NEW EVIDENCE FOR A PLUME BENEATH EAST AFRICA

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Although East Africa has long been regarded as a classic area to study the early stages of continental break-up, the origin of the Cenozoic rifting, volcanism, and plateau uplift found there is still poorly understood. The East African rifts can be linked to extensional stresses associated with the development of the Afar triple junction (Baker *et al.*, 1972), but it is unlikely that the small amount of crustal extension (≤ 10-15 km) across East Africa could have led to sufficient pressure-release melting to account for the large volume of extrusive rock present in several parts of the rift system, or to enough thinning of the lithosphere to cause significant (≥1 km) plateau uplift (McKenzie and Bickle, 1988; Ebinger, 1989). Consequently, a purely "passive" origin for the Cenozoic tectonism has been frequently rejected, and, instead the tectonism has been commonly attributed to a mantle plume, or to "passive" rifting above a mantle plume (e.g., Green *et al.*, 1991; Latin *et al.*, 1993; Macdonald, 1994; Slack *et al.*, 1994; Achauer, 1994; Prodehl *et al.*, 1994; Paslick *et al.*, 1995; Burke, 1996; Fuchs *et al.*, 1997). However, geochemical and geophysical data from East Africa provide few constraints on mantle composition and structure beneath the lithosphere where the plume is hypothesized to lie, and thus whether or not a plume exists beneath East Africa is debatable.

Results from two recent studies of broadband seismic data, when combined, shed new light on the East African plume debate. The seismic data come from teleseismic earthquakes recorded on twenty stations deployed across Tanzania between June 1994 and May 1995 (Nyblade *et al.*, 1996). In the first study, travel times from P and S waves were inverted for upper mantle seismic velocity variations (Ritsema *et al.*, 1998). The patterns of P and S wave velocity variation obtained are similar, and so we limit our discussion to the S velocity model. The model shows higher-than-average velocities beneath the Tanzania Craton and predominantly lower-than-average velocities beneath the rifted mobile belts surrounding the craton. The low velocity region under the Eastern rift extends vertically to a depth of at least 400 km and laterally over a region about 300 km wide. The lithospheric keel beneath the craton, as defined by the relatively fast velocities, extends to a depth of ≥ 200 km. Between depths of 200-300 km the low velocity structure associated with the rifts begins to creep under the fast structure of the cratonic lithosphere.

There are two likely causes for the contrast in the upper mantle S wave velocities between the craton and Eastern rift. High S wave velocities (~3-6%) commonly characterize cratonic lithosphere to depths of at least 200 km in other Precambrian shields (e.g., Grand, 1994; Grand *et al.*, 1997; Van der Lee and Nolet, 1997; Ekstrom and Dziewonski, 1998) and are expected for the cold, chemically depleted structure of the cratonic keel (Jordan, 1979; 1988). It is possible that the entire velocity contrast observed (~5-6%) could result from the fast cratonic structure alone. However, the contrast in velocities could also result, in part, from temperature anomalies in the mantle beneath the Eastern rift. Since both effects can create velocity anomalies of several percent, it is not easy to partition the velocity variation between the fast cratonic keel, on the one hand, and thermally perturbed rift structure, on the other. Additional constraints on the thermal and/or compositional structure of the upper mantle are needed to separate these effects.

A study of the topography on the 410 and 660 km discontinuities beneath Tanzania provides the required constraints (Owens *et al.*, 2000). The 410 and 660 km discontinuities are generally interpreted as mineral phase transitions in olivine from α-spinel to β-spinel, and for γ-spinel to perovskite+magnesiowustite, respectively (Bina and Helffrich, 1994). The Clapeyron slopes of the equilibrium phase boundaries indicate that the depth of the 410 km discontinuity should be deflected downwards in regions of warmer temperature and the 660 km discontinuity should be deflected upwards (Bina and Helffrich, 1994). Therefore, topography on the discontinuities can provide information about upper mantle thermal structure in the 400-700 km depth range.

Topography on the 410 and 660 km discontinuities beneath Tanzania has been estimated by geographically stacking receiver functions (Owens *et al.*, 2000). Results show that the 410 km discontinuity is depressed by 20-30 km over an area 200-400 km wide beneath the Eastern rift. This depression corresponds to a temperature anomaly of about 200-300 K (Owens *et al.*, 2000) and coincides directly with the low velocity anomaly seen beneath the Eastern rift.

The coincidence of the depressed 410 km discontinuity and the low velocity region beneath the Eastern rift indicates that at least some of the S wave velocity variation between the craton and Eastern rift is due to temperatures beneath the Eastern rift elevated by 200-300 K. Using laboratory measurements of the temperature derivatives of wave speeds in olivine (Isaak, 1992), a 200-300 K temperature increase in the upper mantle would reduce S wave velocities by ~2%. In comparison to the 410 km discontinuity, little relief is observed on the 660 km discontinuity.

These findings are important for understanding the origin of Cenozoic tectonism in East Africa. The depth extent of the upper mantle thermal disturbance beneath the Eastern rift is not easily explained by small-scale convective upwellings induced by "passive" stretching of the lithosphere. In passive rift models, sublithospheric mantle flow fills in "voids" created by the stretched lithosphere. This leads to small-scale convective instabilities near the base of the lithosphere but not likely throughout the upper mantle (Buck, 1986; Mutter *et al.*, 1988). Additionally, in fast-spreading oceanic rifts, anomalously slow mantle structure caused by upwelling and decompression melting extends to depths of only ~200 km (Toomey *et al.*, 1998; Webb and Forsyth, 1998). Thus, in northern Tanzania where the total amount of lithospheric extension during the past 5-8 Ma has been only 10 km over an area 300 km wide (Foster *et al.*, 1997; Ebinger *et al.*, 1997), it seems highly improbable that small-scale convective upwellings could extend from the transition zone all the way to the base of the lithosphere.

Tomographic images of the mantle beneath Africa show broad S velocity anomalies in the upper mantle beneath East Africa that may possibly connect with low velocity structure in the lower mantle beneath southern Africa (Grand *et al.*, 1997). If the low velocity structure is continuous across the 660 km discontinuity, then it would appear that there might be a broad thermal upwelling extending from the core-mantle boundary all the way to the uppermost mantle. While the S wave model and the depression of the 410 km discontinuity are consistent with a broad thermal upwelling extending from the lower mantle into the upper mantle, the flat 660 km discontinuity under Tanzania is not easily explained by a thermal upwelling that is continuous across the transition zone, at least not by one beneath Tanzania.

In contrast, the depth extent of thermally perturbed structure beneath the Eastern rift, the 200-400 km wide depression of the 410 km discontinuity, and the flat 660 km discontinuity can be explained by a plume model, if a plume head lies beneath the eastern margin of the Tanzania Craton. In this interpretation, the thermal structure beneath the Eastern rift is caused by buoyant (warm) plume material that has migrated around and up the eastern side of the cratonic keel modifying the mantle lithosphere beneath the Eastern rift. Plume head temperatures are estimated to be 100-300 K above ambient mantle temperatures (McKenzie and Bickel, 1988, Campbell and Griffiths, 1990; Farnetani and Richards, 1994), sufficient to reduce S wave velocities by a few percent.

The 200-400 km wide depression of the 410 km discontinuity is explained in this model by the plume head lying across the 410 km discontinuity. Fluid dynamic studies of plumes suggest that plume heads can be several hundred kilometers across and between 150-200 km thick (Grifftiths and Campbell, 1991). Hence, if a plume head impinged on thick ($\geq$200 km) craton lithosphere, it is possible that the bottom of the plume head could extend to depths of $\geq$~400 km, giving rise to a depression of the 410 km discontinuity that is several hundred kilometers across while not affecting the 660 km discontinuity. A plume head situated beneath the eastern side of the craton could also explain why there is little volcanism in the Western rift compared to the Eastern rift. For a plume located under the eastern margin of the craton, most of the plume material would flow around the craton lithosphere to the east and a lesser amount would make its way around the craton lithosphere to the west.

Our plume interpretation is supported by a recent study of mantle xenoliths from Labait Volcano, which lies on the eastern edge of the Tanzania Craton in northern Tanzania (Chesley *et al.*, 1999). In this study, the xenolith from the deepest depth (~160 km) exhibits a major element composition similar to estimates of primitive mantle. In addition, its $^{187}Os/^{188}Os$ ratio is radiogenic, overlapping with the range seen for metasomatic xenoliths and the host melilitite. Shallower samples, in comparison, have refractory compositions typical of old cratonic lithosphere. These observations suggest that the sublithospheric mantle under Labait Volcano has geochemical characteristics of a plume and is distinct from MORB-source mantle.

As mentioned previously, a number of plume models have been proposed for East Africa, and our seismic images can be used to evaluate them. Ebinger and Sleep (1998) recently argued for a model in which one large plume head impinged on the lithosphere in southern Ethiopia, and the plume material flowed outward from this location channeled by topography on the lithosphere-asthenosphere boundary. Considering that material from the hypothesized plume must flow outward in all directions from southern Ethiopia to explain the widespread occurrence of Cenozoic volcanism in Africa, it is difficult to explain with the Ebinger and Sleep model the 200-400 km wide thermal anomaly throughout most of the upper mantle beneath northern Tanzania. In order to do so, it would require most, if not all, of the plume material to have flowed from Ethiopia south along the eastern margin of the Tanzania Craton.

Many authors have suggested that a plume exists beneath the Kenya rift based on geological, geochemical, and geophysical observations. A plume head centered beneath the Kenya rift is consistent with, although not required by, our findings. Even though our study area does not extend far into Kenya, our seismic results suggest lower mantle velocities and greater depression of the 410 km discontinuity beneath Kenya than beneath northern Tanzania. The fact that we observe little relief on a deep 660 km discontinuity beneath Tanzania is also consistent with a plume centered beneath Kenya, since the tail from a plume beneath Kenya would affect the 660 km discontinuity outside of the region that we image.

In summary, seismic velocity models of the upper mantle combined with topography on the 410 km discontinuity provide evidence for a deep thermal anomaly beneath the Eastern rift that is consistent with the presence of a mantle plume lying under the eastern margin of the Tanzania Craton. Cenozoic uplift and volcanism in East Africa can be readily attributed to this plume. The warm plume material would lead to isostatic uplift of the lithosphere, and convection within the plume head could contribute a dynamic component to that uplift, as indicated by the gravity field (Ebinger *et al.*, 1989). The volcanism may result from decompression melting of warm plume material at shallow mantle depths under the rifts and by melting of heated mantle lithosphere above the plume.

The extent to which the rifting is influenced by the plume, however, remains uncertain. The timing between plateau uplift and rifting is needed to unravel the relative contributions to the extensional stress field from the plume (i.e., membrane stresses due to lithosphere uplift) versus far-field plate boundary forces associated with the opening of the Afar triple junction. The history of plateau uplift in East Africa relative to the timing of rifting is poorly constrained, and thus it is difficult to determine how much influence the plume may have had on rifting. Constraints on the timing of uplift are needed to address further the cause of rifting, constraints which may possibly be obtained in the future from fission track, cosmogenic isotope or possibly other types of data that are sensitive to lithospheric uplift.

Upper mantle structure beneath the East African rift reported in this paper is significantly different than that found beneath other major continental rift systems. Low velocity structure in the upper mantle is found beneath the Rio Grande and Baikal rifts, for example, but there is little indication from existing studies that the anomalous structure extends to depths of several hundred kilometers as found under East Africa. In fact, there is a growing body of evidence from the Baikal rift suggesting that the upper mantle has not been significantly perturbed away from the rift axis. The difference in the depth extent of thermally modified upper mantle beneath these rifts reflects their different origins. The origin of the Baikal rift is clearly associated with far-field plate stresses caused by the collision of the Indian and Eurasian plates, while the origin of the Rio Grande rift is linked to the development of the Basin and Range province in the western U.S.

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