GEOPHYSICAL RESEARCH LETTERS, VOL. 19, NO. 13, PAGES 1395-1398, July 6, 1992

THE RELATIONSHIP BETWEEN THE HEIGHT OF A VOLCANO AND THE DEPTH TO ITS MAGMA SOURCE ZONE: A CRITICAL REEXAMINATION

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Abstract. It is commonly assumed that hydrostatic pressure balance arguments can be used to establish a relationship between the maximum height to which a volcanic edifice is able to grow and the depth at which the partial melts providing its magma supply are formed. Such a relationship has been used to infer various aspects of the thermal and stress state of the lithosphere beneath volcanic constructs on Earth, Mars, Io and Venus. We examine the assumptions behind this relationship (which are that: (1) a continuous pressure connection exists between source and summit, (2) the pressure around the magma source is the local hydrostatic pressure dictated by the depth below the geoid, and (3) the melt erupting at the summit has a net positive buoyancy), and show that many of them require geologically unreasonable conditions. We then critically assess the evidence cited in the literature for the relationship and find that there are other factors that may explain the observations. We conclude that volcano heights on the terrestrial planets cannot be related in any simple way to lithospheric thickness or depth to the magma source zone and we review the range of other factors controlling volcano height.

Introduction. It is commonly assumed that the maximum height, measured above the regional geoid level, to which a volcanic edifice is able to grow can be simply related, using hydrostatic pressure balance arguments, to the depth at which the partial melts providing its magma supply are formed [Eaton and Murata, 1960; Vogt, 1974; Epp, 1984; Hartmann, 1983]. This relationship has been used to infer various aspects of the thermal and stress state of the lithosphere beneath volcanic constructs on Earth [Vogt and Smoot, 1984; Decker, 1987; Smith and Cann, 1992], Mars [Carr, 1981], Io [Carr, 1986] and Venus [Nikishin, 1990; Schaber, 1991]. We examine first the assumptions and then the evidence on which this simple relationship is based.

Existing models of the links between the heights of volcanoes, their large-scale internal structures, and the depths to their magma source zones [Eaton and Murata, 1960; Vogt, 1974; Epp, 1984] make the basic assumptions, either explicitly or implicitly, that: (i) a continuous pressure connection via a Newtonian fluid exists within the volcanic edifice between the melt source zone and the vent; (ii) the pressure in the country rocks surrounding the melt source is the local hydrostatic pressure dictated by the depth below the geoid (with no load contribution by the edifice itself); and (iii) the integral over depth of the density difference between the melt and the surrounding rocks is such that the melt erupting at the volcano summit has a net positive buoyancy.

These assumptions are commonly used as follows. Let the depth from the geoid to the top of the melt source zone be H_s (measured positive downwards) and the height of the summit of the volcano above the geoid be H_v (measured positive upwards); the densities of the volcanic edifice rocks, the melt, and the lithosphere in the vicinity of the volcano, averaged over the vertical distances to which they apply, are ρ_v, ρ_m and ρ_l, respectively; g is the acceleration due to gravity. Then on balancing the pressures due to the weights of the magma and the country rocks at the partial melting level:

$$\rho_l \, g \, H_s = \rho_m \, g \, (H_s + H_v) \quad (1),$$

and thus

$$H_s = H_v \, [\rho_m / (\rho_l - \rho_m)] \quad (2).$$

When the melt is on average positively buoyant relative to the lithosphere through which it rises (the commonest case), $(\rho_l - \rho_m)$ is positive and so H_v is also positive, and a volcanic edifice can form. In the case of dense basalts rising through the lower density anorthositic lunar lithosphere, $(\rho_l - \rho_m)$ was commonly negative, and thus H_v must also have been negative. This situation has been used to explain why eruptions only took place in, and could only partly fill, the mare basins [Solomon, 1975; Wilson and Head, 1981; Head and Wilson, 1992a].

Analysis. We propose that each of the three assumptions underlying the above treatment contains flaws, as follows:

(i) Continuous Pressure Connection. There is unlikely to be a continuous pressure connection between the magma source zone and the surface. The evidence from volcanoes such as Kilauea [Ryan, 1987] is that discrete batches of magma ascend from the upper parts of zones of partial melting in the mantle to accumulate in a shallow magma reservoir under the summit of the volcano. These batches move through a region which offers less resistance (both viscous and elastic) than elsewhere since it has by definition been pre-heated by earlier magma batches [Turcotte, 1989]. The stresses generated around a body of buoyant melt are such that each batch probably proceeds as an isolated dike opening a new fracture ahead of it [Weertman, 1971; Shaw, 1980; Spera, 1980; Sleep, 1988], the fracture nearly closing again behind the batch, but leaving a thin ribbon of melt behind [Stevenson, 1982] which then cools to the ambient (sub-solidus) temperature. There is little possibility of a continuous fluid column extending from the source region to the reservoir or to the surface (though a connection from the shallow reservoir to the surface probably exists during eruptions). Even if several successive batches of melt follow the same path, so that each new batch rises along the hot ribbon left by the previous batch, the ribbon will commonly have cooled sufficiently [Delaney, 1987; Bruce and Huppert, 1989] that the magma in it exhibits non-Newtonian rheology [Shaw, 1969]; a finite yield strength will then be involved in transmitting stresses along the ribbon and the hydrostatic condition will not prevail.

(ii) Stress State at the Magma Source. The state of horizontal stress in the country rocks adjacent to the magma source may well be close to hydrostatic. It can be shown [e.g., Jaeger, 1969] that, depending on how rocks are emplaced, the vertical and horizontal components of the stresses within them may differ by as much as the factor $\{(1-\nu)/\nu\}$ ~3, where ν, ~1/4, is the Poisson's ratio of the rocks. If this is so, then the relationship between the pressure in the magma in any kind of source zone and the horizontal stress in the rocks surrounding

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Paper number 92GL01073
0094-8534/92/92GL-01073$03.00

it will be very complex [Sleep, 1988; Tait et al., 1989]. However, Rubin and Pollard [1987] and Dieterich [1988] have advanced arguments for assuming that, at depths of only a few kilometers in volcanic rift zones, the combination of repeated dike intrusion and inelastic creep of the heated country rocks is enough to produce a near-hydrostatic stress field. McGarr [1988] argues that such conditions are, in fact, very widespread in the lithosphere.

However, even if it is assumed that the stress conditions surrounding a magma source zone are indeed purely hydrostatic, it is still difficult to make a case for completely neglecting the load imposed on the source zone by the local volcanic edifice. If this load is included, (1) becomes

$$\rho_l\, g\, H_s + \rho_v\, g\, H_v = \rho_m\, g\, (H_s + H_v) \qquad (3),$$

and so

$$H_s = H_v\, [(\rho_m - \rho_v)/(\rho_l - \rho_m)] \qquad (4).$$

Now, although $(\rho_l - \rho_m)$ can still plausibly be assumed to be positive, $(\rho_m - \rho_v)$ is likely to be negative, since much of the volcanic edifice consists of the solid, cooled (and hence denser) form of its own melt; this implies that no positive volcano can form in the first place! Clearly, the problem can be avoided by assuming that the extra load imposed by the edifice is shared by a finite region at the depth of the source zone, so that only some fraction α of the term $\rho_v\, g\, H_v$ is involved. Equation (4) then becomes:

$$H_s = H_v\, [(\rho_m - \alpha\, \rho_v)/(\rho_l - \rho_m)] \qquad (5),$$

and H_v can be positive if

$$\alpha < \rho_m/\rho_v \qquad (6)$$

where ρ_m/ρ_v, at least for terrestrial basalts, is likely to be of order 0.92. The value which should typically be assumed for α must be based on an assessment of the distribution of crustal stresses in the vicinity of the volcano, which will be a complex function of the surface topography [McTigue and Mei, 1981], the spatial variation of country rock density [Dieterich, 1988] and the extent to which the stresses are inelastic. The fact that the stresses are not simply hydrostatic relative to the geoid, however, ensures the violation of the second of the original assumptions as embodied in equation (1).

A final issue is that the above derivation entirely ignores the origin of the pathway connecting the melt source to the surface. It is effectively assumed that fractures already exist into which melt can flow whenever it is produced. In fact, whether melt utilizes an existing fracture or creates a new one (i.e., forms a dike), elastic stresses, which are inherently non-hydrostatic, are required to dilate the crack and allow magma flow. Typically, the stresses required to allow a dike to form and continue to exist involve an excess pressure of order 10 MPa being present inside the dike, relative to the local horizontal regional stress [Weertman, 1971; Pollard and Muller, 1976; Rubin and Pollard, 1987]. Only if the dike walls behaved in a purely plastic fashion on the time scale of magma eruption would the internal pressure and external stress be equal. It is therefore a basic inconsistency in the original model that elastic conditions must exist in the magma source region to create a continuous dike from the source region to the surface and must then be arbitrarily (and unjustifiably) replaced by plastic conditions operating over very short time scales to justify the pressure-stress equality.

(iii) Magma Always Positively Buoyant. The net buoyancy of magma between its source and the surface is far from guaranteed. For example, sedimentary rocks forming the upper parts of the terrestrial continental crust often have a density lower than that of any basalt, yet basalts do reach the surface in such regions. Similarly, the densities of the upper few km of volcanic edifices such as Krafla in Iceland and Kilauea in Hawaii [Gudmundsson, 1987; Rubin and Pollard, 1987] have densities lower than that of the basaltic liquid from which they are formed - in this case because gas exsolution at shallow depths leads to the eruption of vesicular lavas or pyroclastics. On hydrostatic grounds it would be argued that negative magma buoyancy at shallow depths will not affect the ability of magma to reach the surface directly because it is only the total weight of the magma column relative to the total weight of the surrounding rocks that is important. However, the common presence of magma chambers at shallow depths in basaltic volcanoes shows that magma commonly stalls at the level at which it is neutrally buoyant [Walker, 1987; Ryan, 1987; Walker, 1989]. Thus, in conflict with hydrostatic arguments, the local density contrast between the melt and the country rocks exerts the major control on the level to which magma can rise, and the simple pressure balance calculation implied by equations (1) or (3) does not apply.

Discussion. The issues raised above draw into question the concept of there being a simple relationship between volcano height and depth to the source. However, evidence is commonly cited in the literature that such a relationship exists, and the relationship is frequently used in interpreting volcano heights in various environments. What then is this evidence cited for the relationship, is it truly strongly supporting, and if not, are there alternative explanations?

Eaton and Murata [1960] first proposed such a relationship for the Hawaiian shields and found that for typical densities, the height of Mauna Loa gave a depth to the magma source of ~60 km consistent with the greatest depth at which earthquakes are recorded below the island of Hawaii. Thus they concluded that simple isostatic arguments gave depths consistent with the seismic evidence for the base of the lithosphere which they interpreted to be the beginning of a partial melting zone equivalent to the source region. The lower height of Kilauea volcano relative to Mauna Kea and Mauna Loa, gave shallower source depths because Kilauea is still relatively young and has not yet reached its maximum height. Therefore, the implication of the Eaton and Murata hypothesis is that the depth to the magma source in Hawaii is ~60 km and thus the maximum height that a Hawaiian shield could grow is ~4 km; Mauna Loa and Mauna Kea have thus achieved their maximum heights. More recent petrologic [Wyllie, 1988] and chemical [Sen and Jones, 1990] analyses have shown that the source of the magma in Hawaii is deeper than 60 km, probably in excess of 80-90 km. If this latter depth is correct, then the maximum heights of Hawaiian shields should be in excess of 6.6 km on isostatic grounds, much greater than the current heights of Mauna Loa and Mauna Kea. Why then has Mauna Kea not grown to greater height than it has? One of the most fundamental reasons for this may simply be magma supply: i.e., the volcano has now moved sufficiently far from the center of the hotspot that it is no longer being supplied with magma and thus cannot grow any higher. Therefore, the apparently good fit that Eaton and Murata found for maximum height and magma source depth is not consistent with the most recent estimates of source depths of magma below Hawaii and thus there is no direct evidence that the heights of Mauna Loa or Mauna Kea are being isostatically limited. In addition, a simple argument that the heights are due to magma supply and volcano growth-time over a moving and locally transient source (the stationary hot spot) seems to provide an equally plausible explanation.

Vogt [1974] considerably extended the Eaton and Murata [1960] isostatic arguments. He postulated that if volcano heights were isostatically limited, then the height of a volcano would be a guide to the thickness of the lithospheric plate on which it grew. He found a square-root relationship between volcano height and age of the plate consistent with the idea that plate thickness increases as the square-root of time. He therefore concluded that volcano heights are isostatically limited. If Eaton and Murata's arguments are weakened in light of the new data and alternative explanations discussed above, is there also an alternative explanation for the relationship observed by Vogt [1974]? For example, one could argue (as we did in the case of Hawaii above) that volcano height is simply a function of how long the volcano has had to grow: this hypothesis would predict a cube-root relationship between volcano height and the thickness (age) of the plate on which the volcano grew (the volume of the volcano being proportional to the cube root of any of its dimensions, in this case its height). In order to test this hypothesis, we took Vogt's volcano height data and plotted it as a function of the cube root (rather than his plot of the square-root) of plate age. Our results show a good correlation (with a slightly better correlation coefficient than the isostatic square-root case). Thus, one can argue that a model of volcano height related to duration of magma supply is as reasonable an explanation of the observations as isostatically-limited volcano heights.

Later papers [e.g., Epp, 1984; Nikishin, 1990; Schaber, 1991; Smith and Cann, 1992] generally state the isostatic-limit model as an assumption, do not provide any independent evidence that it is correct, and then interpret their observations in the context of the initial assumption. In summary, the initial assumptions about the depth to the source region on which Eaton and Murata [1960] based their isostatic limit on volcano growth model are likely to be incorrect. In addition, it can be reasonably argued that correlations that are undoubtedly found between volcano height and plate thickness or depth to magma source [e.g., Vogt, 1974] could have causes other than an isostatic one. The simplest alternative hypothesis is that volcano height is related to the total magma supply and time of volcano growth, an hypothesis that Vogt's data fit equally well. In this case, the heights of the volcanoes are not necessarily maximum heights, as is necessary to assume for the isostatic model to apply.

What, then, are other factors that appear to control the height of volcanoes, and how are these manifested on different planets? Factors important during the formation of a volcano include mantle magma supply rate, presence or absence and depth of neutral buoyancy zone(s), eruption frequency, eruption volumes, flow lengths, and total duration of the edifice building phase. Where several volcanic edifices evolve over an extended source, the spatial distribution of major vents may determine the relative magma supply rates to them. Factors important during the modification of volcanoes (which overlaps with the formational phase) include thermal structure and thickness of the lithosphere (and the resultant loading, flexure, and subsidence), large-scale slumping of volcano flanks, and erosion.

On Earth, in the case of Hawaii, magma ascends at an average supply rate of about 3 m^3/s, and pauses at a neutral buoyancy zone at a depth of several km [Ryan, 1987]. Most magma is injected laterally into rift zone dikes, and eruptions from these rift zones, together with summit eruptions, build the edifice vertically. As the edifice increases in elevation, the neutral buoyancy zone and the magma reservoir rise accordingly [see Figure 10, Head and Wilson, 1992b]. As the volcano increases in height, there is a corresponding increase in the load on the lithosphere, and the volcano ultimately subsides several kilometers [Moore, 1987]. During its growth and subsidence, flank failure occurs along broad listric faults and this enhances the formation of flanking rift zones and broadens the profile of the volcano [Swanson et al., 1976; Dieterich, 1988]. The laterally moving lithosphere on which Hawaii is situated eventually transports the volcano off the source region, causing activity to cease, and the volcano further decreases in elevation due to subsidence and erosion.

On the Moon, major shield volcanoes are not present, and the global volcanism rate during the period of mare volcanism [10^{-2} km^3/a; Head and Wilson, 1992a] is comparable to the output of a single eruptive center such as Hawaii on the Earth today [Crisp, 1984]. The density of mare basalt magmas is less than that of the lunar mantle, but greater than that of the highland crust; melts will thus ascend through the mantle, but arrive at a density trap at the base of the crust [Solomon, 1975; Head and Wilson, 1992a]. Buildup of magmatic pressure in this region to levels that exceed local lithostatic pressure can result in dike propagation toward the surface, but most dikes will penetrate only into the lower crust and will not result in surface eruptions [Head and Wilson, 1992a]. In order for magma to reach the surface in dikes, very high magmatic pressures are required to overcome the negative buoyancy in the crust and to keep the dike open to the surface; such dikes will be infrequent in occurrence, will have volumes of 20-60 km^3 [far in excess of volumes typical of terrestrial shield volcano eruptions), and will produce very long flows or sinuous rilles [Head and Wilson, 1991; 1992a]. Thus, lack of shallow neutral buoyancy zones, and the presence of volcanic centers characterized by infrequent eruptions of great volume and length, appears to account for the lack of large shield volcanoes on the Moon [Head and Wilson, 1991].

On Mars, major shield volcanoes are often a factor of three higher than at Hawaii, and are characterized by much larger summit calderas and longer flows [Carr, 1981]. The low atmospheric pressure on Mars (relative to Earth) will favor gas exsolution and pyroclastic eruptions and will nominally result in neutral buoyancy zones at somewhat greater lithostatic pressure levels and, because of the lower gravity, considerably greater depths than on the Earth [Wilson and Head, 1983; 1990]. The stable lithosphere on Mars results in the accumulation of melt products in a single location over a long period of time, providing a significantly greater potential source volume (and thus height) for a single edifice than on Earth. Correspondingly larger reservoirs at neutral buoyancy zones may produce higher volume, longer flows, adding to the breadth of the shield. The relatively thicker elastic lithosphere on Mars during volcano growth [Solomon and Head, 1990] means that flexure and subsidence should be less significant that on Earth.

On Venus, major shield volcanoes are generally of much lower height than on Earth [averaging less than about 2.5 km; Schaber, 1991]. Factors believed to account for this include the higher surface temperature and the correspondingly greater average length of lava flows [Wood, 1979; Head and Wilson, 1986]. In addition, the high surface atmospheric pressure results in neutral buoyancy zones being unlikely to form at the lowest elevations, in the inhibition of vertical migration of neutral buoyancy zones formed at higher elevations, and in the favoring of lateral migration of magma from neutral bouyancy zones creating larger reservoirs and more dispersed surface vents [Head and Wilson, 1992b]. Loading and subsidence must be an additional factor in decreasing edifice height, but quantitative estimates are not yet available.

In summary, a number of factors other than hydrostatic pressure balance arguments influence the maximum height to which a volcanic edifice is able to grow and there appears to be no reason to expect, or indeed any conclusive evidence for, a simple relationship between edifice height and the depth at which the partial melts providing edifice magma supplies are formed. Each situation and environment must be considered separately and a thorough analysis of the factors outlined above must be undertaken in order to determine magma source depths and the causes of volcanic edifice heights.

Acknowledgements. We greatfully acknowledge discussions with G. P. L. Walker and C. E. Johnson. L. Crumpler and two anonymous reviewers provided helpful comments on the first version of the paper. The work was supported in part by the NASA Planetary Geology and Geophysics Program of the Solar System Exploration Division through grants NAGW-713 and NAGW-2185 to JWH at Brown University and partially through grant NAGW-1216 to LW at the University of Hawaii. Thanks are extended to Mary Ellen Murphy for help in manuscript preparation.

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Received: April 9, 1992
Accepted: May 6, 1992