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Geological and Geophysical Parameters of Mid-Plate Volcanism [and Discussion]

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Geological and geophysical parameters of mid-plate volcanism

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Although volcanism occurs mainly at oceanic ridges and subduction zones, it also takes place within continental plates. Mid-plate Cainozoic volcanism is widespread in Africa where it is alkaline in composition, is characteristically associated with uplift and faulting, and is almost completely restricted to non-cratonic areas. Cratons in Africa are characterized by low elevations, heatflow less than 45 mW m^{-2}, lithospheric thicknesses greater than 200 km and higher than average seismic velocities. In contrast, non-cratonic regions stand higher and have greater heatflow and lower lithospheric thicknesses and seismic velocities. There is a clear time correlation between the pause at *ca.* 45 Ma in the African apparent polar wander path and the outbreak of volcanism at *ca.* 35 Ma. Accepting that sub-lithospheric heat input is regionally variable, geophysical modelling indicates that the spatial and temporal distribution of uplift-associated mid-plate volcanism can be almost entirely explained in terms of plate thickness and velocity. We examine, quantitatively, conductive models that determine the thermal disturbances in the lithosphere explicitly in terms of perturbation strength, plate thickness and plate velocity. For example, for a plate over 200 km thick, a relatively modest movement (> 2 cm a^{-1}) will suppress the upward propagation of most sub-lithospheric thermal anomalies, thereby precluding mid-plate volcanism. Where such thermal anomalies are confined to the base of the lithosphere, uplift alone, without surface volcanism, would result. The thinning of the lithosphere predicted by the theoretical models can be accomplished by deep, strong thermal perturbations or by upward migrating perturbations of lesser strength. The latter is consistent with configurations for the East African–Ethiopian lithothermal systems derived from geophysical and geochemical data. Mesozoic volcanism of Gondwanaland was more widespread and voluminous, and erupted through both cratons and non-cratons, thereby suggesting that the causative thermal anomalies were much more vigorous than those in the Cainozoic.

1. Introduction

Although most of the Earth's magmatism occurs at or adjacent to plate margins, volcanic activity has repeatedly occurred within both oceanic and continental plates. Here we study African Cainozoic volcanism to investigate the character and distribution of mid-plate volcanic activity and how it is related in time and space to plate motion and lithospheric thickness.

First we identify and justify, on both geological and geophysical data, the structural subdivision of African lithospheric régimes into cratonic and non-cratonic (Pan-African) and consider the depth to which such contrasts could exist. We then examine the effects, in terms of surface heatflow and vertical movement, that sub-lithospheric thermal disturbances could cause, and then consider quantitatively the effect of these disturbances on the configuration of the base of the lithosphere; several models relating to varying plate thickness and velocity are presented. Lastly, in discussion, we attempt to relate these models to the lithospheric structure beneath various active volcanic areas.

2. Geological framework

Africa consists of numerous Cainozoic volcanic areas, seemingly unrelated to plate margin activity (figure 1), which range from small isolated volcanic cones (e.g. J. Uweinat and Bayuda) through large volcanic complexes with composite volcanoes and plateau lavas (e.g. Tibesti, Haruj and J. Marra) to the vast volcanic provinces of East Africa and Ethiopia. Smaller areas of Cainozoic volcanism are scattered across southern and southwest Africa, and Madagascar. In volume, the volcanic rocks of East Africa and Ethiopia total *ca.* $10^6\,km^3$ – about 95 % of all African Cainozoic volcanic products.

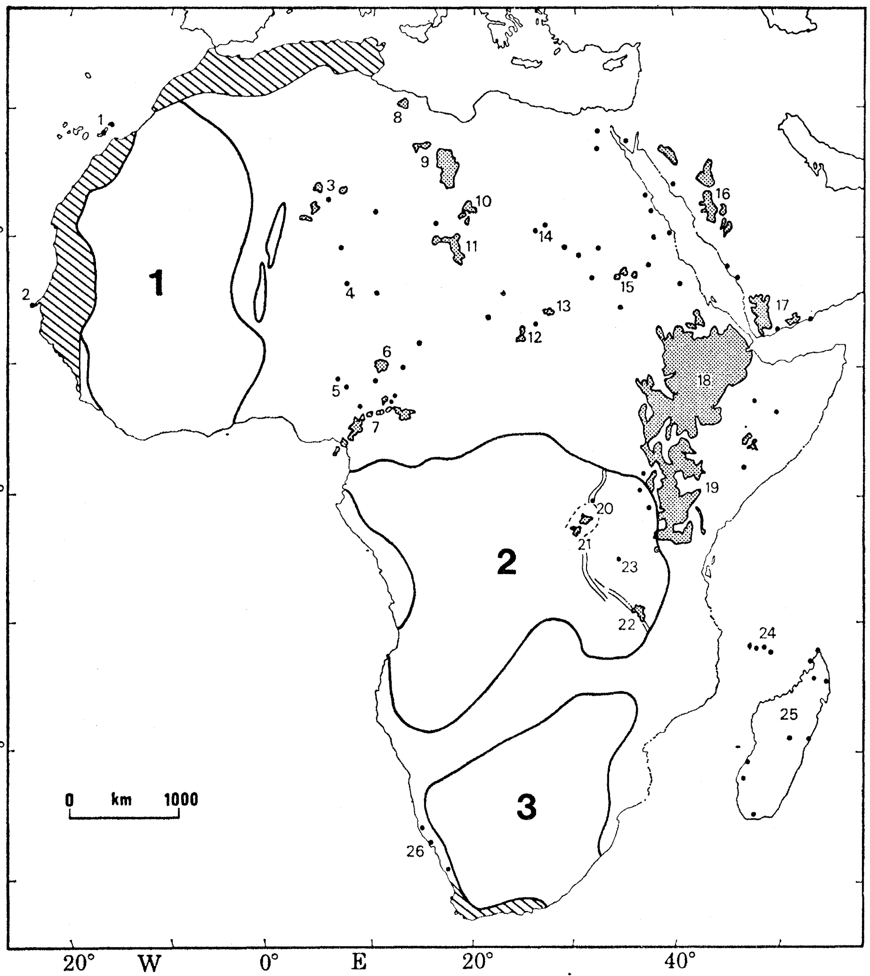

Figure 1. Cainozoic volcanism and structural units of Africa. Extensive Cainozoic volcanic fields are shown stippled, and smaller areas of Cainozoic volcanism are shown as single dots. Individual volcanic areas are as follows: 1, Canary Islands; 2, Dakar; 3, Hoggar; 4, Air; 5, Jos; 6, Biu; 7, Cameroon Line; 8, Tripolitania; 9, J. Haruj; 10, Eghei; 11, Tibesti; 12, J. Marra; 13, Meidob; 14, J. Uweinat; 15, Bayuda; 16, West Arabia; 17, South Arabia and Aden; 18, Ethiopia; 19, East Africa (Kenya) Rift; 20–22, West African Rift with 20, Birunga; 21, Kivu; 22, Rungwe; 23, Igwisi Hills; 24, Comores; 25, Madagascar; 26, Southwest Africa. Thin lines outline the trend of the Western Rift. Cratons are shown in outline as follows: 1, Northwest Africa Craton; 2, Congo Craton; 3, Kalahari Craton. Craton boundaries are not everywhere accurately defined; rocks of cratonic age are known from isolated locations in belts affected by later orogenies. The hatched areas have been affected by post-Pan-African orogenies and volcanics associated with these young fold belts are not shown. (After Thorpe & Smith 1974, fig. 1.)

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Structurally, Africa can be divided into (1) cratons, areas which generally are topographically lower than the global continental mean elevation and have been unaffected by regional tectono-thermal activity during the last 1100 Ma, and (2) non-cratonic areas which generally stand near or above the continental mean elevation and were affected by the 700–400 Ma, Pan-African tectono-thermal event (Kennedy 1964). The distribution of volcanics in relation to these major structural units is shown in figure 1, where their almost total concentration in non-cratonic areas, first noted by Kennedy (1965), is evident (cf. Thorpe & Smith 1974). The only exceptions are in the Western Rift and the Igwisi Hills, which are volumetrically insignificant, petrologically atypical, and an interesting exception to the general rule.

African mid-plate volcanics are characteristically alkaline and include a variety of lineages with varying degrees of silica-saturation and under-saturation. Common soda-rich alkaline associations include (*a*) alkali basalt–trachyte–peralkaline rhyolite; (*b*) alkaline basalt–trachyte–phonolite; (*c*) nephelinite–phonolite and (*d*) carbonatite series. Potash-rich alkaline associations are characteristic of volcanic areas of the West African Rift and the Igwisi Hills, which have many kimberlitic features (Kennedy 1965; Dawson 1971) are regarded as being formed at depths of 150–200 km (Reid *et al.* 1975). 'Transitional' or tholeiitic rocks occur as minor, late-stage members of several mid-plate volcanic provinces (Jos: Wright 1972; Kivu: Denaeyer & Schellink 1965; Tibesti: Vincent 1970), and as substantial components of the volcanics in the Afar (Barberi *et al.* 1975; Raschka & Muller 1975) and elsewhere in Ethiopia (Jones 1976).

There is a general spatial and temporal relation between mid-plate volcanism, domal uplift and rifting (Le Bas 1971; Gass 1970, 1972). Many volcanic areas, for example those of the Hoggar, Tibesti, J. Marra, the Bayuda and the volcanics of the East African–Ethiopian rift system, are all located on crystalline basement which experienced uplift of between 1–3 km extending over areas of 200–3000 km diameter before or during volcanic activity. The East African–Ethiopian rift system has overlapping areas of uplift and volcanism, and smaller discrete areas of volcanism, within the system, are commonly associated with lesser topographic culminations (Burke & Wilson 1972). However, despite well documented association of uplift with mid-plate volcanism it is clear that there is no ubiquitous relation. Indeed, several important volcanic fields, such as J. Haruj and Tripolitania, were erupted into Lower Cainozoic sedimentary basins. Conversely, some areas, which have experienced both local and estensive Cainozoic uplift (e.g. the Drakensberg), are devoid of Cainozoic volcanism.

It is also well established (Le Bas 1971; Gass 1970, 1972; Bailey 1972) that volcanism is associated with major faulting and rifting that often follow older pre-Cainozoic structural trends (McConnell 1974). The link between faulting and volcanism is clear in Air, Tibesti, Jos, Biu, the Cameroon line and, perhaps best of all, in the Ethiopian, Eastern and Western rift systems of East Africa. In the Red Sea and Gulf of Aden, rifting and continental separation have taken place. However, as with uplift, there is no unique relation between rifting and volcanism. Parts of the Eastern and much of the Western rift are devoid of volcanics and the Zambian part of the central African Plateau, an uplifted area with active faulting, is devoid of volcanic products. Yet two of Africa's largest Quaternary volcanoes, Mounts Kenya and Kilimanjaro, lie some 100 km east of the Eastern rift and are not associated with any major fracture zone.

A genetic progression has been proposed (Burke & Dewey 1973) from 'unrifted' domes (e.g. Tibesti, J. Marra) through rifted domes (e.g. the Kavirondo region of the East African rift) to triple-spreading ridge junctions of the Afar type. Most of the areas (see figure 1) have not passed

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the first of these stages despite some 30 Ma of volcanic activity. Only in the case of the Red Sea and the Gulf of Aden have uplift, rifting and volcanism (not necessarily in that order) culminated in continental separation and the development of new oceanic constructive margins.

3. Geophysical framework

Here, those geophysical parameters which might provide a control on the processes that culminate in mid-plate volcanism are discussed. Because of the clear significance of the geological cratonic–non-cratonic subdivision we concentrate on geophysical parameters which relate to this demarcation. These include seismic, heatflow and gravity variations. Seismic data include surface, P and S wave velocity and attenuation studies. Although only geographically limited surface wave investigations have been made in Africa (see Gumper & Pomeroy 1970), they have proved to be a powerful means of distinguishing between cratonic and non-cratonic areas in other continents. Such studies use Rayleigh and Love wave dispersions as a means of distinguishing upper mantle structure in these contrasted continental provinces. Surface wave experiments in South America (Alexander & Sherburne 1973), North America (Wickens 1971; Biswas & Knopoff 1974) and Australia (Goncz & Cleary 1976) all indicate that shield areas are characterized by the absence, or weak development, of an S-wave low-velocity zone and that this character persists to depths of up to 400 km (Alexander & Sherburne 1972, 1973; Alexander 1974, 1975). So although Africa lacks continent-wide, surface-wave dispersion measurements, correlations elsewhere suggest that fundamental differences between African cratonic and non-cratonic areas could be identified by this technique (cf. Goncz & Cleary 1976, fig. 5).

Seismic data available for Africa include P and S wave velocities, P-wave travel-time residuals and S*n* wave attenuations which all provide information on the properties and thickness of the lithosphere. Regional variation in P travel-time residuals (Toksoz, Arkani-Hamed & Knight 1969) indicate negative residuals in cratonic areas, and recent data for residuals adjacent to the East African rift system show the presence of positive residuals in this volcanic area (Long, Backhouse, Maguire & Sundar Lingham 1972; Long 1976; Long & Backhouse 1976; Fairhead 1976; Green 1976.) Similar positive residuals have also been reported in Ethiopia as being most marked in the Afar area (Berckhemer *et al.* 1975). The overall correlation of negative residuals with cratonic areas is consistent with the regional variations of upper mantle P-wave velocity (P*n*), for it is known that the Afar and East African volcanics are associated with lower P-wave velocities in the upper mantle (Arkani-Hamed & Toksoz 1968; Long 1976; Berckhemer *et al.* 1975).

Data for S-waves are represented by studies of regional S*n*-wave attenuation (Molnar & Oliver 1969). This study shows a broad association of efficient S*n* transmission with cratonic areas (Molnar & Oliver 1969, fig. 1). In Africa, Gumper & Pomeroy (1970) found S*n* to propagate well over paths that do not cross either the Red Sea or the East African rift north of 10° S. For those regions of poor propagation, they infer a 'gap' in the high-velocity mantle portion of the lithosphere. On a wider scale, significant lateral heterogeneity of the upper mantle has been demonstrated by study of ScS travel-times (Sipkin & Jordan 1975, 1976). Again, these studies indicate longer travel-times (lower velocities) for continental regions with Phanerozoic (generally non-cratonic) orogenic histories than for Precambrian (cratonic) continental regions (Sipkin & Jordan 1976). These results also suggest that continental structure might persist to depths exceeding 400 km (Sipkin & Jordan 1975, 1976; Jordan, 1975).

Regional differences in continental heatflow correlate directly with the geologically fundamental cratonic–non-cratonic subdivision. Heatflow through cratons is low (generally less than 45 mW m^{-2}) in comparison with that through non-cratonic continental areas (Hamza & Verma 1969; Polyak & Smirnov 1968). Figure 2 shows heatflow variation over Africa by the use of observational and synthetic data from Chapman & Pollack (1975*a*) but contouring directly on 5° × 5° mean values, so providing a more detailed picture of heatflow variation than that given by the smooth, spherical harmonic representation (Chapman & Pollack 1975*a*, fig. 7). It will be noted that heatflow for non-cratonic areas in Africa is 50–100 mW m^{-2}.

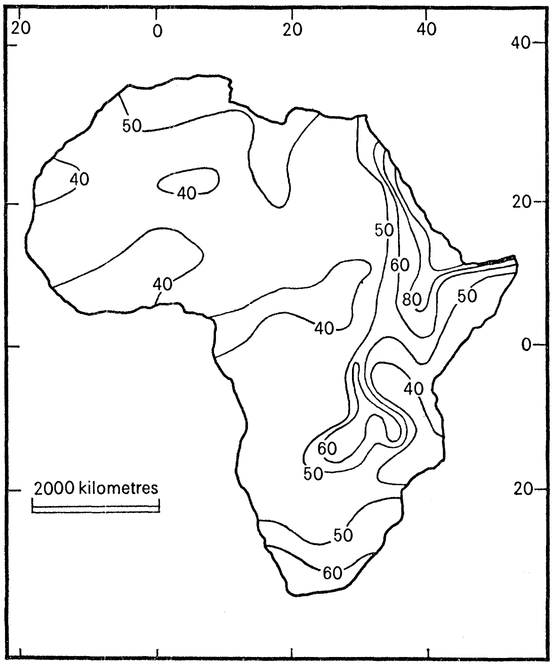

FIGURE 2. Heatflow variation over Africa. Contours (in mW m^{-2}) controlled by measurements in much of the area east of a line from Cairo to Capetown, and in limited areas of West Africa; elsewhere, heatflow estimated from empirical relation between heatflow and age of last tectono-thermal event.

Other geophysical observations also reflect the gross tectonic characteristics. A well defined negative Bouguer gravity anomaly exceeding 120 mGal in magnitude covers the Afar–East African rift region and extends in an arc south-westwards, overlapping the Congo and Kalahari cratons, to the west African coast at 12° S (Slettene, Wilcox, Blouse & Sanders 1973; Girdler 1975). There are also electrical conductivity anomalies in several non-cratonic areas. These are around the Kenya rift (R. J. Banks, personal communication), between and to the east of the Kalahari cratons (de Beer, Gough & Van Zijl 1975) and to the south of the Kalahari cratons in the Cape Fold Belt (Gough 1973). The conductivity anomaly between the Kalahari and Congo cratons is in an area with heatflow about 50 % greater than that of adjacent cratons (Chapman & Pollack 1975*b*). These data have been interpreted in terms of lithosphere fracturing and/or thermal highs in the upper mantle.

Seismic, heatflow and gravity data all imply variation in lithospheric physical properties and thickness. The implication is that an asthenosphere is absent or poorly developed below cratons but distinct and well developed beneath non-cratonic areas. These data have been incorporated in recent quantitative models which indicate a regional variation in African lithosphere thickness (Pollack & Chapman 1977; Fairhead & Reeves 1977). In addition, Pollack & Chapman (1977) have constructed regional geotherms parametric with surface heat flow. The thickness of the lithosphere is then taken as the depth at which the temperature reaches a specified fraction of the mantle solidus. This fraction ranges from about 85 to 100% of the solidus, depending on the amount and proportions of the various volatiles present. Figure 3*a* depicts the thickness of the African lithosphere determined from heatflow data in figure 2.

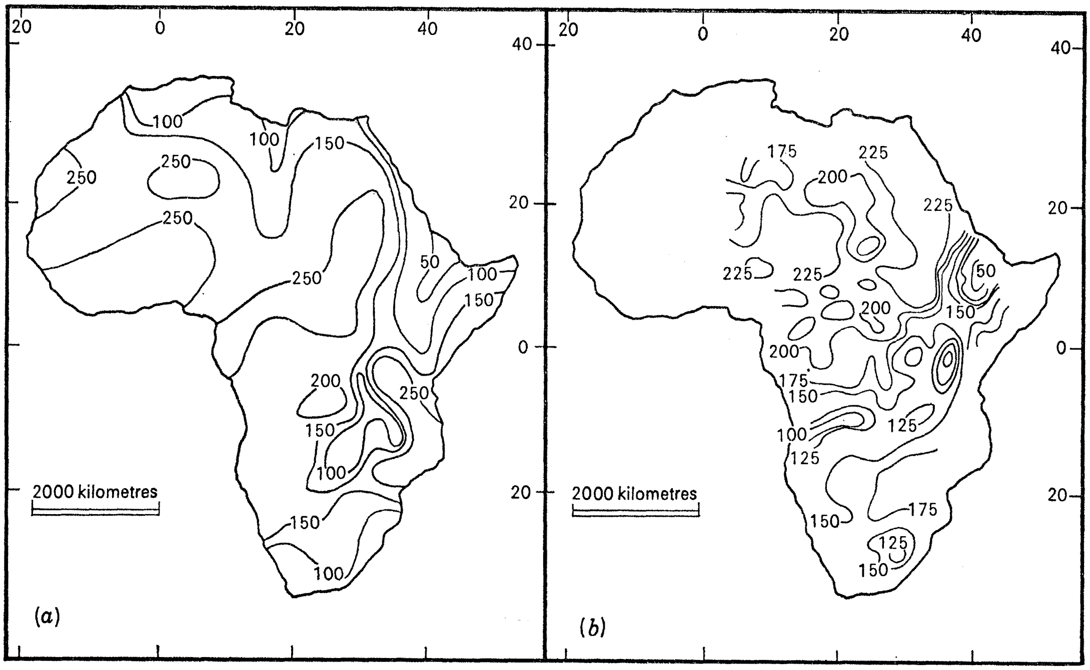

Figure 3. Thickness of the African lithosphere (*a*) based on thermal data of figure 2, (*b*) based on interpretation of teleseismic delay times and the regional Bouguer anomaly map (J. D. Fairhead and C. V. Reeves 1977).

Another means of determining the thickness of the African lithosphere has been devised by Fairhead and Reeves (personal communication; 1977) who use teleseismic P-wave travel-time anomalies with the surface elevation and Bouguer gravity anomaly data (see figure 3*b*). Figures 3*a* and 3*b* show considerable agreement. Both indicate lithosphere thinner than 150 km extending in an arc from Afar, through the East African rift and terminating in Zambia–Angola. Thicker lithosphere (> 200 km) characterizes large areas of west Africa, north of the equator, in both models and also part of the Congo craton. Specific differences between the models can be accounted for by a variety of factors, but considering that figures 3*a* and 3*b* are based on independent data sets, agreement between the two models is good and is here used in considering the development of thermal systems responsible for African Cainozoic volcanism.

To summarize, briefly, the fundamental contrast between cratonic and non-cratonic regions of Africa, we observe that the cratonic regions are characterized by heatflow less than 45 mW m^{-2}, elevations generally less than the global continental mean elevation, lithospheric thickness greater than 200 km, and higher than average seismic velocities. In contrast, the non-cratonic regions of Africa, to which Cainozoic volcanism is virtually restricted, are characterized by heatflow greater than 45 mW m^{-2} and topographic elevation near or greater than the continental mean, and lesser lithospheric thicknesses and seismic velocities. Although we do not develop the arguments as to why this contrast exists, we believe its origin lies in a small, residual thermal effect dating from the Pan-African.

4. Development of lithothermal systems

There is no reason why sublithospheric temperatures and heat flux should be spatially uniform. Significant thermal perturbation could be produced by (i) chemical plumes resulting from radio-element inhomogeneities in a static lower mantle (Anderson 1975); (ii) more or less constricted convection in a mobile upper mantle – the thermal plumes of Parmentier & Turcotte (1974) or the larger scale convective movements of McKenzie, Roberts & Weiss (1974) and (iii) non-localized shear heating at the base of the lithosphere (Froidevaux & Schubert 1975). The thermal characteristics of these mechanisms are catalogued in table 1. As the measured effects are time dependent, gradational, and sensitive to parameters of the model chosen, the accuracy of the numerical values given is difficult to evaluate. However, although these mechanisms vary in origin there is a similarity in their thermal effect.

Table 1. Characteristics of sub-lithospheric thermal perturbations

mechanism	ΔT_l/°C	Δq_l/(mW m^{-2})	lateral extent km	reference
chemical plume	200–400	30–75	200–500	Anderson (1975)
thermal plume	400	30–45	200–700	Parmentier & Turcotte (1974)
convective upwelling	200–300	25–35	up to 700	McKenzie *et al.* (1974)
shear heating	100–400	2–10	not localized	Froidevaux & Schubert (1975)

The possible effects of these thermal disturbances at the base of the lithosphere are diverse: thermal expansion, doming and thinning of the lithosphere, enhanced heatflow through the lithosphere, topographic rifting at the surface and, ultimately, mid-plate volcanism. We now examine the circumstances under which one of other of these thermal phenomena could occur.

Briden & Gass (1974) observed that magmatic and metamorphic events of the past 800 Ma in Africa have been temporally associated with pauses in the apparent motion of the African plate as determined palaeomagnetically. They suggest that sublithospheric heat sources can produce crustal metamorphism and magmatism only if they are focused for long periods of time on the same part of the lithosphere, that is, only if the plate is at rest or moving very slowly. A fast moving plate would suffer only minor perturbations in passage over an asthenospheric thermal disturbance. It is clear that the terms fast and slow must be referred to the time required for a thermal disturbance to propagate upward through the lithosphere. This time is governed principally by the mode of heat transfer, either conduction or penetrative magmatism, and by the thickness of the lithosphere. We now examine, quantitatively, two conductive models that determine the thermal disturbance in the lithosphere in explicit terms of perturbation strength, plate thickness and velocity. The first model is one which, we believe, adequately represents the thermal régime beneath cratonic regions; the second represents the thermal régime characteristic of the non-cratonic regions that are sometimes host to volcanic activity.

The cratonic model envisages a convecting asthenosphere (i.e. the small scale convection or McKenzie *et al.* 1974) in which the upwellings and downwellings yield warmer and cooler regions

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over which the lithosphere passes. Thus, the base of a given segment of lithosphere, as it moves over the convecting pattern, experiences changes in temperature with time. Purely for analytical convenience, we consider this temperature change to be temporally periodic. The steady-state conductive solution (Carslaw & Jaeger 1959; Pollack & Chapman 1974) is given by:

$$\left.\begin{aligned} T(z) &= A(z)\sin\{2\pi t/\tau - \phi(z)\}; \\ \begin{Bmatrix} A(z) \\ \phi(z) \end{Bmatrix} &= \begin{Bmatrix} \mathrm{mod} \\ \arg \end{Bmatrix} \frac{\sinh kz(1+\mathrm{i})}{\sinh kl(1+\mathrm{i})}, \end{aligned}\right\} \tag{1}$$

where T is the temperature disturbance within the lithosphere, l is the thickness of the lithosphere, and τ is the temporal periodicity of the disturbance. The periodicity is, of course, determined by the distance between upwellings and the velocity of the plate. The solution is seen to be a thermal wave propagating upward from the temperature disturbance at the base of the lithosphere. The wave amplitude A diminishes and the phase shift ϕ grows with increasing distance from the base. The thermal wave is characterized by the dimensionless parameter kl, the product of a thermal wave number and the thickness of the lithosphere. The wave number is defined by $k = (\pi/\kappa\tau)^{\frac{1}{2}}$, where κ is the thermal diffusivity. Figure 4 shows the loci of the dimensionless parameter kl for a range of values of geophysical interest. Large values of the parameter correspond to thick and/or fast-moving lithosphere, while small values correspond to thin and/or slow-moving lithosphere.

Figure 5 shows the thermal wave profiles through the lithosphere for a range of the parameter kl values and demonstrates two important generalities. For $kl \leqslant 4$, there is a temperature perturbation throughout the lithosphere as well as a perturbation of heatflow at the surface. For $kl > 4$, the temperature disturbance is confined to progressively greater depths within the lithosphere, with no disturbance of heatflow at the surface. Throughout the illustrated range of kl, the temperature disturbance will lead to thermal expansion and contraction of the lithosphere and concomitant development of surficial basins and swells.

We now consider more explicit numerical examples and begin by recalling that the period τ of the thermal disturbance can be interpreted as the ratio d/u, where d is the separation between convective upwellings and u is the plate velocity. Thus the wave number can be expressed as $k = (\pi u/\kappa d)^{\frac{1}{2}}$. A value of 1500 km is adopted for the spacing between the centres of convective upwelling, following McKenzie *et al.* (1974), the thermal diffusivity is fixed at 32 $\mathrm{km^2\,Ma^{-1}}$ and the surface thermal conductivity at 3.2 $\mathrm{W\,m^{-1}\,K^{-1}}$.

Within these parameters, what are the phenomena observable at the surface, such as enhanced heatflow and uplift? The maximum change in the surface heatflow that would arise from a 100 K periodic disturbance at the base of the lithosphere, for various lithospheric thicknesses and plate velocities, is shown in figure 6. For instance, such a disturbance beneath a thin (75 km) stationary lithosphere would give a change in the surface heatflow, of the order of 4.2 $\mathrm{mW\,m^{-2}}$. However, if the plate is moving at a velocity greater than 5 $\mathrm{cm\,a^{-1}}$, the surface heatflow signal is less than 1.0 $\mathrm{mW\,m^{-2}}$ and is effectively lost. For a 150–200 km thick lithosphere, the surface heatflow signal is only marginally observable even when the plate is at rest, and is rapidly reduced by any plate motion. Taking a maximum thermal disturbance of 300 K (see table 1) and a 75 km thick lithosphere at rest, then there would be a maximum surface heatflow disturbance of 3×4.2 or 12.6 $\mathrm{mW\,m^{-2}}$. But in a cratonic situation, with lithosphere thickness of 200 km, even the largest thermal anomalies have little chance to be seen as surface heatflow disturbances.

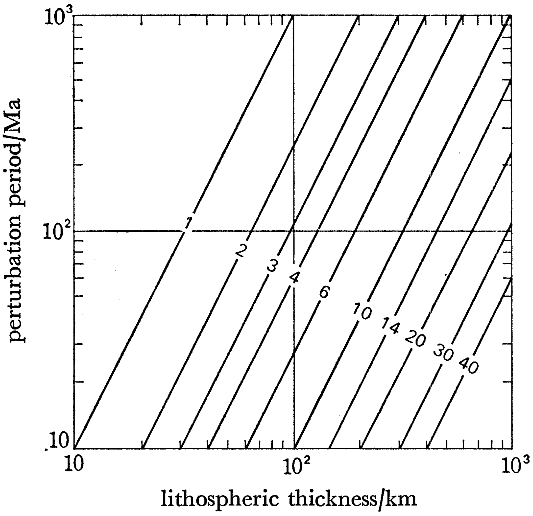

FIGURE 4

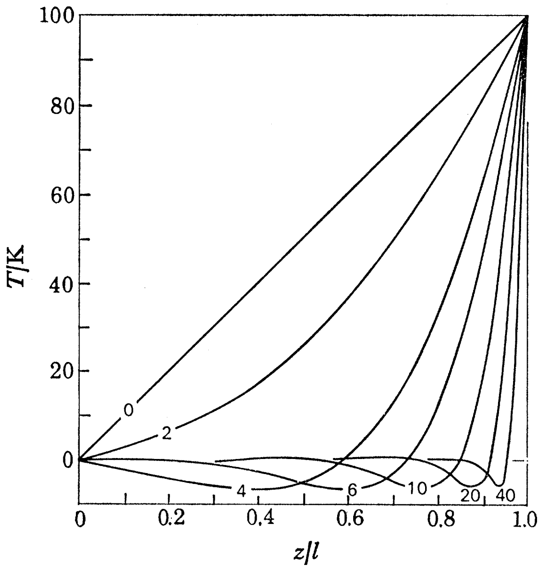

FIGURE 5

FIGURE 4. Loci of thermal wave dimensionless parameter kl (see text for definition) for range of lithospheric thickness and perturbation periods of geophysical interest.

FIGURE 5. Thermal wave profiles through lithosphere of thickness l; amplitude of thermal disturbance is 100 K at base of lithosphere. Numbers on profiles indicate dimensionless parameter kl.

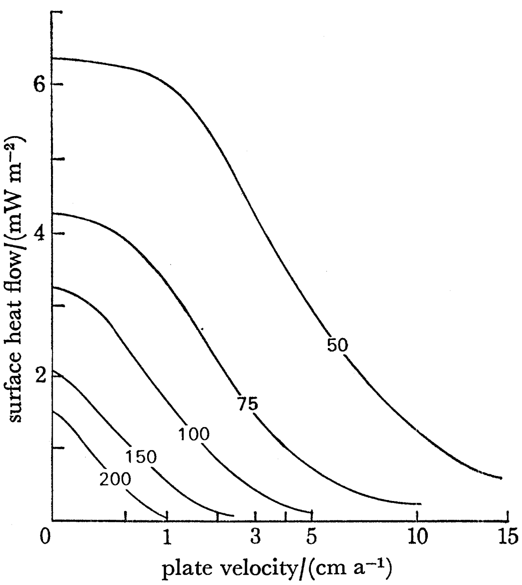

FIGURE 6

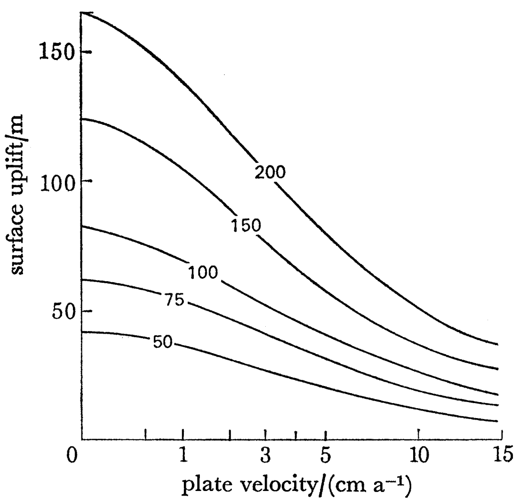

FIGURE 7

FIGURE 6. Perturbation to heatflow at surface of moving lithosphere, due to periodic temperature disturbance of 100 K amplitude at base, arising from passage at indicated velocities over thermal anomalies separated by 1500 km. Numbers on curves are lithospheric thickness in kilometres.

FIGURE 7. Surface uplift of moving lithosphere arising from thermal expansion due to conditions described in figure 6. Numbers on curves are lithospheric thickness in kilometres.

The surface uplift that would result from thermal expansion due to a 100 K periodic disturbance at the base of the lithosphere is calculated as

$$(3\lambda+2\mu)/\{3(\lambda+2\mu)\}\int_0^l \alpha T(z)\,\mathrm{d}z,$$

where λ and μ are elastic constants assumed equal, and α is the coefficient of volume thermal expansion. The complementary uplift associated with phase changes involved in the transition from garnet through spinel to plagioclase peridotite is not considered here, although it may be of comparable magnitude to the thermal expansion. Figure 7 shows the uplift due to thermal expansion. Again one observes the greatest effects in a stationary plate. For example, where a change of 300 K occurs at the base of a stationary, 200 km thick lithosphere with an expansion coefficient of $3\times10^{-5}\,\mathrm{K}^{-1}$, 500 m of uplift or subsidence would result, a magnitude equal to about half the basin and swell topography of cratonic Africa (Faure 1971, 1973). At fairly rapid plate velocities (*ca.* 5 cm a^{-1}) the uplift and subsidence which could be produced on a lithosphere of the same thickness would be reduced by a factor of about two. At this velocity, while the uplift is still apparent, the surface heatflow signal would be too small to be observed. The expansion of stationary or slowly moving lithosphere, alone, is somewhat inadequate to account for the full magnitude of basin and swell topography. However, it is clear that for a thermal perturbation to exist at the base of the lithosphere, a substantial section of the underlying mantle must also exhibit elevated temperatures and be expanded. The additional uplift from this sublithospheric expansion, along with effects of mineralogical phase changes, are more than adequate to make up the deficiency in uplift from lithospheric expansion alone.

What processes will occur at the base of the lithosphere as a consequence of a thermal perturbation? In attempting to answer this question quantitatively, we first define the base of the lithosphere, as do many others, as being the mantle solidus, i.e. the intersection of the geotherm with the periodotite incipient melting point curve. Also, though the solidus will shift according to the volatile content and composition, the variability does not affect our general conclusions. The principal effect of a temperature change at the base of the lithosphere will be the movement of that boundary itself, and an upward or downward movement can be regarded as a thinning or a thickening of the lithosphere.

First we examine a cratonic situation with a geotherm yielding a surface heatflow of 45 mW m^{-2} and grazing the mantle solidus, thereby defining the base of the lithosphere, at 200 km (figure 8), and developing a weak, underlying low-velocity zone. We then describe the perturbation of this boundary by a thermal excess of 300 K, by the use of the thermal wave solution (equation 1 and figure 5) for the parameter $kl = 6$, which, for a 200 km lithosphere, corresponds to a plate velocity of *ca.* 1.4 cm a^{-1}. As can be seen from figures 5 and 6, for this combination of plate thickness and velocity, the thermal wave damps out within the lithosphere and yields no heatflow signal at the surface, and the thermal disturbance is confined to the lower quarter of the lithosphere. When this perturbation of 300 K is added to the cratonic geotherm at the base of the lithosphere, it produces a perturbed geotherm (figure 8) which is steeper in the 150–200 km depth range, and intersects the solidus at about 160 km, thinning the lithosphere by some 40 km from the original thickness. The temperature gradient in the perturbed zone below 150 km is elevated to 9 K km^{-1} and produces a 'kink' in the geotherm in the same depth range and of the same magnitude as the controversial pyroxene geotherm of Boyd (1973). A cooling perturbation would lower the geotherm below the solidus everywhere, leading to substantial thickening of the lithosphere.

So, topographic variations of *ca.* 150 m, little or no surface heatflow perturbation, and lithospheric thinning of some 40 km over warmer regions or substantial thickening over cooler regions, are characteristics of a stationary, or slowly moving, 200 km lithosphere with a 300 K thermal disturbance at its base. These characteristics, we believe, typify the African cratonic lithosphere.

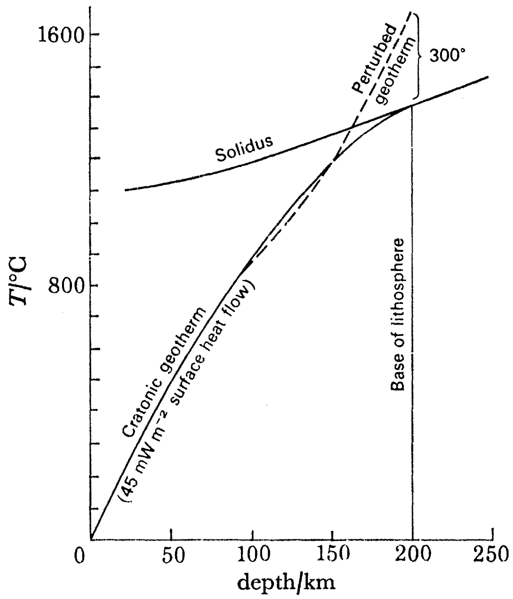

FIGURE 8. Perturbation of cratonic geotherm and thinning of lithosphere due to 300 K periodic temperature disturbance at base. Plate velocity = 1.4 cm a^{-1}. Only perturbation due to warming is shown; cooling perturbation not illustrated.

We choose a second analytical model to examine non-cratonic areas. We take a point source of heat, fixed at depth beneath a moving lithosphere (Carslaw & Jaeger 1959; Birch 1975). The resulting temperatures are given by

$$T(x,y,z) = \frac{q}{4\pi K}\left[\frac{\exp\{-(R_1-x)\,U/2\kappa\}}{R_1} - \frac{\exp\{-(R_2-x)\,U/2\kappa\}}{R_2}\right], \tag{2}$$

where q is the strength of the source, R_1 and R_2 are the distances from the field point to the source and an image source, K is the thermal conductivity and U is the velocity in the x direction. This model is intended to represent a 'hot spot': a thermal or chemical plume. One should note however, that the point source may be replaced by any temperature boundary condition that is compatible with the temperature distribution derived from the point source. Thus, the model could also represent a broad 'thermal cushion' as well as a confined point source; the resulting temperatures above the cushion will be identical to those developed from a point source. We set the strength of heat source to be such that if it were located at 50 km depth, it would yield a 45 mW m^{-2} increment to the surface heatflow directly above, i.e. it is strong enough to double the cratonic heat flow if it were emplaced to a shallow depth.

Figure 9 shows the thinning of a cratonic lithosphere by a point thermal disturbance at its base, for various plate velocities. The thinning is symmetrical and maximum for a stationary plate, but for even modest velocities the perturbation is lessened, and distorted in the direction of plate motion. A velocity of only 1 cm a^{-1} halves the thermal penetration, allowing the lithosphere to remain close to its original thickness. Significantly, even for greater source strengths, slow motion of the plate is an effective heat distributor, prohibiting extensive thermal penetration and lithosphere attenuation.

One observes that for the given source strength, the 200 km lithosphere is thinned only to about 130 km, far short of the thinning to 50 km indicated by teleseismic studies of the volcanic areas of Ethiopia and east Africa (Long, 1976; Makris, Menzel, Zimmerman & Gouin 1975). The requisite thinning to 50 km can be achieved by a source at 200 km, but this would have to be an order of magnitude greater than the source strength illustrated in figure 9. Such an augmented lithothermal system is shown in figure 10*a*; its effects range horizontally over about 1000 km, beyond which the lithosphere retains a cratonic thickness. Directly above the source, the lithosphere is only 50 km thick; the thinning is accompanied by a 35 $mW\,m^{-2}$ increment to the surface heatflow.

An alternative model for thinning, which requires little or no augmentation of the source strength but which does allow the source to be upwardly mobile, can also be constructed. We consider an initial perturbation at the base of the lithosphere which leads to a thinning of the lithosphere. If, as the lithosphere thins, the source follows the base of the lithosphere upward, then stages of the progressive thinning would develop approximately as in figure 10*b*. As the source moves upward, the position of the solidus about it is also elevated, but the region involved in the partial melting becomes progressively less as the source is located at progressively shallower depth. This contraction of the supersolidus region occurs because of the greater difference between ambient temperatures and the solidus at shallow depths, compared with the situation at 200 km where the ambient temperature is at the solidus. Thus, at shallow depths only a small region about the source receives enough heat to reach partial melting; near the base of the lithosphere, small temperature perturbations at a considerable distance from the source are sufficient to bring the ambient temperature up to the solidus. The thinning in figure 10*b* is accomplished with a smaller heat source than that of figure 10*a*, but because the source ascends to 50 km, it yields a surface heatflow increment of 45 $mW\,m^{-2}$.

The models depicted in figure 10 offer sufficient flexibility, through adjustment of source strength and degree of penetrative magmatism, to embrace the full range of observed or deduced lithospheric thinning, horizontal dimensions, and heatflow anomalies of the east African lithothermal systems. In the discussion to follow, we compare these models to others deduced from other geophysical and petrological observations in the Afar region. As with the first model, the effect of plate motion on this point-source model is to attenuate the perturbation and confine it to greater depths.

Both analytical models discussed are steady-state solutions and it is proper to enquire whether the times required to achieve a significant fraction of the steady state are compatible with the time scale of African Cainozoic lithothermal systems, with characteristic life times of 10–30 Ma (§ 2) and with the cratonic thermal régimes where surface uplift–subsidence cycles have typical periods of 10–30 Ma (Faure 1971). For conductive thermal disturbances confined to the lower 50 km of a 200 km cratonic lithosphere, such as that illustrated in figure 8, the bulk of the perturbation is achieved in about 7 Ma; at lesser velocities (*ca.* 0.3 $cm\,a^{-1}$, cf. Faure 1973), the perturbation penetrates upwards through about $\frac{2}{3}$ of the lithosphere and is well established in 50 Ma. Thus conductive times are compatible with observed basin and swell evolution. The smaller times characteristic of lithothermal system evolution in non-cratonic regions suggests that penetrative magmatism, which proceeds on a time scale one or two orders of magnitude greater than heat conduction, is significant (Marsh 1978, this volume, pp. 611–625).

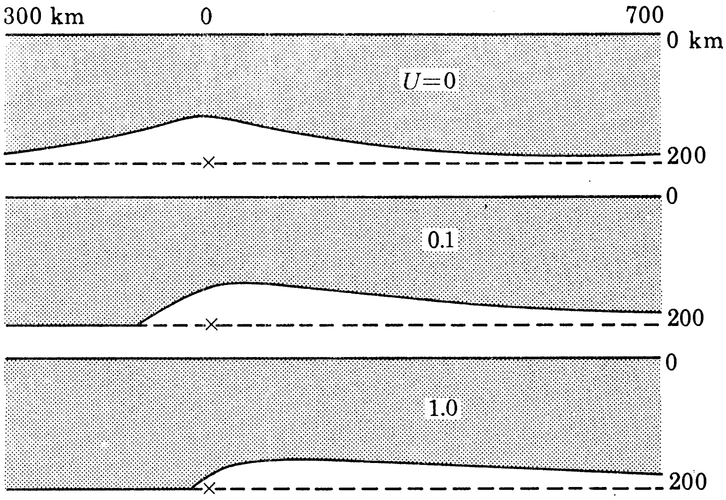

FIGURE 9. Thinning of 200 km lithosphere moving to the right at indicated velocities (cm a^{-1}), by a heat source located at base in position indicated by a cross. Final configuration of lithosphere, depicted by shaded region, is asymmetrical.

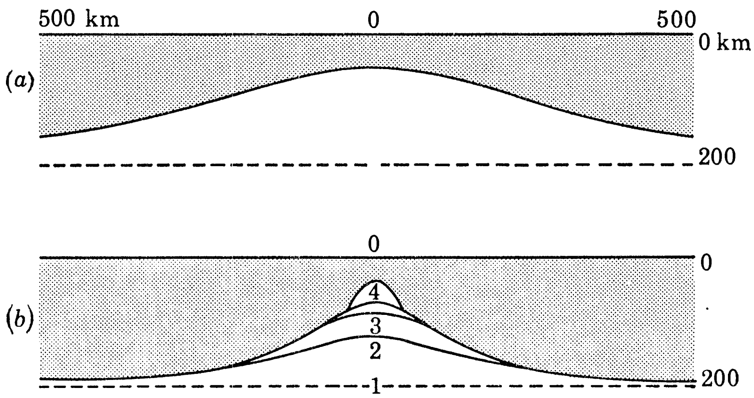

FIGURE 10. (*a*) Thinning of 200 km stationary lithosphere by a strong heat source at base. (*b*) Thinning of 200 km stationary lithosphere by an ascending heat source sequentially located at sites 1–4. Shaded regions depict final configuration of lithosphere.

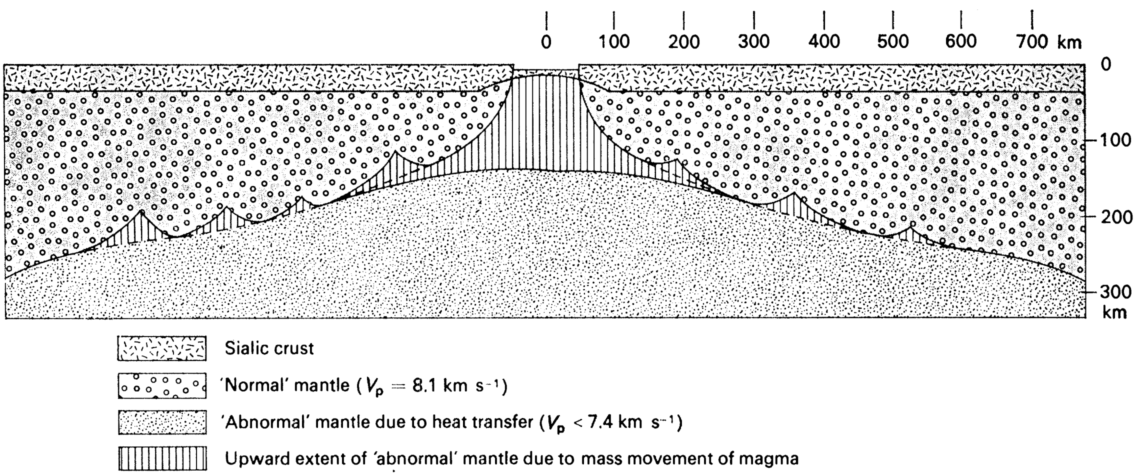

FIGURE 11. East Africal lithothermal system derived from teleseismic evidence (Long 1976) and geochemical data (Gass 1970, 1975).

5. Discussion

We now examine the composition and the spatial and temporal distribution of African Cainozoic volcanic rocks and enquire whether they can be realistically explained within the geological and geophysical parameters presented, and in the light of the various models given in § 4. For this evaluation we have used three variables: (i) sublithospheric temperature and heat flux, (ii) lithosphere thickness, and (iii) plate velocity. Models based upon these parameters account for many major features of Cainozoic volcanism and tectonics. These include the virtual restriction of volcanism to non-cratonic areas (see figure 1). They also explain the occurrence of uplifted areas, either with anomalously high heatflow but no volcanism (e.g. Zambia, Chapman & Pollack 1975*b*), or without either heatflow anomalies or volcanism. The models proposed to account for lithospheric thinning (p. 591) adequately depict the form and character of the East African–Ethiopian lithothermal systems. We discuss these systems in more detail below.

First we consider anomalous situations: volcanism within cratons and/or sedimentary basins. While the virtual absence of Cainozoic volcanic products in cratonic areas is readily explained by the models, the quantitatively insignificant and petrologically unusual products of Cainozoic volcanism in cratonic areas, around the Western Rift, are inexplicable in terms of penetrative magmatism of the type depicted in figure 10*b*. It is relevant that African kimberlites are currently regarded as having been formed at depths of *ca.* 200 km and to have 'drilled' their way through cratonic lithosphere by processes of gas-fluidization (Harris, Kennedy & Scarfe 1970). It seem likely that the 'kimberlitic' volcanics of the intra-cratonic Igwisi Hills (figure 1) had a similar mode of emplacement (Reid *et al.* 1975; Kennedy 1965; Dawson 1971) while the potassic Cainozoic volcanics of the Western Rift may also have been generated at great depth (Holmes & Harwood 1932).

A second anomaly is the occurrence of volcanics in the North African Cainozoic sedimentary basins (§ 2). As these volcanics are not associated with uplift, they are also inexplicable in terms of lithothermal systems governed by the parameters we have discussed.

Considering the non-cratonic areas, the simplest case of non-cratonic volcanism is that of an isolated volcanic field embellishing a domal uplift in the underlying crystalline basement (e.g. Bayuda, Jebel Marra, Tibesti). Here a thermal event of the type depicted in figure 10*b*, with a time scale of 0–20 Ma, could well produce the observed structural and volcanic phenomena. We suggest that the major control in the horizontal and vertical dimension of the surface uplift and the quantity of the volcanic products is the regional extent and magnitude of the causative thermal anomaly, although mantle heterogenity on a similar scale cannot be precluded.

We now consider East Africa, where there are four areas of topographic uplift, each having volcanic products of distinctive composition and abundance and all cut by major rifts which collectively form the East African rift system. It has been proposed (Gass 1970, 1972; Long 1976) that the primary mechanism was the production of overlapping thermal (lithothermal) systems in the mantle and that the uplift, rifting and magmatism were consequent thereon. In the context of this discussion, in the East African systems the magma has moved preferentially upwards along the rift zones; distortion upward of the boundary of the anomalous mantle follows these trends. Beneath the rift zones the sialic crust has been thinned, often to 5–10 km (Fairhead 1976; Makris *et al.* 1975), but little or no crustal separation has taken place. The extreme case is that of the Red Sea and Gulf of Aden where tectonic and magmatic processes have culminated in the separation of the continental crust and the production of new constructive margins.

Just why new constructive margins formed along the line of the Red Sea and the Gulf of Aden, and not along the East African rift system, has been the subject of much speculation. We shall not add greatly to this speculation, but note that if extent of uplift and the volume of volcanic products are related to the size and vigour of the underlying thermal system, then there is no doubt that the Afro-Arabian dome represents by far the largest African Cainozoic lithothermal system. The vertical uplift will give gravitational instability, and this lithospheric stress will be augmented by that related to the nearby spreading in the Indian Ocean and subduction between Afro-Arabia and Eurasia.

Finally, and briefly, we note that the other major episode of African Phanerozoic magmatism, associated with the disruption of Gondwanaland, also occurred at a time when the African apparent polar wander path slowed or paused, a situation which, we contend, promotes the development of lithothermal systems. However, in contrast to the Cainozoic lithothermal systems in Africa, those of the Mesozoic occur in both cratonic and non-cratonic terrains, and suggest that the Mesozoic thermal anomalies were of a magnitude sufficiently great to thin, weaken and fracture thick cratonic lithosphere. We have already suggested above that the dimensions of thermal doming and the volume of volcanic products are related to the vigour of the lithothermal system, and that only the largest of the systems, such as the Afro-Arabian dome, evolve into constructive plate margins. In this context, it seems unsurprising that the outpouring of the Gondwana volcanics, volumetrically so immense, heralded the opening of the modern southern oceans.

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Discussion

C. Froidevaux (*Laboratoire de Physique des Solides, Université Paris Sud, 91405 Orsay, France*). I should like to raise two questions. First, Professor Gass seems to favour a hot spot model for intra-plate African volcanism. But is not the assumed thermal anomaly under the plate reduced to a mathematical point, i.e. without lateral extension? This may explain why its effect disappears as soon as the plate is set in motion. Also, is not the suggestion that the plate has to be 'at rest' for volcanism to occur, in contradiction with the existence of Hawaii and other conspicuous volcanoes on the fast moving Pacific plate? Secondly, what is Professor Gass's reaction to the suggestion made in my paper (Froidevaux & Souriau, this volume, pp. 387–392) that extended intra-plate volcanism occurs *after* a period of rapid plate motion, i.e. after a time constant of the order of 100 Ma, as suggested by the data in the preceding paper?

I. G. Gass. With regard to the first question, the use of a point source in our models was for analytical convenience only; a distributed source can of course be constructed by integration. However, it is not the geometry of the source that gives rise to the rapid dissipation; rather it is the convective flux in the direction of the plate motion that distributes the heat so effectively.

We have argued that plates are most vulnerable to penetrative magmatism when they are thin and are at rest or moving only slowly with respect to an underlying thermal anomaly. Certainly the Pacific plate over much of its extent is thinner than the lithosphere in Africa, and thus may be vulnerable even though moving at a higher velocity. Of course we do not exclude the possibility that other mechanisms, for example that proposed by Turcotte & Oxburgh (this volume, pp. 561–579) may contribute to the Hawaiian volcanism.

Concerning the second question, we think that 100 Ma is a representative time constant for conduction through a 150 km lithosphere. However, penetrative magmatism culminating in volcanism proceeds much faster than does conduction. Accordingly, we associate its initiation with the coming to rest of the plate shortly before the appearance of volcanism, rather than with a period of rapid plate motion long before the volcanism.