m^{-2} at Mead crater, while Karimi and Dombard (52) used relaxation models of the same feature to infer a range of 55 to 90 mW m^{-2}. Modeling of small-scale tessera deformation requires heat fluxes in excess of 80 mW m^{-2} (53); this is consistent with the low crustal plateau T_e values.

In short, there is little agreement on Venus's heat flux, and it almost certainly varies spatially. Crustal plateau relaxation provides one possible way to constrain this heat flux, as will be argued below.

Method

To explore plateau relaxation on Venus, we use the model of ref. 54. This model treats the crust as a viscous, non-Newtonian fluid with temperature-dependent viscosity and describes the evolution of elevation as an axially symmetric nonlinear diffusion problem (*Methods*). Airy isostasy is assumed, and the vertical temperature gradient is assumed to be constant. The model does not include elastic effects because these operate on timescales much shorter than the relaxation timescales of interest here, e.g., ref. 55 and because crustal plateaus appear to be mostly isostatically compensated (14). Physically, relaxation happens via lateral flow in the lowermost part of the crust, where temperatures are highest and viscosities lowest. The relaxation timescale depends on the effective viscosity, the thickness of the channel in which flow occurs, and the width of the relaxing region, e.g., ref. 56. The role of the mantle is unimportant because its relaxation timescale is short compared to that of the crust (57). Relevant parameter values are given in Table 1.

Results

Fig. 1*A* shows the evolution of a model plateau topography over 450 Myr assuming a dry diabase (mafic) rheology (17). We take the initial elevation (2 km) and radius (300 km) from ref. 11, appropriate to the smallest crustal plateau (Alpha Regio). Initial relaxation, modifying the starting flat-topped profile, is rapid. However, the relaxation rate decreases strongly as a function of time; this is because the rheology is non-Newtonian, so that as the stresses (due to elevation) decrease, the effective viscosity increases. Elevation changes are slower away from the center, partly because of the axisymmetric geometry assumed but also because the flow rate is sensitive to the local crustal thickness. For the particular conditions adopted here, only moderate relaxation occurs, consistent with the model results of ref. 11. A remnant of the initially steep-sided central plateau survives, so this outcome is at least marginally consistent with observed plateau characteristics.

Fig. 1*B* shows the evolution of the same initial conditions but now assuming a dry anorthite (more felsic) rheology (27). The relaxation in this case is much more rapid; the initial sharp-edged plateau morphology is completely destroyed, while the rise of the surrounding terrain would lead to intense extensional deformation. Neither of these results is consistent with observed plateau characteristics. At least for the particular conditions assumed here, survival of crustal plateaus is not consistent with a felsic rheology.

Fig. 2 shows the relaxation time of our model plateau as a function of the heat flux for six different rheologies. This relaxation time (defined as a 30% reduction in the central

Table 1. Parameter values used in the viscous relaxation model

Parameter	Symbol	Value	Units
Surface gravitational acceleration	g	8.9	$m\ s^{-2}$
Crust-mantle density contrast	$\Delta\rho$	500^{+}	$kg\ m^{-3}$
Crustal density	ρ	$2,800^{+}$	$kg\ m^{-3}$
Surface temperature	–	740	K
Thermal conductivity*	–	2	$W\ m^{-1}K^{-1}$

* Thermal conductivity value follows (58).
$^{+}$ For diabase, a density of $2,900\ kg\ m^{-3}$ is assumed, and the crust-mantle density contrast is $400\ kg\ m^{-3}$.
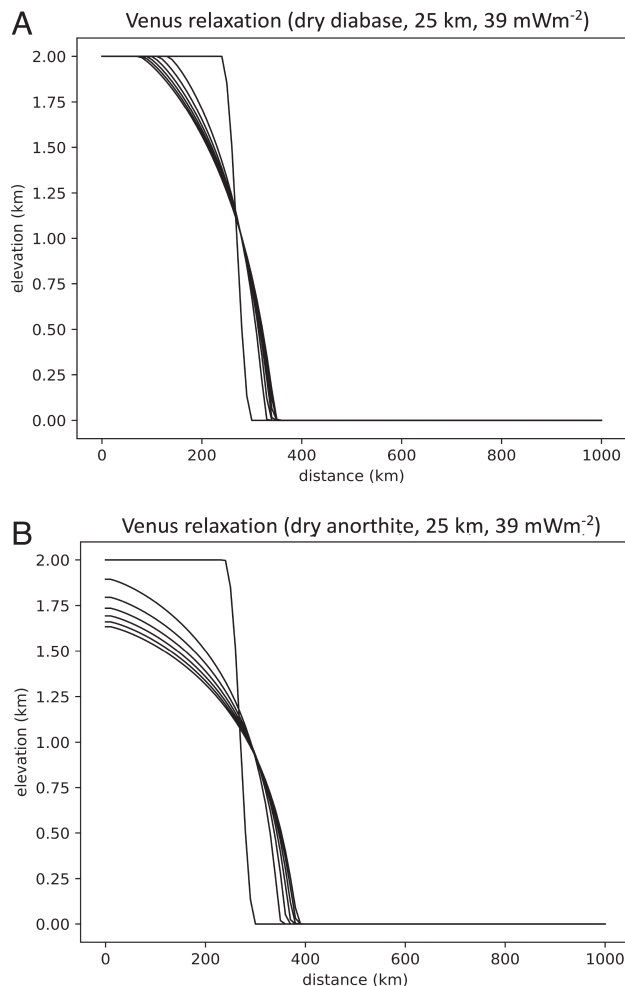

Fig. 1. Elevation profiles plotted at different times for a crustal plateau on Venus. Profile are plotted at 50 Myr intervals from 0 to 450 Myr. Here, a 25-km background crustal thickness and a heat flux of 39 mW m^{-2} are assumed. (*A*) Dry diabase rheology (17). (*B*) Dry anorthite rheology (27).

elevation) should not be less than the estimated age of about 1 Gyr (horizontal dashed line). If the background crustal thickness is 20 km (Fig. 2*A*), neither dry nor wet quartz would permit plateau survival over 1 Gyr. A wet anorthite rheology is only permitted if the heat flux is less than the lower end of the radiogenic value. In this case, Venus would be heating up in the present day. In contrast, dry rheologies, including dry anorthite, could sustain higher heat fluxes, up to ≈80 mW m^{-2} in the case of dry diabase.

If the background crustal thickness is 30 km (Fig. 2*B*), things become a good deal more restrictive. Now, dry anorthite and dry albite would require heat fluxes less than the lower and upper bounds on radiogenic heat flux, respectively.

Fig. 2 shows that rocks containing substantial volumes of quartz, either wet or dry, cannot form crustal plateaus. Thus, these plateaus are unlikely to be strictly granitic. However, anorthite-rich rocks are present on the Moon, and likely on Mars (see above), so we will focus below on this mineral. Dry albite is somewhat stronger than dry anorthite, but since it is commonly found with weaker quartz, the albite rheology is probably less relevant. Potassium feldspar is weaker than anorthite (59), so we are being conservative in focusing on the stronger mineral.

Fig. 3 shows the competing effects of crustal thickness and heat flux on relaxation times for dry and wet anorthite. To retain the same relaxation time, higher heat fluxes require lower crustal

thicknesses. Wet anorthite and dry anorthite require background crustal thicknesses less than 20 km and 29 km, respectively, unless the heat flux of Venus is less than the present-day radiogenic lower bound. A heat flux of 80 mW m^{-2} as suggested by ref. 53 is not consistent with either rheology but could be sustained with a dry diabase rheology and a 20-km-thick crust (Fig. 2A).

We remind the readers that the relevant rheology is that of the lowermost crust, since that is where flow occurs. Dehydration sets in at 600 to 800 °C for feldspars (25), which means that the base of the crust should be anhydrous. On the other hand, an anorthite-rich crust is unlikely to melt at its base, as shown by the dashed white lines in Fig. 3.

Discussion

A possible complication for our results is that plateau crust may concentrate radioactive elements, like the terrestrial continental crust. If heat-producing elements are moved from the mantle into the crust, the temperature at depth will be lower and the maximum allowable heat flux higher. In practice, this effect appears to be small because it is offset by the increase in channel thickness with higher crustal heating (56). For instance, if a 20-km-thick crust receives ten times the mantle concentration
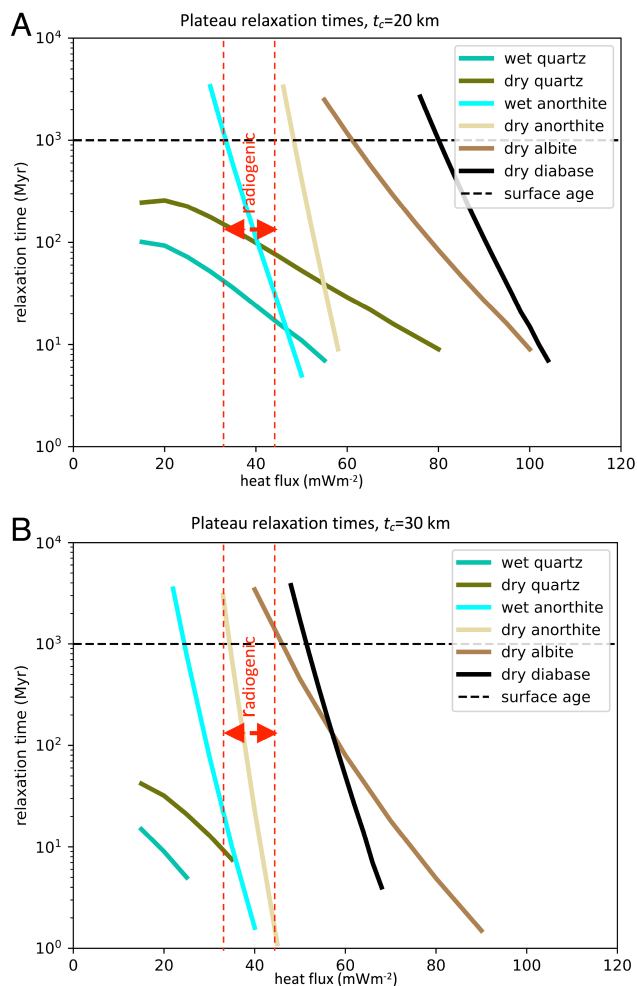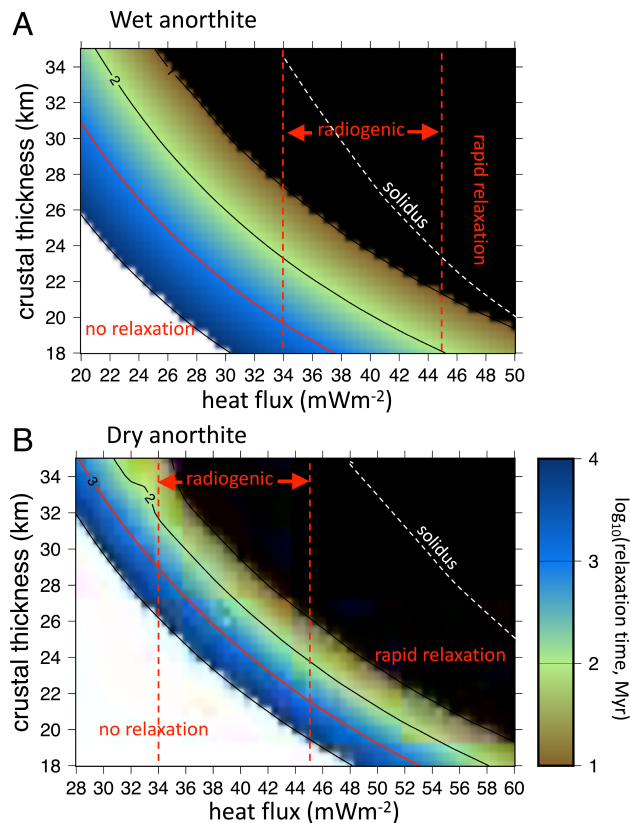
Fig. 3. Relaxation time as a function of background crustal thickness and heat flux. (*A*) Wet anorthite case. The red curve denotes the 1-Gyr contour. The dashed white lines show where the base of the crustal root (11 km deeper than the background crust) reaches the wet anorthite solidus (1,240 °C at 5 kbar) (67), assuming a thermal conductivity of 2 W m^{-1}K^{-1}. (*B*) Dry anorthite case. Here, the dry solidus (1,560 °C) (68) is plotted.

of radiogenic elements (for an internal heat production rate of 0.16 μW m^{-3}), then the surface heat flux required to produce a relaxation time of 1 Gyr increases by only 3% for dry anorthite.

A perhaps more serious complication is that some of the rheological parameters used are quite uncertain. As an example, we used the anorthite data from ref. 27 and not (60) because in the latter case, dislocation creep behavior (which is dominant at the stress levels expected) is barely discernable. Similarly, we generally use monomineralic creep data, rather than construct a composite, polymineralic rock rheology in the style of ref. 53 because doing so would introduce another layer of modeling. Another important question is how dry the minerals tested really are. Even small amounts of water can strongly affect creep properties, so the fact that the nominally dry anorthite of ref. 27 contains 640 ± 260 ppm H means that it will be weaker than a truly dry specimen[†]. Further experimental efforts to characterize rock and mineral rheologies relevant to Venus would be a great help.

On the Earth, crustal plateaus are maintained by the competition between viscous relaxation and compression due to plate tectonics, e.g., ref. 61. We have neglected the effect of far-field compression or of downwellings (62) in this work. If arguments for lateral surface motion on Venus (24) turn out to be correct, our models will need to be modified accordingly. It will also be of interest to use apply more sophisticated models, e.g., ref. 11 to felsic rheologies for comparison with our results. While we do not expect the inclusion of elastic effects to change the

Fig. 2. Relaxation timescale as a function of heat flux for different crustal rheologies (see text). The horizontal dashed line indicates the estimated age of crustal plateaus. The vertical dashed lines indicate the range of heat fluxes expected due to radiogenic decay. (*A*) Background crustal thickness t_c of 20 km. (*B*) Background crustal thickness of 30 km.

[†] For reference, the H concentration of the dry diabase in ref. 17 was <10 ppm.

relaxation timescales (see above), they will likely yield different topography profiles.

Our use of a relaxation time defined by a 30% reduction in topography is very conservative. Fig. 1*A* shows that the steep-sided plateau morphology is lost long before a 30% reduction occurs. This means that all our heat flux estimates are conservatively high. On the other hand, the relaxation time goes roughly as the square of the wavelength, e.g., ref. 56, so a plateau three times as wide would relax an order of magnitude more slowly. For dry anorthite, this would only increase the permitted maximum heat flux by about 5 mW m^{-2} (Fig. 2).

We generally assumed a crustal density of 2,800 kg m^{-3} to resemble that of the terrestrial (felsic) crust (63). For diabase, we assume a higher density which results in a thicker crustal root, higher temperatures at the base of the crust, and thus more rapid relaxation, though the outcomes are not very sensitive to this parameter. By adopting a crustal density at the low end of likely values, we are being conservative in that relaxation could be more rapid and the bounds on heat flux more stringent, than depicted in Figs. 2 and 3.

In our modeling, we have assumed a constant heat flux. If the heat flux has varied over time, what we are really constraining is the highest heat flux recorded: because initial relaxation is so rapid (Fig. 1), even a brief period of elevated heat flux will dominate the total relaxation. Similar comments apply if the surface temperature has varied.

Finally, although we have assumed a plateau age of 1 Gyr, we do not know whether that all plateaus have the same age. The crater data are insufficient to tell, and the stratigraphic evidence is debated. Older plateaus should be more relaxed, other things being equal, but this prediction is not testable with current observations.

Future Progress. The fact that crustal plateaus have not undergone extensive relaxation (11) can be satisfied only by limited combinations of rheology, heat flux, and crustal thickness (Fig. 3). For instance, quartz-rich rheologies can be ruled out (Fig. 2). At present, however, heat flux and crustal thickness are currently too poorly known to make definitive statements. How can progress be made?

Deriving heat fluxes from elastic thickness measurements is likely to always be subject to large uncertainties, as the range of estimates in ref. 15 makes clear. Crustal thickness, however, can be more directly measured from gravity-topography techniques. Thus, the improved gravity measurements that will be provided by the VERITAS (64) and EnVision (65) missions should provide more reliable estimates of this parameter. At the same time, these missions are likely to provide much-improved constraints on plateau mineralogy. Of course, this will provide only a surface measurement, so that (for instance) detection of hydrated felsic minerals would not necessarily imply the presence of such minerals at depth. Indeed, based on Figs. 2 and 3, it would be surprising to find such minerals dominating plateau rheology. If quartz-rich mineralogies are detected, a very thin crust and/or low heat flux would be required (Fig. 2*A*).

It therefore seems likeliest that future mission results will constrain plateau mineralogy and crustal thickness, in which case an upper bound on the heat flux can be derived. If the magnitude of crustal plateau relaxation can be estimated, the local heat flux can be constrained. Since this quantity is so uncertain at present, doing so would represent a significant step forward.

Better gravity data will also reveal whether mountain belts (66) are supported isostatically or by flexure. If the former, these features may provide more stringent relaxation constraints because of their narrower widths and higher elevation contrasts.

Conclusions

If crustal plateaus are in fact made of felsic material, maintenance of their topography over billion-year timescales is not assured. Despite the current uncertainties in crustal thickness and heat flux, quartz-dominated rheologies can be ruled out entirely. Anorthite-dominated rheologies would imply a heat flux that cannot significantly exceed the expected radiogenic value (Fig. 2), in conflict with at least some heat flux estimates (15, 53). We expect that future spacecraft missions will resolve the question of crustal plateau composition; with that information in hand, the survival of plateaus can then be used to infer the likely heat flux range.

Materials and Methods

The viscous relaxation code is described in detail in ref. 54, so only a brief summary is given here. Viscous flow takes place in an axisymmetric geometry where the fluid is non-Newtonian and the viscosity varies with temperature (and thus depth). The evolution of the shell thickness D is given by

$$\frac{dD}{dt} = -\frac{1}{r}\frac{d}{dr}\left(r\int_0^D v dz\right), \quad [1]$$

where r is the radial coordinate, and $v(z)$ is the horizontal flow velocity. This flow velocity is given by

$$v(z) = AE_b\left(g\Delta\rho\frac{\partial D}{\partial x}\right)^n\int_0^z |z'-\alpha|^{n-1}(z'-\alpha)e^{-z'/\delta}dz', \quad [2]$$

where g is the surface gravitational acceleration, $\Delta\rho$ the crust-mantle density contrast, α is a constant of order unity determined by the requirement that $v = 0$ at the surface, and δ is the effective channel thickness in which flow occurs, which is governed by the heat flux and activation energy E_b. The velocity is maximum near the base of the crust and decreases upward owing to the increased crustal viscosity. Airy isostasy is assumed to operate throughout. The rheological parameters A, E_b, and n describe the strain rate of a deformed rock $\dot{\epsilon}$:

$$\dot{\epsilon} = A\sigma^n\exp(-E_b/RT), \quad [3]$$

where T is the temperature, R the gas constant, and σ the stress.

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