Volcano spacing and plate rigidity

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ABSTRACT

In-plane stresses, which accompany the flexural deformation of the lithosphere under the load of adjacent volcanoes, may govern the spacing of volcanoes in hotspot provinces. Specifically, compressive stresses in the vicinity of a volcano prevent new upwelling in this area, forcing a new volcano to develop at a minimum distance that is equal to the distance in which the radial stresses change from compressional to tensile (the inflection point). If a volcano is modeled as a point load on a thin elastic plate, then the distance to the inflection point is proportional to the thickness of the plate to the power of 3/4. Compilation of volcano spacing in seven volcanic groups in East Africa and seven volcanic groups of oceanic hotspots shows significant correlation with the elastic thickness of the plate and matches the calculated distance to the inflection point. In contrast, volcano spacing in island arcs and over subduction zones is fairly uniform and is much larger than predicted by the distance to the inflection point, reflecting differences in the geometry of the source and the upwelling areas.

INTRODUCTION

The regular spacing of volcanoes in various provinces has attracted the attention of scientists for more than a century (see summary by Vogt, 1974). Two types of explanations were suggested for the regular spacing. (1) Volcanoes form at intersections of fractures whose spacing is approximately equal to the thickness of the lithosphere (Vogt, 1974). (2) The spacing of volcanoes is governed by gravitational instability of a thin low-density and low-viscosity layer, which is embedded within a higher density and a higher viscosity medium. This instability, which is similar to the classical Rayleigh-Taylor instability, tends to form diapirs at regular spacing (Marsh and Carmichael, 1974; Mohr and Wood, 1976; Sigurdsson and Sparks, 1978; Whitehead et al., 1984). Theory predicts that the spacing, d, is a function of the thickness of the unstable layer, h_2, and the viscosity contrast with the more viscous medium, η_1/η_2 ($\eta_1 >> \eta_2$):

$$d = \frac{2\pi h_2}{2.15}\left(\frac{\eta_1}{\eta_2}\right)^{1/3} \qquad (1)$$

(Marsh, 1979). Marsh tested this relation experimentally and suggested that the regular spacing of 70–80 km at island arcs arises from a thin (>1 km), narrow (~30 km) viscous layer on the Benioff zone.

Following a hypothesis developed for the characteristic eruption history of Hawaiian and other midplate volcanoes (ten Brink and Brocher, 1987), I suggest that the spacing of hotspot volcanoes is determined by the flexural stress field induced in the lithosphere by the load of adjacent volcanoes.

MODEL

The flexural bending of the lithosphere under a volcano is accompanied by in-plane stresses, which extend well beyond the perimeter of the volcano and which have opposite signs at the top and the bottom of the plate (Fig. 1A). If the volcano is considered as a point load on an elastic plate, which is supported on a fluid substratum, the deviatoric stresses from this load can be separated into radial and tangential components. The radial component, σ_{rr},

$$\sigma_{rr} = -\frac{Ez}{(1-\nu^2)}\left(\frac{d^2w}{dr^2} + \frac{\nu}{r}\frac{dw}{dr}\right) \qquad (2)$$

(Timoshenko and Woinowski-Krieger, 1959) in the upper part of the elastic plate, T_e, ($0 < z < T_e/2$) will be compressive in the vicinity of the volcano and tensional farther away. E and ν are, respectively, Young's modulus and Poisson's ratio, and w, the deflection of the plate, is the solution to Hertz's equation (e.g., Nadai, 1963, p. 261–267). There will be an inflection point between the two stress regions (whose distance is marked by a heavy circle in Fig. 1B), where the radial stresses throughout the plate fall to zero. The distance to this inflection point increases with the flexural parameter λ, where λ is proportional to $T_e^{3/4}$ (Fig. 1C). Thus, the dis-

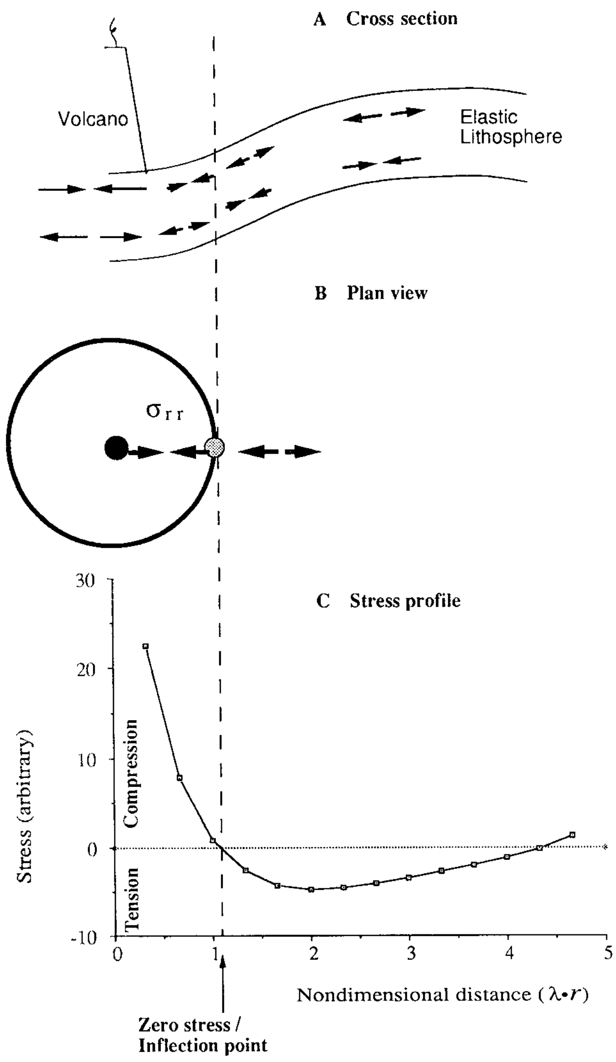

Figure 1. A: Horizontal deviatoric stresses that accompany deflection of thin elastic plate under loading of volcano. B: Distribution of radial stresses (σ_{rr}) in upper part of plate (thick arrows) due to volcano (solid circle). Large circle shows radius to inflection point, where radial stresses change from compression to tension. C: Profile of radial flexural stress due to point load on thin elastic plate overlying fluid substratum, as function of nondimensional radial distance. Inflection point is located at ~1.06λ, where λ is flexural parameter. Model suggests that new volcano (dotted point in B) cannot develop within circle of compressive stresses, but only at minimum distance, which corresponds to distance of inflection point.

tance, r, as a function of lithospheric rigidity (expressed as elastic thickness, T_e), can be defined as

$$r = k \cdot T_e^{3/4}, \tag{3}$$

where k is a constant whose value depends on the choice of Young's modulus and the density difference, $\Delta\rho$, between the upper mantle and the material that fills the flexural depression.

The model suggests that magma is prevented from upwelling to the surface by in-plane compressive stresses in the crust, so the minimum distance for the development of another volcano is the distance to the inflection point of the radial stress component. Although the tangential stress component at the inflection point is still compressive, a stress reduction in only one component is sufficient to establish a path of hydraulic fracturing within the lithosphere that will serve as a conduit to the volcano (Weertman, 1971). Note that the distance to the radial inflection point represents an idealized case, because, in fact, surface eruption is prevented only when the horizontal compressive stresses exceed the pressure head of the magma. Moreover, the contribution of the regional stress field, such as in rift zones, must be added to the flexural stress field.

Two simplifying assumptions were made in the model. (1) The lithosphere behaves elastically. Although the response of the lithosphere is not purely elastic, the distance to the inflection point in an elastic plate model is similar to that in the more realistic elastic-visco-plastic plate model (e.g., Bodine et al., 1981). (2) The volcano can be represented by a point load. For volcanoes with a radius <25 km and elastic thickness greater than their radius, the difference in the distance to the inflection point between a point load and a distributed conical load is <10%, but for larger volcanoes, such as in the Hawaiian-Emperor chain, this assumption may fail.

DATA

Volcano spacing of volcanic groups in diverse tectonic regions and their associated lithospheric rigidity (expressed as elastic thickness) are compiled in Table 1 and compared to the theoretical prediction of the distance to the inflection point in Figure 2. Volcano spacings in seven volcanic groups in East Africa and their standard deviation were measured by Mohr and Wood (1976) (Fig. 2, Table 1). Their criterion for selecting a volcano was an edifice of lava or ash-flow tuffs that is at least 10 km in diameter and was built to an elevation of 500 m or more. In many instances the volcano is lower now because of caldera collapse. Most volcanoes are clearly defined with little overlap between them. The volcanic groups are located both inside and outside the rift valley and differ from each other in many criteria, including geometry, age, size, and composition.

Groups of oceanic volcanoes range from small on-ridge volcanoes (e.g., Lamont seamounts) to giant off-ridge volcanoes (e.g., Louisville and Hawaiian chains). In Hawaii, where volcanoes overlap in space and in time of activity, the distance to the largest nearby volcano should be considered rather than simply the distance to the nearest volcano (e.g., the distance from Loihi to Mauna Loa rather than from Loihi to Kilauea). Ideally, of course, the integrated stress field from several nearby Hawaiian volcanoes should be calculated (see ten Brink and Brocher, 1987, Fig. 14). Two additional problems should be considered in the case of the Hawaiian-Emperor data. (1) The measured spacing of volcanoes (Jackson et al., 1972; Vogt, 1974) was often based on inaccurate bathymetry (e.g., the central Emperor Ridge; Smoot, 1982). (2) The modeling of many of the volcanoes as point loads is inappropriate because of their large basal radii.

Volcano spacing in the Quaternary andesitic volcanoes of many convergent margins appears to be uniform (~70 km; Alaska Peninsula and the Aleutian Islands, Marsh and Carmichael, 1974; Indonesian arc, Scotia arc, Marsh, 1979; see Table 1), but a calculation of elastic thickness exists only for the northwestern United States (Table 1). The calculated elastic thickness for the Wanganui basin (20 km), which lies at the southern tip of the Quaternary volcanic line of the North Island of New Zealand, is consid-

TABLE 1. SPACING OF VOLCANIC GROUPS AND THEIR ASSOCIATED ELASTIC THICKNESSES

Group (See Fig. 2)	Name	Volcano spacing* (km)	Elastic thickness† (km)	Method of calculating thickness	Reference for spacing	Reference for elastic thickness
1	Ethiopian Plateau	109 ±22	57 ±12	Coherence	Mohr and Wood (1976)	Ebinger et al (1989)
2	Eastern Uganda Plateau	72 ±9	40 ±10	Coherence	Mohr and Wood (1976)	Ebinger et al (1989)
3	Harar (Somalian) Plateau	70 ±10	35 ±5	Coherence	Mohr and Wood (1976)	Ebinger et al (1989)
4	Ethiopian Rift	43 ±13	14 ±2	Coherence	Mohr and Wood (1976)	Ebinger and Harding (unpublished)
5	Kenya Rift	42 ±11	25 ±3	Coherence	Mohr and Wood (1976)	Bechtel et al. (1987), Ebinger et al (1989)
6	Dubbi line, Afar	19 ±6	9 ±2	Coherence	Mohr and Wood (1976)	Ebinger and Harding (unpublished)
7	Erta-Ali line, Afar	10 ±3	9 ±2	Coherence	Mohr and Wood (1976)	Ebinger and Harding (unpublished)
8	Loihi-Mauna Loa	70 ±4	30 ±7	Seismic and gravity	Watts and ten Brink (1989)	Watts (1978),
9	Hawaiian chain	72 ±37(a)	30 ±7	Seismic and gravity	Vogt (1974)	Watts (1978), Watts and ten Brink (1989)
10	Northern Louisville Ridge	46 ±19(b)	15 ±5	Gravity	Lonsdale (1988)	Watts et al. (1988)
11	Southern Louisville Ridge	69 ±28(a,b)	35 ±8	Gravity	Lonsdale (1988)	Watts et al. (1988)
12	Southern Cook Islands	40 ±11	16 ±2	Raised atolls	D. Epp, unpublished	McNutt and Menard (1978)
13	Marquesas Islands	54 ±10	20 ±2	Seismic and gravity	from Monti and Pautot (1973)	P. Filmer, personal commun., 1990
14	Ojin seamount, Emperor Ridge	58 ±9	17	Gravity	from Smoot (1982)	Watts and Ribe (1984)
15	Lamont Seamounts, East Pacific Rise	10 ±3	0-5	Stacked gravity and bathymetry	from Barone and Ryan (1990)	Madsen et al. (1984)
16	North Island, New Zealand	80 ±10	20	Seismic and basin modeling	T. Stern, personal commun. 1990	Stern and Davey (1989)
16	Cascade Range, U.S.	90 ±22	12 ±4	Coherence	Marsh and Carmichael (1974)	Bechtel et al. (1990)
17	Tharsis Montes, Mars (Excluding Pavonis Patera)	650-780	100-400(c)	See comment	from Scott and Carr (1978)	
18	Elysium Mons, Mars	370-440	100-400(c)	See comment	from Scott and Carr (1978)	

* Volcano spacing: (a) Spacing > 150 km was excluded from the statistics. (b) Averaged separately north and south of the break in the trend of the ridge at 37.5° S, because this break is analogous to the bend in the Hawaiian-Emperor Ridge (Watts et al., 1988).
† Elastic thickness: (c) Estimates range 110-260 km (from long-wavelength loading of Tharsis Rise; Willemann and Turcotte, 1982), 100-400 km (from comparison of predicted stress trajectories to observed tectonic features; Banerdt et al., 1982), and >150 km for Olympus Mons (Thurber and Toksoz, 1978; Comer et al., 1985), 20-50 km (distance to concentric grabens; Comer et al., 1985).

ered an upper limit for the volcanic line (Stern and Davey, 1989).

DISCUSSION

Control on Spacing

The close agreement between the observed volcano spacing in East Africa and in oceanic hotspots and the calculated distance to the inflection point (Fig. 2) suggests that the spacing is determined by flexural stresses that are induced by the load of adjacent volcanoes. With the exception of large overlapping volcanoes, volcano spacing can be fit by equation 3, k varying between 6.06 and 4.76 (for r and T_e in kilometres), which corresponds to a reasonable range of values for Young's modulus, E, and the density difference, $\Delta\rho$ (for $r = 6.06$, $E = 7 \times 10^{10}$ N·m, $\Delta\rho = 600$ kg/m^3; for $r = 4.76$, $E = 4 \times 10^{10}$ N·m, $\Delta\rho = 900$ kg/m^3) (Fig. 2).

In comparison, the observed volcano spacing in subduction zones is much larger than predicted by their flexural rigidity (Fig. 2). The misfit between the observed and the calculated volcano spacing in subduction zones can be explained in several ways. (1) Large regional compressive stresses in the upper plate overprint the flexural stresses. (2) Rapid variation in flexural rigidity above Benioff zones makes regional estimates difficult. (3) Spacing in subduction zones is determined by a different mechanism, such as the gravitational instability of a thin viscous layer (discussed in the Introduction). The regular spacing of ~70 km in many island arcs and subduction zones precludes, in my opinion, the first two explanations. The stress field in subduction zones is known to vary between relative tension and compression (e.g., Jarrard, 1986), and the thermal regime, which controls lithospheric rigidity, is variable and usually high. Gravitational instability, however, is a viable mechanism because the viscosity ratio between the lithosphere and the melt layer and the thickness of the melt layer is expected to be fairly constant (equation 1).

The explanation of gravitational instability cannot, however, be applied to hotspot volcanoes. Spacing among different groups of hotspot volcanoes (including East Africa) varies by almost one order of magnitude, which, according to equation 1, should correspond to variation by one order of magnitude in the thickness of the melt layer or by three orders of magnitude in the viscosity ratio. Moreover, the trend of the variation is contrary to common sense: the gravitational instability hypothesis will predict a thinner melt layer or a smaller viscosity contrast inside the rift zone than outside because the observed volcano spacing is smaller inside the rifts. Therefore, different explanations must be applied for volcano spacing in hotspots and in subduction zones.

Implications for the Geometry of Upwelling

The different explanations for volcano spacing imply different source and upwelling geometries for these two provinces. Shallow magma chambers under volcanoes in subduction zones are supplied by diapirs, which rise from a thin and narrow melt layer on top of the subducting plate (Marsh, 1979). It is envisioned that hotspot volcanoes are replenished by oblate zones and/or conduits of hydrofractures (e.g., see Ryan, 1988, Fig. 24), which connect the volcano and a large-volume source zone at the base of the lithosphere. The inferred plumbing system under Hawaii is an example of this geometry. The deepest seismicity under Hawaii (40–60 km) is enclosed within a belt 80 km long and ~20–30 km wide located under the southeastern coast of Hawaii (Ryan, 1988). The northeastern-southwestern trend of this belt is perpendicular to the radial stress component, as expected from the distribution of flexural stresses at the tip of the island chain. From this belt of seismicity rises the conduit feeding Kilauea (as well as the conduits for Mauna Loa and Loihi), which is cylindrical to oblate in shape and 3–9 km in diameter (Ryan, 1988).

Volcano spacing does not correlate with plate velocity (in the hotspot frame) because Africa is almost stationary. Among the Pacific hotspots listed in Table 1, the northern Louisville Ridge has the slowest velocity, followed, in order of increasing velocity, by the southern Louisville Ridge (Lonsdale, 1988), the Marquesas (McNutt et al., 1989), the Emperor chain, and the Hawaiian chain (Clague and Dalrymple, 1987), which is the fastest. The lack of correlation contradicts the suggestion by Skilbeck and Whitehead (1978) that the interaction of asthenospheric horizontal flow and a buoyant plume produces an inclined channel along which discrete plumes develop and rise to become regularly spaced island chains. Rather, the effects of asthenospheric flow and plate motion are to disperse the mantle plume and to produce a large source area in the lower lithosphere (Sleep, 1990). The migration of the volcano away from the source of magma is, nevertheless, expected to result in tilting of the magma conduit with time as observed in Kilauea, where the conduit is tilted by ~20° between depths of 12 and 37 km (Ryan, 1988). The effect of conduit tilting is to reduce the pressure gradient that drives the flow in the conduit. The tilting, coupled with increased in-plane compression under the volcanic load, may ultimately cause the conduit to shut and a new more vertical conduit may break near the inflection point where the ambient stress field is less compressive and the pressure gradient is larger.

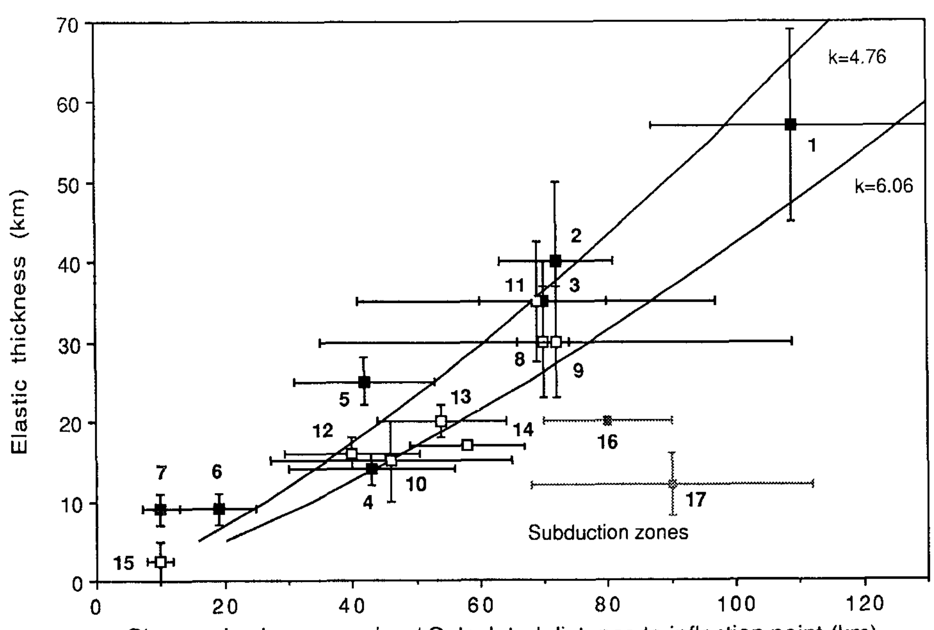

Figure 2. Comparison between data and model. Numbered squares with error bars are volcano spacing in groups of volcanoes in East Africa (solid) and in oceanic hotspots (open) plotted against elastic thicknesses in these areas. Numbers correspond to Table 1. Two continuous curves: Calculated distance to inflection point (zero radial stress) for point-load model as function of elastic thickness of plate. Two curves correspond to two values of *k*, which bracket range of reasonable values of Young's modulus (4–7 × 10^{10}) and density contrasts (600–900 kg/m^3; see text for explanation).

Volcano Spacing as a Predictive Tool

The relation between volcano spacing and plate rigidity provides a simple tool for the evaluation of the elastic thickness of the lithosphere at the time of loading. For example, the spacing of volcanoes on Mars may be used to evaluate the elastic thickness of the Martian lithosphere.

Because of their large size, Martian volcanoes were modeled as multiple disk loads on a thin, spherical, elastic shell (Solomon and Head, 1979) rather than point loads. On the basis of their intervolcano distance (Table 1), elastic thicknesses of 300 km and 130 km, respectively, are predicted for the Tharsis and Elysium volcanic groups, which are consistent with most estimates of elastic thickness for Mars (Table 1) except that of Comer et al. (1985).

The relation between the elastic thickness of the oceanic lithosphere and its age during loading (Watts, 1978) can be modified to create a similar relation between volcano spacing, r, in kilometres, and the age of sea floor at the time of the emplacement of the volcanoes, t (in m.y.).

$$r = k \cdot c^{3/4} \cdot t^{3/8}, \tag{4}$$

where c is a constant whose value is 4.2 (McNutt, 1984) or 2.7 (Calmant et al., 1990). This relation suggests that the distribution of volcanoes on the ocean floor is influenced by the age of the sea floor at the time of volcanism. In the case of several episodes of volcanism, millions of years apart, the true distribution of volcanoes with age is expected to be complicated by the fact that older volcanoes which formed on a younger age sea floor also modify the stress field for volcanoes that erupt many millions of years later. Further complication may arise in the presence of a regional stress field.

CONCLUSIONS

The model presented in this paper suggests an interaction between the conduits of upwelling magma from a hotspot source and the stress field in the upper part of the lithosphere, which is created by the products of this upwelling. Volcano spacing, according to the model, depends on the thermomechanical properties of the lithosphere and can therefore also serve as a simple tool to deduce these properties. The inability of the model to predict volcano spacing in subduction zones suggests that volcano spacing there is probably controlled by localized magma generation and diapirism, and not by the near-surface mechanical properties.

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