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THE AFRICAN PLATE

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The entire African Plate in both its continental and oceanic areas has evolved in distinctive and unusual ways over the past roughly thirty million years. I attribute these peculiarities to the African Plate having come to rest with respect to at least part of the convective circulation in the underlying mantle. On the grandest scale interaction has produced a plate-wide structure of basins and swells with wavelengths ranging from 200 to 2,000 km and a common above average elevation of both the land surface and the ocean floor. Denudation of the swell areas on land has generated escarpments and extensive erosion surfaces and has resulted in rapid progradation of the deltas of those of the continent's great rivers which reach the sea. During a preceding interval lasting from at least 65 to about 30 Ma, Africa was a low-lying continent with widespread deep weathering. Igneous activity, apart from that generated by the Tristan and Deccan plumes, was continuous only in an area in Cameroon. Sporadic volcanism elsewhere was generally short lived with the exception of a ten million year episode of basaltic eruption between 45 and 35 Ma close to the future site of the Southern Ethiopian Rift near the Kenya-Ethiopia border. Between 65 and 30 Ma some of the intracontinental rifts that had formed at about the beginning of Cretaceous time (~140 Ma), for example: the Anza, Benue, and Sirte Rifts, were still receiving substantial volumes of sediment although rift igneous activity had stopped. The earlier Cenozoic from ~65 Ma to ~30 Ma thus appears to have been a relatively quiet interval in the African Plate's igneous, topographic and tectonic evolution.

A simultaneous outbreak of volcanic activity in many separate regions of the African Plate about 30 million years ago signalled the end of the quiet time. Episodic volcanism has marked some of these areas ever since. The concentrations of volcanism are interpreted as indicating the sites of impact on the base of the lithosphere of rising hot material or mantle plumes. These plumes, from their surface manifestations, appear to be of a great variety of sizes. Those beneath the Afar, the Canaries and the Cape Verde islands appear to be the largest. The persistence in the same places of the eruption of volcanic rock from forty or more discrete mantle plumes is the most direct evidence that the plate has stopped moving with respect to at least part of the underlying mantle circulation. Volcanism at the hot spots that mark the sites of underlying plumes is typically distributed over elliptical areas with semi-radii of up to 250 km. Horizontal propagation of volcanic activity at rates of 10 mm/yr or slower has been reported from within some of these elliptical areas, but a plate-wide consistent azimuth of hot spot propagation cannot be discerned.

Although intraplate volcanism is now widespread on the African Plate, only the Tristan-Gough island and seamount complex and Réunion island among active volcanoes lie near the ends of volcanic hot spot tracks. At Tristan, intermittent volcanism has been going on in the same elliptical area for about 30 million years. The Réunion hot spot, although close to the projected prolongation of the hot spot track from the Deccan traps, is unrelated to the Deccan source plume. That plume died at the time of the approximately 30 Ma ridge jump which formed the Central Indian and Carlsberg Ridges.

Plume-generated, intraplate or hot spot volcanism is widely, although patchily, distributed over the oceanic parts of the African Plate. The volcanic rocks generally have a HIMU geochemistry indicating that they represent only a small degree of partial melting of the mantle and perhaps also indicating derivation from source material above the 670 km discontinuity. On the African continent and in Madagascar, post 30 Ma volcanism shows some concentration in elevated areas. This volcanic activity has been almost entirely restricted to areas underlain by crust that had been reactivated during Pan-African times (roughly 900 to 400 Ma). The restriction has been attributed to the presence of fertile mantle lithosphere under those reactivated areas. Fertile material was emplaced under the continental crust following lithospheric delamination during Pan-African continental collisions. Interaction with the underlying circulating mantle over the past 30 My has extracted magma from the fertile areas. In the cratonic areas of the African continent, beneath which there is only sporadically distributed and locally-metasomatized fertile mantle lithosphere, there are swells but no active volcanism. It has also been suggested that the restricted distribution of post 30 Ma volcanism is simply because the cratonic lithosphere is thicker or stronger than the lithosphere reactivated in Pan-African times and has mechanically resisted penetration by rising plumes.

There have been several episodes of rifting in Africa over the past 270 million years. A major episode began ~140 Ma near the beginning of Cretaceous times. Several of the rifts formed at that time continued to receive sediments until ~30 Ma but their rift-associated igneous activity had long ended. No new rifts formed in Africa between ~140 and ~30 Ma when the current East African rifting episode began. I regard the current episode of rift activity in East

Africa, which began at the same time as the plate-wide upsurge in intraplate volcanism and the formation of basin and swell structures all over the African Plate, as related phenomena. Recognition of the coherent behavior of the entire African Plate and of the development of the East African Rift System as part of that coherent behavior helps in understanding the origin and evolution of the East African Rift System. Volcanism in and around the active rift system is in some places concentrated in areas such as the Afar, Turkana and at Samburu beneath the Gregory Rift that may mark the sites of discrete plumes. Those plumes are part of the 30 Ma and younger population of plumes which is sporadically distributed under the entire African Plate. Other volcanism in the East African Rift System is attributable to pressure reduction in areas that have been extended by the rifting process.

The petrology of the igneous rocks associated with the East African Rift System has been the subject of exhaustive research throughout the past 100 years. Because volcanism of hot spot volcanoes on the oceanic parts of the African Plate has been dominated by the HIMU source over the past 30 My I suggest that the volcanism of the contemporary hot spots and rift areas on the African continent should also be considered as primarily controlled by a HIMU source. This simple idea appears to work quite well. Other isotopically distinct reservoirs identified as EM1 and EM2 are reported to have played an important role in African continental volcanism over the past 30 My. I suggest that this material has come from interaction between HIMU magmas generated from sub-continental plumes and the fertile mantle lithosphere that underlies the Pan-African reactivated areas of the continent. The great petrologic variety of the East African Rift System can be interpreted as largely related to episodes of emplacement within the crust and mantle lithosphere of igneous rocks from HIMU, EM1 and EM2 sources at various times during the past 30 My. Emplacement in the lithosphere may have been followed by partial remelting, fractionation and eruption related to heating by later or continuing plume activity or, alternatively, to pressure release melting under actively extending rifts. The abundance of post 30 Ma carbonatites, which are known from both continental and oceanic areas of the plate, can also be interpreted as related to a HIMU dominated source. The absence of post 30 Ma kimberlites in contrast to the abundance of carbonatites may indicate generally shallower rather than deeper separation of immiscible liquids. The Tristan and Gough island and seamount province, which is distinct from all other African Plate volcanic provinces, represents a plume that impinged on the base of the lithosphere 130 Ma and may be the world's oldest active plume. Tristan erupts through 30 My old oceanic lithosphere and, in contrast to products of the much younger plume population, is dominantly an EM1 and EM2 derived object. In spite of its great age and eruption through fairly old lithosphere, Tristan still has roughly the same ^{3}He signature as typical MORB. The Afar overlies a very large plume that has erupted a great mass of basalt beginning at 30 Ma. Afar has a high ^{3}He signal, yields rocks suggestive of HIMU, EM1 and EM2 derivation and erupts through continental as well as transitional to oceanic lithosphere. Réunion shows some isotopically distinct features and a high ^{3}He signal.

Earthquakes and active faulting are concentrated in the general area of the East African Rift System and its surrounding swells but there have been sporadic earthquakes with magnitudes as large as 7 in other parts of the plate interior. Active tectonism in the form of faulting has been reported from many areas of Africa. Older faults have been reactivated in many parts of the plate within the past 30 My. These faults have commonly behaved as strike-slip or thrust faults perhaps because of the general dominance of the ridge push force on the African Plate. By contrast, areas of active volcanism and rift-reactivated faults are normal faults because buoyancy related to plumes and to rift related elevation has locally modified the general stress distribution.

Elevation of shorelines and Later Cenozoic sediments at the coast of Africa is widely recognized but has proved difficult to relate simply to eustatic change. The reason for this appears to be that eustatic changes are superimposed on local tectonic effects related to the development of Africa's active basins, swells and rifts. These effects are distributed very unevenly around the continent and are particularly well developed around the shores of southern Africa where they have been intensely studied.

Seismic reflection data from the African continental margins show evidence of spectacular erosion and canyon cutting during the Oligocene (34-22 Ma) followed by an episode of low-stand deposition and salt tectonics. These events are attributable to erosion from the newly elevated African continent. Similar, although commonly less spectacular, low-stand episodes around other continents at about the same time are attributable to world-wide sea level lowering consequent on formation of the East Antarctic ice sheet. The circum-African phenomena certainly appear more striking than those around other continents, but better temporal resolution and more local studies will be needed before the exact role of African continental elevation in generating the erosional unconformity and subsequent low-stand depositional episode can be fully distinguished from the glacially controlled phenomena.

The development of basin and swell structure about 30 Ma has modified Africa's drainage pattern on a continental scale so that internal drainage has developed in several areas. The deltas of the great rivers that do reach the sea, such as the Nile, Niger, Zambesi and Limpopo, have prograded rapidly over the past 30 million years depositing abundant sediment eroded from the newly elevated areas of the continent. The Zaire, which had been a river with internal drainage, has reached the ocean as a result of river capture. The giant Zaire submarine canyon and deep sea fan have formed as a result.

How can being at rest with respect to at least part of the underlying mantle circulation account for all these distinctive and peculiar properties of the African Plate? A possible general explanation is this: When a plate is moving over the circulating mantle or asthenosphere at a typical plate speed of 50 mm/yr, an area 500 km in diameter at the base of the plate is carried away from an underlying plume or other hot area within a time interval of 10 My. Opportunity for interaction between the plume and the plate is therefore limited in time. Only huge plumes such as that beneath Hawaii are able to produce spectacular effects at the surface on fast moving overlying plates. Where there is no relative motion, hot rising mantle will interact over time scales of tens of millions of years with an overlying plate.

Hot, rising mantle may interact with a continent-bearing plate in several ways: (1) Because it is hot and of relatively low density, the plume will elevate the overlying area isostatically. It can do this even when the plume lies vertically below, but is not immediately in contact with the lithosphere at the base of the plate. (2) A hot plume, where in contact with the base of the continental lithosphere, will heat that lithosphere conductively. This is a slow process in thick continental lithosphere, the effects of which would be imperceptible at the surface for a long time. Conduction would have been unlikely to have made much difference in 30 My. (3) The plume can thin the overlying lithosphere by inducing partial melting. This process will operate more effectively on any fertile material which forms part of the mantle lithosphere.

The first process, isostatic uplift, appears to have happened in many parts of the African Plate over the past 30 My. The different length scales of the various swells indicate that underlying plumes may be of very diverse sizes and at various depths. It is not possible to tell whether the deeper parts of the continental lithosphere are now being heated from below by the second process at a rate that is faster than was common in earlier times. Eruption of volcanic rocks on the continent attributable to the third process has been somewhat concentrated in elevated areas, perhaps because of an isostatic response to low density, hot or depleted material at depth. Continental volcanism is almost entirely restricted to areas reactivated during Pan-African time which appear to be underlain by fertile mantle lithosphere.

The oceanic parts of the African Plate are underlain by relatively hot, young, thin and depleted mantle lithosphere. Plume interaction shows a different expression from that on the continent in the form of a fourth process: the development of swells, on at least two of which a conductive thermal anomaly has been mapped. Many of these swells are capped by 30 Ma and younger hot spot volcanic complexes such as are seen in the Canary and the Cape Verde islands. These young volcanic complexes are interpreted to be the products of rising plumes that have impinged on the base of the African Plate only within the past 30 My. Evidence of plume youth comes from the observation that, with the exception of Tristan, the hot spot volcanic islands of the African Plate do not have tracks. Geochemistry indicates relatively little involvement of the oceanic lithosphere that is cut by the young volcanoes. A fifth distinct process is represented by those parts of the African Plate and the conjugate parts of the American, Antarctic and Indian plates which have formed by sea-floor spreading within the past 30 My. There are no old plumes under this young ocean floor and hot spot volcanoes in these areas, such as Ascension, Shona and Bouvet, overlie new plumes.

Teleseismic studies have long been interpreted as indicating that the mantle beneath the African Plate shows heterogeneity with suggestions of slower velocities, usually interpreted as indicating hotter regions, beneath some high areas. With the advent of tomographic analysis of various kinds, this picture is beginning to come into better focus. Unfortunately the extremely small number and the uneven distribution of modern, broad-band seismic stations on the African Plate means that seismically derived information about the structure of the underlying mantle is at best tantalizingly incomplete. Nevertheless various studies indicate the occurrence of hotter material at both shallow and great mantle depths beneath the interior of the African Plate. The Afar, East African Rift System, Cameroon volcanic area and Cape Verde islands are places where indications of hotter mantle at depth appear in various different models. There appears to be a large and quite separate volume of low velocity material in the deeper mantle beneath the southern part of the African continent, perhaps because no cold lithospheric slabs have been subducted into that region over the past 200 My.

Why did the African Plate come to rest over the mantle circulation? One good answer might be that we do not know. Attention has been drawn to a temporal coincidence between the beginning of the current phase of igneous activity and rifting in Africa, and a change in the rate of continental collision in the Mediterranean. Perhaps this change arrested the motion of the African Plate over the underlying mantle? Another possibility is that the impingement on the base of the lithosphere of one or more giant plumes, such as those beneath the Cape Verde archipelago and the Afar, at about 30 Ma, served in some way to anchor the plate. A somewhat similar mechanism has been suggested to have slowed the motion of the Pacific plate with respect to the Earth's spin axis at about 110 Ma.

The past ~30 million years have seen radical changes in the global environment dominated first by the development of the East Antarctic Ice Sheet and, much later (at ~3 Ma), by the development of northern hemisphere glaciation with its huge oscillations in ice-volume. The peculiar condition of Africa with its rifts, volcanism and especially its basins and swells has played a part in its response to these changes. The influence of the formation of the East Antarctic ice sheet between 30 and 40 Ma appears to have been swamped in the off-shore seismic record of Africa by the effects of roughly contemporary continental elevation. The onset of Antarctic glaciation may have initiated, or at least strongly modified, the flow of the Benguela current which contributed to desert formation in Southwestern Africa and to the beginning of deposition of the arid environment sediments of the Namib formation. Elevation beginning at about 30 Ma was surely sufficient to cause some climatic modification on the continent but the biggest climatic change came much later with northern hemisphere glaciation and the development of the Sahara some 3 million years ago. That development was contemporaneous with those global changes that mark the onset of northern hemisphere glaciation.

Since the beginning of northern hemisphere glaciation, the African environment has shown fluctuations between wet and dry conditions in many areas. The rivers of northern Africa terminate in interior basins during dry intervals as the River Chari does in Lake Chad today. Great northern African rivers only reach the ocean, as the Nile, Senegal and the Niger do today, during wetter intervals.

Lakes have been abundant in Africa over the past 30 My both in basins of internal drainage and in the East African Rift System. The benign nature of the lake shore environment for hominid evolution has often been noted. But wherever hominids evolved, and the case for their having evolved in Africa is strong, we would not have access to such wonderful stores of fossil hominid and related remains were it not for outcrops exposed by faulting during the current episode of activity in the East African Rift System.

How can the Africa stationary hypothesis be tested? The most important need is for better temporal resolution of all the different phenomena that are involved. The good news is that new techniques such as fission-track and cosmogenic nuclide dating are becoming available and new efforts are generating the kinds of information that will make the required tests feasible. There is also a clear need for better understanding of the underlying mantle structure. This will call for deployments of modern broad-band seismometers as well as more study of deeply derived mantle xenoliths.

A continuing challenge will be to maintain a comprehensive approach to the solving of what has to be seen as a single problem: How has the African Plate behaved over the past 30 Ma and why has it behaved in such a distinctive manner? Integrated use must be made of all the different kinds of relevant data from all parts of the African Plate and from all of the different environments that extend from the deep mantle to the highest erosion surfaces, from the faults, volcanic rocks and sediments of the East African Rift System to the sediments and terraces of the continental margin and the hot spot volcanoes of the deep ocean floor. Alex L. du Toit set us an example in "Our Wandering Continents" by bringing many different kinds of evidence to bear on a single central problem. Du Toit's example will surely help to keep us all on the right track.

Introduction

One of the most striking aspects of *Our Wandering Continents* (du Toit, 1937) is the comprehensive way in which Alex L. du Toit approached the testing of the hypothesis of continental drift. His masterwork treated such diverse subjects as stratigraphy, paleontology, structural geology, paleoclimatology and continental shape in a thoroughly integrated way. He made use of all in establishing the history of the continents.

My thesis is that the African Plate has developed in a distinctive way over the past 30 My because it has been at rest with respect to at least part of the underlying mantle convection pattern during that time. Africa's development during those past 30 million years has differed both from its immediately previous behavior and from the behavior of the other contemporary great lithospheric plates. Du Toit's example has inspired me to attempt a comprehensive review because full understanding must surely require an integrated approach. I will therefore try to review what has happened in both the oceanic and continental areas of the African Plate while considering morphological, structural, igneous, erosional and depositional changes that have taken place. I will briefly discuss how these developments may have responded to global environmental changes and speculate about the lithosphere of the African Plate and the underlying asthenosphere.

Geological peculiarities of the African continent were discerned by Herodotus. Explorers encountering the Great Escarpment and the Great Rift Valley recognized more. Holmes (1944; 1965, his figure 763), following Krenkel (1922; 1957) and Argand (1924, his figure 6), was early in emphasizing that the continent has a distinctive basin and swell structure (figure 1 and 2). Krenkel suggested that young volcanism on the crests of swells in the northern part of the continent indicated a role for the mantle in generating Africa's topography. Wilson and I (Burke & Wilson, 1972), impressed by Krenkel's argument and by the observation that most African hot spots have no tracks, revived Krenkel's idea by suggesting that the African Plate had come to rest over the mantle circulation ~25 million years ago. More recently acquired data have led me to change this estimate of age, which is still in no way precise, to ~30 million years. Wilson and I thought that the present topography of the plate imaged the planform of the mantle circulation. McKenzie and Weiss (1975) made the same interpretation of Africa's basin and swell topography. They too recognized that Africa had moved little with respect to the spin-axis of the earth over the past 200 My (Burke & Dewey, 1974) and suggested that interaction over that long period had generated the topography. Thiessen *et al.* (1979) estimated that although the distribution of the crests of high areas on the African Plate is irregular, a crestal separation of about 700 km is quite common and comparable to the depth to the base of the transition zone. Later Ashwal and I (1989) attributed the virtual absence of Neogene volcanism on the older cratonic areas of Africa and its concentration in areas of Pan-African reactivation to the abundance of fertile lithospheric mantle beneath the reactivated areas in contrast to older cratons which were underlain mostly by highly-depleted mantle lithosphere.

Bailey (1992; 1993) suggested that a change in the velocity of collision between Africa and Europe about 30 million years ago might have been a triggering event which induced the change in the style of behavior of the African Plate. The beginning of convergence in the Atlas Mountains has also been considered roughly coincident with both the start of the present episode of rifting in East Africa and with the restructuring of Indian Ocean spreading centers. Bailey pointed to collisional change in the Mediterranean as possibly having influenced a contemporary upsurge in igneous activity in Africa. The idea of the triggering of African change by an event related to the continuing African-European convergence is not of course incompatible with the earlier recognition of a stationary Africa. Tarduno & Sager (1995) observed that the Pacific Plate moved very slowly with respect to the spin-axis during an episode of formation of several giant oceanic plateaus from plumes at about 110 Ma. They pointed out that Morgan & Smith (1992) had suggested plume activity might reduce asthenospheric viscosity in neighboring areas and slow down relative motion between a plate and the underlying mantle circulation.

Nyblade & Robinson (1994), in discussing a large area of the African Plate on both continent and ocean (the "African Superswell"), suggested that a volume of the mantle beneath that area might be hotter than average. They cited Anderson (1982) who pointed out the possible warming effect on the mantle of blanketing of an underlying volume by the Gondwana continent over the interval between about 500 to 160 Ma. Nyblade & Robinson, citing the work of Richards & Engebretson (1992) and Scrivner & Anderson (1992), also drew attention to the observation that cold slabs of lithosphere have not been subducted beneath the area presently occupied by the African Plate for about 200 million years. Deep mantle tomography (Su *et al.,* 1994; Zhou, 1996) has improved understanding of the location of a deep mantle low velocity volume beneath southernmost Africa. Nyblade & Robinson (1994) further suggested that the development of the present peculiarities of the African Superswell and the present elevation of that area might be linked to "a thermal perturbation of the lithosphere."

Here I review a variety of aspects of the evolution of the African Plate over the past 30 million years and consider how those developments have been linked to changes in the global environment and to the underlying, but as yet poorly known, deeper structure of the mantle. I point to some new developments that are making the testing of the "Africa Stationary" hypothesis easier and suggest some directions for future research, but first I must describe Africa as it was in the interval immediately before 30 Ma to show that something distinctive really did happen at about that time.

Africa from 65 Ma to 30 Ma: A quiet interval?

Sea-level was high over much of the world during the Late Cretaceous and extensive regions of Africa were then submerged (Sahagian, 1988). The great Jurassic to Cretaceous rifting episode, which had extended over a large area of the African continent, started at about 160 Ma. This rifting episode that accompanied the departure from Africa of Antarctica, North and South America, Madagascar, and India with the Seychelles, had run its course by 65 Ma. Later, several of the intra-continental rift systems continued to accumulate thick sediments suggesting some continuing active faulting on those rift margins (Figure 3).

Marine sediments that were deposited on the African continent during Paleocene and Eocene times (from 65 to 35 Ma) are nearly all preserved close to the edge of the continent and are best developed in the Mediterranean coastal area, particularly in the Sirte Rift Basin. Thicknesses of more than a few hundred meters of sediment are reported from other basins at the continental margin including the Aaiun Basin, the Togo-Benin Basin and the southwestern part of the Benue Rift. Some rifts of the West African and Central African Cretaceous systems continued to accumulate non-marine sediment (Genik, 1993, his figures 8 and 14). Near the east coast of the continent, a kilometer or more of non-marine Paleogene sediment was deposited in the Anza Rift (Bosworth & Morley, 1994). The marine section is thick in the Lamu embayment at that rift mouth (Walters & Linton, 1972). More than a kilometer of earlier Cenozoic sediments accumulated in the coastal basins of Tanzania (Kent *et al.*, 1971) and in Mozambique around the area of the Zambesi mouth (De Buyl & Flores, 1986).

While marine deposition was concentrated at the edge of the continent, most of the interior continental surface of Africa was dry land. Many of Africa's widely distributed and deeply weathered rocks, including laterites and bauxites, appear to have developed during the 65 to 30 Ma interval. It is important to emphasise that, except in those cases where weathered material is interbedded with datable rocks, it is almost impossible to determine the age of weathering of an old rock. In South Africa: ". . .in the case of the oldest of the post-Mesozoic [erosional] surfaces deep kaolinization of the underlying bedrock was present beneath massive laterite or silcrete duricrusts. . . ." (Partridge & Maud, 1987, p. 182). In West Africa (Figure 3 based on work of Egbogah, 1975), bauxites are widely distributed and have been elevated to varying extent during the past 30 My. Bauxites occur in West Africa at elevations from sea level in Sierra Leone and Guinea to 2.5 km above sea level in Cameroon. There is some indication in the Senegal Basin (Spengler *et al.*, 1966) that deep weathering is recorded in Eocene sediments interbedded within a Late Cretaceous to Late Eocene sequence. In Eritrea, laterites that underlie the earliest basalts ". . .represent a widespread and lengthy period of humid conditions in the early Tertiary and developed on a near sea level plain" (Drury *et al.*, 1994, p. 1372). Laterites in Yemen occur at the top of the Cretaceous to early Tertiary marine to terrestrial sandstones of the Tawilah Formation (Menzies *et al.*, 1990; Geukens, 1966). In northern Arabia, at that time part of the African continent, bauxite was formed during the Cretaceous (Collenette & Grainger, 1994). Remnants of a possibly related extensive surface are preserved at high elevations in Kenya and Uganda where a thick lateritic cap and deep weathering have developed. This surface is known by various names including "Mid-Tertiary." The overall picture is of a continent with low relief, rolling topography and few high mountains that was undergoing deep weathering under humid conditions.

During the 65 to 30 Ma interval, nearly continuous igneous activity within the African Plate appears to have been restricted to the hot spot track of the Walvis Ridge in the Atlantic and to a part of the Cameroon line on the continent. Most of the volcanic material erupted from the Deccan trap plume during this interval lies on the Indian Plate and the plume appears to have underlain the active spreading center between the two plates only episodically. On the African Plate, only parts of the Mascarene Ridge including the Saya de Malha bank and Nazareth bank are made of material erupted from that source. Isolated, short-lived igneous events have left a record in various parts of the plate. The emplacement of the Mwadui kimberlite took place at 41 Ma (Davis, 1977) and granite was intruded into Jebel Uweinat at 45 Ma (Schandelmeir *et al.*, 1983). Ten million years of basaltic eruption close to the Kenya-Ethiopia border between 45 and 35 Ma (Ebinger *et al.*, 1993) constituted a relatively long-lived local intraplate episode. The early basalts of the Lotikipi area at the Kenya/Sudan border (Morley *et al.*, 1992) have ages from 35 to 30 Ma. Isolated areas of volcanic rock record eruptions between 60 and 35 Ma at several localities near the coast in southwestern Africa and on the Agulhas Bank (Duncan *et al.*, 1978; Dingle & Gentle, 1972).

The Cameroon line is in some ways the most extraordinary geologic feature of the African Plate. It is marked by a dozen volcanic areas, on shore and offshore, that have been active during the Quaternary. Volcanism has been episodic at some of these sites for up to 30 Ma (Lee *et al.*, 1994). Lee *et al.* (1994) also report that "at least 17 plutonic complexes" in the continental part of the Cameroon line represent "igneous activity from 65 to 30 Ma." The general area of Cameroon and the neighboring Benue trough was the site of much igneous activity during the Cretaceous (Whiteman, 1982). The Burashika volcanics, dated at 135 Ma by Richard Armstrong (Burke, 1976a, p.100; Maluski *et al.*, 1995; Fitton, 1980), represent the oldest volcanism of the Benue trough. Igneous activity in this general area may have been close to continuous since the Early Cretaceous (~140 Ma). Lee *et al.* (1994) present a model of a "sub-lithospheric hot-zone" underlying the Cameroon line which has propagated into the ocean and interacted with several rising plumes since 30 Ma.

A not wholly dissimilar possibility is that the Cameroon area has lain above the same volume of hot asthenospheric mantle since the Cretaceous while other parts of the African Plate have moved over the mantle. My reason for suggesting this is indicated in Figure 4 where Mesozoic and Neogene magnetic poles for the African Plate are plotted. During both intervals Africa has been at rest with respect to Earth's dipole magnetic field, and by inference, the earth's spin axis. In between, the plate has moved with respect to the spin axis by a rotation of about 20 degrees about a poorly localized pole in the Gulf of Guinea but simply remembered as being close to zero degrees latitude and longitude. If being at rest with respect to the spin axis of the earth is the same thing as being at rest over the mantle circulation pattern, which is possible, then the Cameroon area may not have moved with respect to an underlying "hot-zone" for 130 My or more.

The oceanic volcanic sites along the Cameroon line are typical examples of the many hot spot volcanic areas that have developed and persisted over much of the African Plate during the past 30 million years. The individual active Cameroon volcanic zone hot spots are by no means peculiar and require no special origin. There is no need to invoke a special hot zone propagating horizontally into the ocean because the volcanoes of the Cameroon line represent only a small part of a large population of 30 Ma and younger plumes. Pavoni (1993) has suggested, using quite other evidence, that the Cameroon area has been critically important in the tectonic evolution of Africa over an interval even longer than 140 Ma; Fitton (1980) long ago drew attention to a simple geometric relationship between the Benue trough azimuth and that of the Cameroon line; and Hartley *et al.* (1996) consider northern Cameroon to be an area of minimal elastic thickness.

About 2000 cubic km of mainly basaltic lavas were erupted between about 45 Ma and 34 Ma to form a 500 m thick pile in the area now occupied by the southern main Ethiopian rift (Ebinger *et al.*, 1993). This episode was unaccompanied by much extension at the surface and has been interpreted to represent hot spot volcanism above a mantle plume. The basalts of Lokitipi, a few hundred kilometres to the southwest (Morley *et al.*, 1992), which lie along an azimuth from the southern Ethiopian basalts that is roughly concentric with the trend of the Walvis Ridge for the interval 40 to 30 Ma, might be considered a later manifestation of the same plume. The products of any plume at Lokitipi soon became indistinguishable from volcanics associated with the early development of the East African Rift.

Collision with Eurasia along the northern boundary of the African Plate slowed at 65 Ma and accelerated at 35 Ma (Dewey *et al.*, 1989; Figure 3). In the Indian Ocean, reorganisation of the plate boundary at ~30 Ma (38 Ma according to Patriat & Segoufin, 1988) to form the present central and northwest Indian Ridgess, was one of the most prominent events that took place at about the time when Africa came to rest over the mantle circulation.

In summary, over the interval 65 Ma to 30 Ma, the African continent and the African Plate appear to have been the site of scattered intraplate volcanism and relatively little tectonic activity. The two more extensive continental areas in southern Ethiopia and Cameroon, where there was intraplate volcanism, were places where there was to be participation in the surge of 30 Ma and younger igneous activity that was about to begin and which continues today. At ~30 Ma, the position of the eastern boundary of the African Plate changed as a result of a ridge jump in the Indian Ocean. This was the time when widespread intraplate volcanism developed over much of the African Plate and, as evidenced by offshore seismic reflection data, the continent began to rise.

Bailey (1993) has suggested that the quiet tectonic state of Africa may have been related to a temporary slowing of collisional processes at the northern margin of the plate in the Mediterranean while the change to the present state relates to the resumption of collision between Africa and Eurasia. Here is how that might have worked. Ridge push is likely to have been the dominant plate-margin generated force on the African Plate between 65 Ma and 30 Ma, as it is today (Zoback, 1992; Figure 3). Continental collision causes elimination of the slab pull force for the length of the boundary involved in collision and substitutes a collisonal force that is probably much smaller. During the interval 65 Ma to 30 Ma, slab pull may therefore have been a more important force on the African Plate than it was after collision resumed. Another possibility is that the eruption of mantle plumes themselves pinned the plate (cf. Morgan & Smith, 1992).

There is a need to better understand the structural history of the African Plate over the 65 Ma to 30 Ma interval. Unfortunately the detailed history of much of the African Plate is relatively poorly known for that time. This is especially true for the area of the continent that was dry land where much of the record is of rock weathering and temporal resolution is poor. It would be interesting to be able to recognise structures indicative of the prevailing tectonic regime that formed during this relatively poorly known time.

A surge in intraplate volcanism

Many Plumes

The African Plate is unique on earth in the number of places at which intraplate volcanoes have erupted during the past 30 My. This volcanic activity is not evenly distributed and people who live in South Africa have a good reason for not being especially impressed by it because they live in the largest area of the plate that has escaped eruption. Histograms of numbers of age determinations against ages of rocks (Figure 5) indicate how volcanic activity has surged on the continent and the map of Figure 6 shows the widespread distribution of young volcanic rocks within the plate interior.

Apart from many of the younger volcanic rocks that are associated with the East African Rift System, volcanic rocks of the African Plate interior have not been generated in exactly the way that most commonly produces pressure-release volcanic rocks. Pressure-release rocks are normally produced when a local reduction of the overlying lithospheric load reduces pressure in the mantle so that a volume of the shallow underlying mantle crosses the basalt solidus to generate magma which mostly erupts as basalt. This mechanism does not normally operate in the interior of plates which are generally in compression from plate boundary forces (Zoback, 1992) but may occur if some kind of perturbation of the plate boundary forces generates local extension (see for example: Burke & Lytwyn, 1993).

I suggest that in much of Africa, the required perturbations over the past 30 My have come from below and not, as is more common, from the plate margins. Mantle plumes, rising hot bodies in the underlying asthenosphere, are the cause of young volcanism in most areas of the African Plate. Because the plate is not moving over these mantle heterogeneities, they have had time to interact with the lithosphere and to produce new elevations and new volcanoes. Some features of the volcanism appear from a glance at Figure 6. The volcanic provinces are numerous (~40), the exact number depending on how neighboring volcanic centers are lumped together. The postulated underlying plumes are not all the same size: Afar, the Canary islands and the Cape Verde islands look very big, whereas Air is quite small. Plumes appear irregularly spaced and underlie both continent and ocean floor. Figures 5 and 6 together give a general idea of the distribution of volcanic activity and of the time when it started, but both the map and the histograms of igneous rock ages conceal a good deal of complexity.

Age distribution of young volcanic activity on the African Plate

The histograms of Figure 5 might be taken to indicate that volcanic activity has been increasing steadily on the African continent over the past 30 My, but this is not necessarily the case. The more recent volcanic rocks are better exposed and preserved so that they are more accessible and more suitable for isotopic age determination. Some of the youngest rocks occur in association with human artefacts and fossil hominids so that they are particularly likely to be studied. It must be emphasized too that the quality of the data in the histograms is uneven. Some of the measurements done a long time ago using the K-Ar age method are proving to be less reliable guides than had been hoped. In cases where new $^{40}Ar/^{39}Ar$ age determinations are being made on the same rocks, there are sometimes significant differences. At some time in the relatively near future it will probably prove appropriate to discard much of the older data set. A clearer picture of time dependent behavior will surely emerge. At present, it looks as though there was an overall change from a condition of little and episodic volcanism on the African Plate before 30 Ma to one of abundant and continuous volcanism afterward. I would be reluctant to infer more about temporal variation in plate-wide volcanic activity, although temporal variations in individual provinces have been well resolved and do show significant variations.

Compilations that include some proportion of the older data are still useful and give an indication of temporal evolution. For example, Table 1 from Mazarovich (1990) gives a good idea of how the volcanic activity of several different hot spots off the northwestern coast of Africa had begun by 25 Ma. Only in Fuenteventura, in the shoreward Canary islands, is there a suggestion of older activity.

The Canary islands, which are one of the larger volcanic provinces, have been interpreted as showing progressively younger ages of the onset of volcanism from west to east over a distance of about 500 km (McDougall & Schmincke, 1976). Because other provinces show no progression over the same interval and because some provinces show a progression of ages in other directions, the Canary age progression is not related to systematic movement of the African Plate. We may rather be seeing some indication of structure or temporal evolution within an underlying plume complex or possibly an effect of proximity to the continental margin (Schmincke, 1982).

Isotopic ages of volcanic rocks around 25 Ma are recorded from many areas on the African Plate and Table 1 of Mazarovich (1990) provides a good example. Many separate plumes, perhaps as many as forty, were operating by the beginning of the Miocene (~22 Ma). The volcanic areas of the African Plate permit a simple test of the “Africa at Rest” hypothesis. The sketch in Figure 7 shows three volcanic areas that lie thousands of kilometres from each other. All have experienced intermittent volcanism within the same roughly elliptical areas since about 30 Ma. These three areas are the Afar (Schilling *et al.*, 1992, their figure 19), Tristan (O’Connor & le Roex, 1992) and Principe (Lee *et al.*, 1994). On the basis of ages from these localities, the rigid African Plate cannot have moved for more than a maximum of 300 km in any direction over the past 30 million years. This reduces to a velocity of 10 mm a year which is below that resolvable for moving plates using conventional observations. I conclude that motion below a threshold velocity of about 10 mm a year is required to produce the extraordinary effects which now distinguish the African Plate from all others.

For shorter intervals, a demonstration of Africa's stationary character can be derived in a different way. Figure 8 shows that for the duration represented by the Nuvel-1 data (DeMets *et al.*, 1990) which extends back ~3.7 My, Africa has hardly moved at all with respect to a fixed hot spot population (Gripp & Gordon, 1990) and is rotating very slowly about an internal pole at a rate below that usually considered resolvable. It is intriguing that the place that appears to have moved least on Gripp & Gordon's (1990) map, which is close to zero degrees latitude and zero degrees longitude, is in the same general area as the long-lived Cameroon igneous activity, as well as close to the pole that describes the rotation of the African Plate with respect to the spin axis between Cretaceous times and about 30 Ma. Continuous igneous activity in this general area for 140 My may be related to a lack of motion with respect to the underlying mantle circulation.

Hot-spot tracks on the African Plate

The Tristan and Gough hot spot and the Walvis Ridge, which forms its track of older volcanic material, were among the first hot spot manifestations identified on the African Plate. Many other hot spot tracks have since been postulated to exist on the plate and Figure 9 illustrates a compilation of those tracks (from Summerfield, 1996).

Analyses of hot spot tracks on the African Plate were made by Morgan (1981) and Duncan (1981). Duncan (1981, p. 30) described how he plotted tracks: "Specifically, the motion of a single plate over hot spots underlying it is determined from the geometry and dated localities along the hot spot paths." Of thirteen African Plate hot spot tracks shown on Duncan's (1981) figure 1, only five are shown with "dated localities" and one of those tracks is very short. The implication is that "the geometry" was relied on for at least eight tracks. I interpret "the geometry" as involving two assumptions: (1) that all hot spots have persisted in existence throughout the life of the African Plate, that is for about 150 My, and (2) that the hot spots have not moved with respect to each other. To the extent that these assumptions are valid, it can be inferred that all hot spot tracks on a single plate are concentric about a pole that describes the motion of one of the set of hot spots. On the African Plate, the temporal record is dominated by ages that represent the position of the Tristan and Gough hot spot source, or mantle plume, as it built the Walvis Ridge.

It is appropriate to review Duncan's way of plotting hot spot tracks for three reasons: 1) much more is known today about relative motion among hot spots, 2) with the advent of space-borne altimetry (Sandwell *et al.*, 1994; Smith & Sandwell, 1994), the bathymetric evidence for volcanic accumulations along postulated tracks can be better examined, and 3) many more $^{40}Ar/^{39}Ar$ ages of volcanic rocks are now available from the oceanic parts of the African Plate, mainly through heroic efforts by Duncan himself.

Relative motion among hot spots has been reviewed by Acton & Gordon (1994) and Atwater (1989, p. 41) who are cautious about assuming that hot spots remain fixed with respect to one another for long intervals. Norton (1995) is more skeptical about that assumption, and Cande *et al.* (1995) occupy what seems to be an intermediate position. Although all of the African hot spots underlie a single plate, the tracks reach a maximum age of about 150 Ma and the possibility of relative motion among hot spots, especially for older times, is worth bearing in mind.

Duncan's (1981) study linked active hot spot volcanoes, mainly on the African Plate, to igneous rocks on older ocean floor or on the neighboring continent, even where there was no strong bathymetric evidence of a track. The assumption appeared to be that the underlying source or mantle plume remained in existence but erupted no volcanic material so that no evidence of a track at the surface was needed. Morgan (1981) applied similar reasoning to a more extensive area of the earth and to a much longer period. Because hot spot tracks like Tristan and Gough and the great tracks of the Pacific Ocean have generated so much abundant volcanic material, I suggest that drawing hot spot tracks through areas with little or no volcanic rock should be done with caution. Hot spot tracks from active volcanoes without older igneous rocks along the track and tracks that link only a small number of igneous rocks of different ages, especially where that number is as low as two, should not generally be used.

The availability of better bathymetry, and especially of satellite altimetry, encourages a more cautious approach to hot spot track mapping. These data sets show that very few of the hot spot volcanoes on the oceanic part of the African Plate are associated with mappable tracks of volcanic material (Figure 10), although there are places where future sampling and dating of other groups of seamounts may reveal additional hot spot tracks. Figure 10 shows no hot spot tracks on the continent of Africa, although there is some evidence for a very short track between southern Ethiopia and Lokitipi dated between 45 and 35 Ma. On the oceanic part of the plate, there are only three well-developed tracks: 1) The Walvis Ridge ending near Tristan and Gough islands; 2) The Mascarene Ridge including the Saya de Malha, Nazareth Banks and Cargados

Carajos; and 3) The Madagascar Plateau extending southward for nearly 1000 km from Madagascar. These three are all the tracks that are mappable in the form of large volumes of ocean floor volcanic rocks forming relatively shallow areas. To complete the map, possible incomplete tracks for St Helena (O'Connor & le Roex, 1992) and the Sierra Leone Rise hot spot (Schilling *et al.*, 1994) could also be plotted. The plume for the Madagascar Plateau track became inactive at about 60 Ma at a time of plate reorganization (Norton & Sclater, 1979) and no volcanic material marks any southward prolongation. The Marion hot spot on the Antarctic Plate has been suggested to be a trackless manifestation of the Madagascar object (Storey *et al.*, 1995) but does not meet the criteria used here.

The hypothesis being tested in this paper is that Africa has been at rest with respect to the underlying mantle circulation for the past 30 My. If that is a valid hypothesis, then plume tracks which relate to active volcanism on the African Plate must end in places where there has been intermittent volcanism since 30 Ma. In the neighborhood of Tristan and Gough islands, which both display evidence of volcanism within the past million years, isotopic ages at three localities range from 31 Ma to 8 Ma. The localities of the two volcanic islands and the three dated seamounts are distributed over an elliptical area of ~500 km by ~200 km (O'Connor & le Roex, 1992) and thus are consistent with the "Africa Stationary" hypothesis (Figure 11).

Réunion, Mauritius and Rodrigues: Young hot spots?

The island of Réunion has been widely regarded as the present site of the plume that generated the rocks of the immense Deccan trap flood basalt province at about 65 Ma (Duncan *et al.*, 1990; Peng & Mahoney, 1995), but the trend of Réunion, Mauritius, Rodrigues island and Rodrigues Ridge (Figure 6), which form a 1000 km long line roughly at right angles to the trend of the postulated Deccan hot spot track, was long ago recognised as calling for some modification of this interpretation (Morgan, 1978).

The evidence of Figure 12 shows that the Réunion hot spot is unrelated to the Deccan trap plume. If the African Plate has not moved with respect to the mantle plume population for the past 30 My, an active manifestation of the Deccan trap plume source would have to be where it was at 30 Ma. That is, under Nazareth Bank or Cargados Carajos Bank. Over the past 30 My, there has been no volcanism in that area. The Deccan trap mantle plume source therefore appears to have been dead for 30 My and has been buried under a thick carpet of carbonate beneath the Cargados Carajos and Nazareth banks. This interpretation is compatible with the

"Africa at Rest" hypothesis. There is one other possibility. All the other plumes might have remained fixed, and Réunion could have moved 700 km independently of the rest of the sub-African Plate plume population. That, I reject.

It has long been recognised that until the restructuring of the Central Indian Ridge took place after anomaly 13 time, 38 Ma (Norton & Sclater, 1979; Patriat & Segoufin, 1988), the Deccan source plume generated on axis and, close to the spreading axis, hot spot volcanic rocks building the Chagos-Laccadive Ridge and the Mascarene Ridge (Duncan *et al.*, 1990). The isotopic, stratigraphic and morphologic data show that when the ridge jumped off the hot spot at the time of plate reconstruction, the Deccan source plume died. Rocks from Nazareth Bank in ODP Site 706 (see Figure 12 for locations) gave an age of 33 Ma, and rocks from the Texaco Nazareth Bank well (NB-1) gave an age of 31 Ma (Duncan & Hargraves, 1990). Both banks are covered by more than a kilometer of carbonate sediment and there is no evidence of progressively younger igneous rock having been generated farther south along the projection of the Deccan hot spot track until the Islands of Mauritius and Réunion are reached. The oldest ages on Réunion are only 2 Ma. The ages of the volcanic rocks beneath Nazareth Bank are therefore interpreted as reflecting the time of the demise of the Deccan trap mantle plume source.

Rocks from Réunion, Mauritius, Rodrigues island and Rodrigues Ridge yield an African intraplate style of age distribution with basaltic eruption persisting in the same places but without the generation of a mappable hot spot track (Figure 12). The ages of intermittent volcanism in these areas extend back to 10 Ma on Rodrigues Ridge (Duncan & Hargraves, 1990). These volcanoes thus form part of the 30 Ma and younger African intraplate hot spot population. They are unrelated to the Deccan plume because no material marking the development of a plume track exists for the interval 30 to 10 Ma.

The distinction of the Réunion hot spots from the Deccan trap source encourages a new look at the results of ODP Leg 115 which drilled in various parts of the Deccan plume track and fostered much laboratory study. For example, it is no longer necessary to postulate true polar wander of the earth's dipole field to account for the paleomagnetic results (Vandamme & Courtillot, 1990). The Deccan eruptions, along with those recorded at ODP sites 706, 715 and 707, all took place close to 25^{o} South. Because Réunion is unrelated to the plume that gave rise to these materials, its eruption at 21^{o} South does not provide evidence of true polar wander. Geochemical data can be reviewed from the same perspective (Lytwyn & Burke, 1995).

In summary, evidence from hot spot tracks on the African Plate shows that they are few in number and that the plate stopped moving over the mantle circulation about 30 Ma. This leads to the question: Does the geographic distribution of post 30 Ma volcanism throw light on the nature of the African lithosphere, and on the underlying circulating mantle since the plate came to rest?

What controls the geographic distribution of 30 Ma and younger volcanic activity on the African Plate?

Kennedy (1964; 1965) pointed out that 30 Ma and younger volcanism on the African continent is almost entirely confined to areas that have been involved in Pan-African events. This concentration is certainly a striking feature of the regional distribution of African volcanism, but there are others as well. For example, young intraplate volcanoes occur on both continent and ocean, and as Krenkel (1922) observed, there is a concentration of young volcanism on the crests of topographic swells in those areas that we now know to have been affected by Pan-African events. These aspects of the geographic distribution of young volcanism are here interpreted in terms of the "Africa Stationary" hypothesis. Relevant data are depicted on Figure 6 which shows volcanic areas, basins and swells and active rifts, as well as plate boundaries.

The most obvious feature of Figure 6 is that it shows a large number of places at which volcanism has begun and continued intermittently within the past 30 Ma. During the previous 35 My, the number of intraplate volcanic areas totalled no more than ten (Figure 3), but volcanism did not persist for long in nine of those areas. Only in a region on the Cameroon line was igneous activity persistent over the whole 65 to 30 Ma interval. The distinctive change in the behavior of the African Plate over the past 30 Ma is clear. Not surprisingly, the distinctive character of young African volcanism reveals itself in different ways on the continent and on the ocean floor, so I discuss these separately.

Young volcanism on the oceanic part of the African Plate

Intraplate volcanic areas on the oceanic part of the African Plate are unevenly distributed and this is certainly in part, but not entirely, because the ocean floor is quite poorly known. Figure 6 shows about fifteen volcanic areas with concentrations at Tristan and Gough (O'Connor & le Roex, 1992), off the northwestern coast of the continent (Mazarovich, 1990; 1994), in the Cameroon line (Lee *et al.*, 1994), in the Comores and in the Réunion-Mauritius-Rodrigues area.

Tristan and Gough are distinct because they lie at the end of Africa's only active hot spot track, but other areas also show distinctive features. The Madeira, Canary islands and Cape

Verde archipelagoes are very large and very well studied. The Cape Verde islands have been the site of a splendid study that combined heat flow, topographic and satellite altimetric data to establish the presence of an underlying mantle plume (Courtney & White, 1986). A second extremely helpful analysis of the gravity field (Morgan *et al.*, 1995) treats the Cape Verde archipelago as a place that has not moved over its underlying plume.

Perhaps because of the success of the Cape Verde study, the idea that mantle plumes have characteristic diameters of 1000 km or more has been much elaborated (White & McKenzie, 1989). This diameter is a bit small for the largest of topographic swells on the African Plate (Nyblade & Robinson, 1994; and the Ethiopian and East African swells depicted on Figure 6). Many discrete areas of active volcanism on the African Plate that are associated with topographic or basement uplifts, and in some cases with well-mapped negative Bouguer gravity anomalies, are much smaller in area. Topographic and structural cross-sections from Dakar and the Ahaggar in Figures 13 and 14 serve to illustrate this smaller scale which can also be seen in the relationship of the volcanoes to topography and negative Bouguer gravity anomaly in the Jos Plateau (Figure 15). In some places (e.g., Ahaggar) separate, elevated volcanic areas are clustered within a regional topographic high and perhaps fine structure within a plume is being discerned. A general conclusion might be that there is a range of elevated areas from the largest, which are the four discrete uplifts that together make up the "African Superswell" of Nyblade & Robinson (1994), to small features, about 200 km across. Elliptical shapes are general with longer axes typically about twice the length of shorter axes. This ellipticity appears unrelated to anisotropy in the surface geology or to lithospheric structure, and is perhaps related in some way to a mantle flow pattern.

The elevation in the Ahaggar area is much more impressive than is the relatively small amount of volcanic material erupted. Apart from areas in the East African Rift System where quite large volumes of volcanic material have been erupted (perhaps a total of 100,000 cubic km from the Samburu plume over ~20 Ma, using data of Smith, 1994), the general rule is that African hot spot volcanism is of small volume. This is to be expected from observations on the ocean floor where the importance of the HIMU source, which is generally represented by only small amounts of partial melting, has been recognised. The HIMU source is one of several isotopically distinct reservoirs which have been inferred, from geochemical studies, to exist within the Earth's mantle (Hart & Zindler, 1986). These reservoirs are defined on the basis of relative proportions of various radiogenic nuclides. I make frequent reference to HIMU and to three of the other distinct mantle reservoirs: EM1, EM2 and DMM, or the MORB source. Discrimination of these reservoirs has proved helpful in improving the understanding of the origins of the igneous rocks of the African Plate. A similar observation of basement elevation applies to the oceanic areas of the African Plate where, in contrast to the huge volcanic accumulations of the Pacific Ocean hot spots, there is much evidence of both basement uplift and eruption of relatively small volumes of volcanic material.

In the Canary and Cape Verde archipelagoes, underlying ocean floor is exposed in the volcanic islands of Fuenteventura and Maio. Similar evidence of basement elevation can be discerned in regional seismic lines (Figure 16a and b; Lehner & De Ruiter, 1977; Rosendahl *et al.*, 1991). Evidence of basement uplift at hot spots is general on the continent of Africa (Figures 13, 14 and 15). Figure 17 shows that this is in striking contrast to the situation in Hawaii and perhaps other Pacific Ocean hot spot volcanoes where volcanic rocks represent a huge pile that has depressed the underlying basement to inaccessible depths, and flexed the lithosphere into a moat and a rim (Wessel, 1993). Hot spot volcanoes of the African Plate, with the exception of Tristan and perhaps the Afar, the northern part of the Cape Verde archipelago (Gerlach *et al.*, 1988) and Réunion, are characterised by HIMU sources and represent only a small degree of partial melting of the mantle. In contrast, the Hawaiian volcanoes represent an enormous amount of partially melted material, and the volcanic piles of Hawaii represent a huge load on the lithosphere. On the African Plate, the lithosphere is not loaded down but is elevated by underlying mass deficient material. This mass deficient material probably includes contributions from three sources: (1) hot, low density material at shallow depths, which includes partly molten material; (2) low density residual material at greater depths, from which relatively iron-rich basalt has been melted; and (3) mass deficient hot plume material at even greater depths (see Morgan *et al.*, 1995).

Only Tristan and Gough, Afar and perhaps Réunion are hot spots with large volumes of volcanic rock rather than dominant basement uplift. It is perhaps significant that these are the only African Plate hot spots with high ^{3}He signatures. These may be the only active localities on the African Plate with plumes from the deep mantle. This is in contrast to plumes from above the base of the transition zone with low ^{3}He signatures that produce small volumes of volcanic material and a dominance of basement uplift.

In summary, hot spot volcanoes, apparently built above mantle plumes of a variety of sizes, are patchily distributed over the ocean floor of the African Plate older than 30 Ma in age.

Several of these volcanic areas are associated with elevated ocean floor and their volcanic accumulations appear to be much smaller than those of the great hot spots of the Pacific Plate. Among active volcanic areas on the African Plate, only the Tristan and Gough complex lies at the end of a long hot spot track marked by large volcanic accumulations. Tristan has a distinctly different geochemistry and its plume source reached the surface at ~130 Ma. All the other hot spots are related to plumes that reached the surface after ~30 Ma. There is no evidence of systematic relative motion between the African Plate and any of these young underlying plumes.

A distinct environment is represented by ocean floor that formed during the past 30 My on the African and conjugate neighboring plates (Figure 18). Near-axis hot spots such as Ascension, Bouvet and Shona lie on this kind of ocean floor. If Africa has been fixed over the mantle circulation, these features must relate to new plume activity. They cannot be manifestations of old plumes which currently must lie under 30 Ma or older ocean floor. Areas of young ocean floor conjugate with the 30 Ma and younger ocean floor of the African Plate have formed on the South American, Antarctic and Indian plates. Ascension island, on the South American Plate, is the best known of the hot spots formed in this distinctive environment. The oldest rocks in Ascension are about 1.5 million years old and the island lies on ocean floor that is itself only a few million years old. Ascension lies very close to the South Atlantic spreading center and has no track on either the African or the South American plate (Sandwell *et al.*, 1994). The track for Ascension depicted in Duncan (1981, figure 2) does not meet the rigorous criteria applied here. Not only is it without morphological expression, but no dated igneous rocks lie along it. The Freetown norite, which has yielded a 194 Ma age, was suggested to have been related to Ascension. Freetown, however, is not close to any mappable geometric track but is an uplifted intrusion emplaced during Jurassic rifting of the West African margin. The topographic uplift at Freetown is one of the familiar 30 Ma and younger uplifts of Africa.

Schilling *et al.* (1992), in developing a suggestion of Morgan (1978), have discerned the influence of a number of currently active South Atlantic hot spots in the modified composition of basaltic rocks along the neighboring South Atlantic spreading center. This intriguing observation suggests that even plumes which appear quite small from their surface manifestations could be influencing magma generation over horizontal distances of as much as 1000 km. Thompson & Gibson (1994) have made similar suggestions of long distance influence from studies of the volcanic rocks of the East African Rift System. Because the magma source for all these hot spots seems to be HIMU material affected by only a small degree of partial melting, it is something of a challenge to picture how this remote sampling might work.

One possibility (Figure 19) is that production of MORB requires extraction of partial melt from a huge volume of mantle (McKenzie & Bickle, 1988). If an active plume is generating a hot spot within a few hundred kilometres of an active spreading center, then that spreading center has access to a somewhat smaller source volume (Figure 19). The subtle distinction detected by Schilling for spreading-center rocks relatively near to off-axis hot spots reflects a change in the source reservoir, as Schilling suggests. This change, however, is attributable to sequestration of a volume of that reservoir by the neighboring hot spot rather than to the supply of material from the plume beneath the hot spot to the spreading center. The difference in MORB geochemistry in sectors of the spreading centers close to hot spots results from access to a smaller reservoir in the mantle.

Young volcanism on the African continent

The dominant controls on the areal distribution of young continental volcanic rocks of Africa are those first identified by Krenkel (1922) and Kennedy (1965). Volcanoes occupy the crests of structural and topographic swells and volcanism is virtually confined to areas reactivated during Pan-African times. The observation that there are young volcanic rocks on the crests of African swells is best seen north of the equator (Figure 6) and work in that region over the past 70 years has only served to confirm Krenkel's (1922) observation. There are also many continental volcanoes in low-lying areas, some of which have been shown to overlie prominent basement elevations as, for example, at Dakar (Figure 13) and Cameroon Mountain (Figure 16). This confirms the observation from other areas of the African Plate that the elevation process involves regional upward flexure and not just volcanic construction.

The Chad Basin (Figures 6 and 20) is perhaps the best example of the way in which African basins simply represent depressions complementary to surrounding, commonly volcano-capped, swells. Eleven discrete volcano-crested, elevated areas form an ellipse about 2000 km across around the Chad Basin (Figure 20) with the basin occupying the low-lying area within the ellipse. The basin is centred over Mesozoic rifts (Genik, 1993) and perhaps the modified lithosphere of the rifted area has played a part in localizing its position (Burke, 1976b). All African continental-interior basins that have formed within the past 30 My, including those in areas among swells that have no volcanoes on their crests such as the Zaire and Kalahari Basins, have formed in the same way. Argand (1924, pp. 151-153, and his figure 6 shown here as Figure 2) characteristically attributed the large scale basin and swell structure of Africa to compression. He suggested that the basins and swells represented "plis de fonds." Krenkel (1922; 1957) had more correctly related the structure to vertical uplift involving partial melting of the mantle, a process that he termed *magmarsis* (Krenkel, 1957, p. 427).

Kennedy's (1965) observation that the 30 Ma volcanism of Africa is confined to areas of Pan-African reactivation (Figure 21) has been interpreted in two ways. Ashwal & Burke (1989) attributed the concentration of volcanic rock to the existence of fertile mantle lithosphere beneath the Pan-African areas and suggested that volcanic material could only be extracted from that source (Figure 22). An alternative interpretation suggests simply that the lithospheric mantle beneath the cratons is thicker and mechanically too strong to rupture so that rising plumes have no path by which to reach the surface. These two solutions are not incompatible, and both could be right.

It is intriguing to consider that if the African Plate is at rest over the mantle circulation pattern, then the distribution of higher areas and of young volcanic rocks might reveal something about the pattern of convective circulation where it impinges on the bottom of the lithosphere. Perhaps the topography of the African Plate mimics, mainly because of isostatic response, the planform of the impingement of the mantle circulation on the bottom of the stationary boundary conduction layer that is the African lithosphere. A simple assumption might be that high areas and volcanic areas lie over rising, lower density, convective plumes and that lower areas lie in the places where diffuse downward convection is taking place. R. Thiessen *et al.* (1979) were unable to resist the temptation to attempt an analysis for the continental area of Africa based on this simple assumption. They used the method of A. Thiessen (1911) polygons and their map and histograms (Figures 23 & 24) show some concentration of separation among both hot spots and "high spots" (crests without volcanoes) of about 700 km. Because this is about the depth to the bottom of the transition zone at 670 km, it is tempting to consider that an upper mantle circulation with comparable length and height scales might be revealing itself. HIMU dominates African hot spot magma sources (Hart *et al.*, 1992; Hay & Wendlandt, 1995; Volker & Altherr, 1995) and has been suggested to represent a small degree of partial melting of typical MORB source type upper mantle from above the base of the transition zone. Alternatively, the spread of separations between hot spots ranges between 200 and 2000 km which might be better interpreted simply as an indication of complex structure in the convecting mantle. A similar study for the ocean floor was not feasible when the Thiessen *et al.* (1979) study was undertaken, but now better data and better models of sea-floor elevation make such a study feasible (Nyblade & Robinson, 1994).

Magmatism in the East African Rift System is considered later when the rift system is discussed. The observations that African magmatism over the past 30 My (1) has been widespread, (2) represents a small degree of partial melting of the mantle and (3) is HIMU source dominated, are important in considering the nature of rift magmatism.

Continental surface

Introduction

The high mean elevation of the African continental surface (Cogley, 1987) and the continent's distinct hypsography (Bond, 1979; Figure 25) are largely features of the African Superswell (Nyblade & Robinson, 1994), but the many lesser swells of the continent (Figure 6) contribute to both. The Kenyan, Ethiopian and Red Sea domes of the Superswell contain the most prominent parts of the active East African Rift System suggesting that rift and Superswell share a common origin. Local additional elevation related to the active rifts contributes further to regional elevation. The Superswell (Figure 6) extends across the South Atlantic Ocean in a northeasterly direction from the Mid-Atlantic Ridge between Tristan and Bouvet. On the continent, it includes the great elliptical topographic swells of southern Africa, East Africa, Ethiopia and the Red Sea. These continental areas form "High Africa" (Cloos, 1937; L. King, 1962; 1963) in contrast to "Low Africa" of the north and west of the continent. In "Low Africa" there are numerous topographic swells which are similarly elliptical in shape but are much smaller in both length and height (Figure 6). Because of the similarity in shape and age of formation of both the larger and smaller swells, I suggest that they have originated in similar ways.

Comparable swells are not familiar to me anywhere else in the world, although some large-scale features of the surfaces of both Mars and Venus look to me reminiscent of those of Africa. Because Venus and Mars are both now planets without plate tectonics, there may be a resemblance in their tectonic styles to the current behavior of the African Plate which is at rest over the mantle convective pattern. Africa is behaving in the same way as an entire planet that has no relative motion between its lithosphere and underlying asthenospheric convection pattern. These objects have been called "one plate planets" (Solomon, 1978).

On Earth, the swell structure of the African Plate is unique. I have been able to demonstrate that the swells were not present during early Cenozoic times (Figure 3) so my suggestion is that elevation of the surfaces of all the topographic swells began about 30 million years ago and, like the simultaneous volcanism and the rifting, is consistent with Africa having come to rest over the underlying circulating mantle at that time. The swells are isostatic responses to the presence of mass deficient volumes, possibly huge plumes, in the underlying convecting mantle. These objects need not necessarily be in contact with the base of the African lithosphere but could lie deeper. Estimating a depth to the source of a mass deficiency might be attempted by characterizing the length over which flexure at the edge of a swell extends.

Figure 26 illustrates a simple conceptual model for southern Africa where there is little Pan-African reactivated lithosphere and, therefore, little volcanism. A generally low-lying continental area, perhaps similar to Australia today, was covered by a deeply weathered surface that formed between about 65 and 30 Ma after the extensive Late Cretaceous seas had retreated from the continent. Since the continent came to rest over the mantle circulation, that surface has been elevated and is now preserved on the top of the the great southern African element of the Superswell. Escarpments, including the Great Escarpment, have formed by erosion during those 30 million years. In this interpretation, the Great Escarpment is not an old feature generated at the time of the breakup of Gondwana at about 150 Ma as L. King (1962; 1963) and some others have concluded.

I do not wish to suggest that I am the first to consider the elevation of Africa and the formation of the Great Escarpment to be a quite recent phenomenon. Perhaps an anecdote will help to make that clear. Sidney Hollingworth, the distinguished British geomorphologist, spent three months as a guest of Lester King at the University of Natal sometime in the early 1960s. On his return to England Hollingworth greeted me, as one of his former students who was active in African geology, by saying "It's just gone up hasn't it?" I had no idea what he was talking about so I replied "What?" His response was one word: "Africa." That brief conversation certainly helped to stimulate my interest in African geomorphology and made me reread Lester King more assiduously.

Many features of the continental surface help to throw light on what has happened in Africa over the past 30 million years. Interest focuses naturally on the extensive erosional surfaces which, when compared with those of other continents, are surely distinct if not uniquely well-developed features of Africa. Linked to these surfaces are the scarps that separate them, especially those escarpments of regional extent. Deep weathering on the erosion surfaces, river gravels and depositional veneers such as pediment gravels, which lie on top of parts of some surfaces, all contribute to the understanding of what has happened. Terraces, both fluvial and marine, elevated marine sedimentary deposits, and the integration of observations on land with those offshore also help the understanding of the evolution of the continental surface over the past 30 My. Many of these features record the more recent environmental developments, particularly those rapid fluctuations of the past 3 million years which have accompanied northern hemisphere glaciation.

There is now a renewed interest in these matters partly because of stimulus from the exploration for, and exploitation of, alluvial and offshore diamond deposits and partly because of the advent of new techniques such as fission track dating and cosmogenic nuclide dating of surfaces. Fission track data for erosion surfaces in Africa are beginning to accumulate (Foster & Gleadow, 1996; Summerfield, 1996; Smith, 1994). There is evidence that only relatively small thicknesses (<2 km) of rock have been removed from most areas where surfaces have been eroded during the past 30 My. For that reason, and given the gentle geothermal gradients that obtain, it may not prove feasible using fission track methods to distinguish more recent events from those that took place during the breakup of Gondwana.

Improved capabilities in computational data handling and numerical methods are making it easier to model the integrated erosional and depositional processes which are especially important near the continental margins (Figure 26). It has become readily possible to incorporate isostatic considerations into model studies including the effects of flexure of the continental margins in response to both the addition of loads such as continental margin sedimentary deposits, and the removal of loads by the erosion, a process which generates a buoyant response (Gilchrist *et al.*, 1994; McGinnis *et al.*, 1993).

The development of surfaces shows a great deal of both local and regional variation in Africa but three coherent areas can be distinguished: (1) Southern Africa where surfaces of regional extent are developed mainly on a single, very large and unrifted swell with no volcanic cap. This southern African swell contains perhaps as many as ten areas of locally greater uplift which I will refer to as subswells. Although there are no young active rifts in this area, older (250-150 Ma) rifts are being reactivated in the Zambesi and Limpopo valleys. (2) Western and northwestern Africa, where basins such as the Zaire and Chad Basin lie among smaller swells. In that region, erosion surfaces, which are confined to individual swells, are less extensive and the escarpments are less prominent than in southern Africa. Several of the individual uplifts in western and northwestern Africa are capped by volcanoes. Rift development in that area is restricted to reactivation of the Cretaceous Benue Rift and mainly takes the form of localized hot spot igneous activity. (3) East Africa and much of the northeastern continent where developments on the East African, Ethiopian and Red Sea swells are closely linked to the active East African Rift System.

Although the three regions are distinct, all are experiencing swell uplift, erosion and scarp formation. Evidence of when these processes began is better in East Africa and in North and West Africa where the occurrence of 30 Ma and younger volcanic rocks and rifts on the crests of swells serves to roughly date the uplifts. The absence of young volcanic rocks and young rifts in southern Africa has made the dating of swell uplift in that area harder.

Perhaps the strongest evidence for timing of African continental uplift comes from the offshore occurrence of the prominent Oligocene unconformity (~34-22 Ma) seen in seismic reflection data all around the continental margin. Although this feature varies in intensity regionally, its ubiquity and its prominence compared with the roughly contemporary global Oligocene unconformity are a strong indication of a general elevation of the continent. Although continental elevation has been general, regional differences warrant separate discussion of the three distinct areas: (1) southern Africa; (2) western and northwestern Africa; and (3) northeastern Africa.

Southern Africa

Southern Africa extends roughly over the countries of South Africa, Namibia, Mozambique, Lesotho, Swaziland, Botswana, Zimbabwe, Zambia, southern Angola and part of Zaire. Interest in the escarpments and erosion surfaces of southern Africa goes back for more than one hundred years. The history of this research has been discussed by Partridge & Maud (1987) who provide a helpful appendix in which highlights are outlined. Here I only note that Alex L. du Toit was himself a student of the erosion surfaces (du Toit, 1910; 1933; Figure 27) emphasizing tectonically influenced river capture and Plio-Quaternary tectonism in the form of axes of uplift in several of the discrete areas within the Great Swell of southern Africa that are here called subswells.

Lester King (1962; 1963), whose masterly studies continue to make stimulating reading, interpreted the erosion surfaces of Africa in the broadest perspective and finally extended his monumental treatment to the whole earth. In South Africa, King's mantle has fallen onto the shoulders of Partridge & Maud who in 1987 published a comprehensive review of the geomorphic evolution of South Africa. Partridge & Maud (1987) supplemented King's approach with detailed field studies and paid special attention to such features as deeply weathered surfaces and correlating coastal sedimentary deposits with inland surfaces. They radically revised King's interpretation by showing persuasively that there are no pre-Mesozoic (>250 Ma) surfaces on the continent, except where exhumed from beneath sedimentary piles. Partridge & Maud (1987) identified an "African Surface" using the same term as Lester King (1962; 1963) but with a somewhat different meaning. They suggested that the African Surface was formed as a result of complex cycles of erosion during the long interval between the Gondwana breakup rifting event in southern Africa (~150 Ma) and the Early Miocene (~20 Ma).

I suggest that South Africa, like much of the rest of the continent, was probably at least partly flooded by the sea at some time during the Late Cretaceous. Sahagian (1988) did not depict the sea as flooding South Africa because of the absence of marine Cretaceous sediments from the interior of the continent. I attribute this absence to erosion, associated with great uplift during the past 30 Ma, and infer that if much of the rest of the continent was flooded so was much of southern Africa. I would certainly concur with Partridge & Maud (1987) that their African Surface is a product of complex cycles of erosion since the breakup of Gondwana. There may well have been, for example, some regional elevation of southern Africa during the time of intense kimberlite eruption between ~90 and ~70 Ma. Possible evidence of erosion from this source can be seen in the relatively thick Upper Cretaceous coastal and offshore section in Mozambique which may represent sediment brought to the coast along the Limpopo valley. (De Buyl & Flores, 1986). The Upper Cretaceous section at the mouth of the Orange river (Miller, 1995, his figure 1) may also represent erosion from that source.

The African Surface of Partridge & Maud (1987) is dominantly the same surface that is so widely represented over much of the continent (Figure 3). That surface, which mainly developed between the end of the Cretaceous and 30 Ma, is characterized by deep weathering. Partridge & Maud (1987, pp. 189 - 190) identified half a dozen localities close to the coast of South Africa where deeply weathered horizons within sediments can be bracketed in age as having formed sometime between the end of the Cretaceous and the end of the Eocene (65 - 34 Ma). These weathered rocks are not themselves expressions of the "African Surface" so much as horizons of weathering within the Paleogene sedimentary sequence. Correlation in age is suggested between the weathered sedimentary horizons and the weathering recorded on the exposed African surface.

In the absence of datable rocks on top of the African Surface, this is the only attainable correlation. An inherent and unavoidable weakness is that the development and preservation of a weathered rock surface depends on many variables besides the timing of a particular weathering episode.

I have suggested that uplift of the continent-wide surface, which I would correlate with the "African Surface" of Partridge & Maud (1987), began during the Oligocene at about 30 Ma rather than in the Early Miocene about 20 Ma. The latter is the date suggested by Partridge & Maud (1987) for cessation of development of their African Surface. I am not sure that the difference of 10 My is very significant. Once a surface begins to be elevated, the evidence of uplift is concentrated in scarps and river valleys close to the coast. It takes a long time for these elements to propagate back into the rising surface and the most elevated regions show few signs of change, even today.

In a way, the African Surface is still developing so it is hard to say when its history was interrrupted. The important thing is to date the beginning of uplift, and that is not easily resolved. One of the reasons for a difference on my part in estimating the onset of uplift may be that I am influenced by the beginning of volcanism in parts of the African Plate remote from South Africa, as well as the age of the spectacular unconformity offshore. I envisage the continent and much of the rest of the African Plate beginning to rise, and the "African Surface" beginning to be cut down and back, at the time when volcanism broke out over much of the African Plate and when the offshore unconformity formed. Partridge & Maud, who concentrate on South Africa, have no volcanism to influence them and date the demise of the surface from about the time when renewed sedimentation begins, in many areas, above the unconformity. I do not see this as a major difference between our interpretations of the erosional history of southern Africa. I have a much bigger problem with interpretation of the age of the Great Escarpment.

The long-standing idea that the Great Escarpment formed during rifting associated with the breakup of Gondwanaland persists in the analysis of Partridge & Maud (1987), but this seems to me to be at least questionable. The Great Escarpment (Partridge & Maud, 1987, their figure 12) extends all around the Republic of South Africa and reaches as far north as Luanda on the west coast of Africa and the Zambesi valley on the east coast. The Escarpment bounds the Great Swell of southern Africa to which it is in my view genetically linked. One reason why the Great Escarpment has been associated by many workers with the breakup of Gondwanaland is that the Great Escarpment is parallel to the rifted continental margin. Because the rifted margin formed at about 150 Ma, the Great Escarpment has been inferred to have formed at the same time. The spatial association does not seem to me a strong argument for a close association in time of formation between escarpment and rifted margin. The entire coastline of Africa was rifted during the breakup of Gondwanaland but the Great Escarpment is developed only along a relatively short length of that coastline. Moreover, there are no comparable Great Escarpments along the conjugate rifted margins or other rifted margins of comparable age on the planet.

There are other difficulties in assigning an old age to the Great Escarpment. The Escarpment lies ". . .more than 200 km inland of the original continental margin in some areas. . ." (Partridge & Maud, 1987, p. 188). If an age of about 150 Ma is assigned to the rifting of the continental margin, then horizontal scarp retreat would have averaged only a little over 1 mm per year for the past 150 million years. This is a slow rate. About 90 mm of rain fell in Durban on the day I delivered this du Toit memorial lecture during April, 1995. It is hard to envisage the Great Escarpment of the Drakensberg lasting so long and retreating so little under such an attack.

The Lebombo monocline, which lies just beneath the escarpment in Natal and Mozambique, is one of only three such monoclinal flexures that outcrop at the continental rifted margins of the world. Elsewhere similar margins have been inferred to exist from geophysical observations, but are deeply buried under the sediments deposited at the continental margin. Evidence of Late Neogene uplift is strong in the two other areas that expose similar monoclines. The Panvel flexure, among the Deccan traps south of Bombay, has been shown to be young from the upward tilting toward the coast of seismic horizons older than 10 Ma. These horizons have been identified on the offshore regional seismic lines of the Indian Oil and Gas Commission. Elevation of East Greenland at about 10 Ma has exposed a comparable monoclinal flexure along the coast near Kangerdluqssuaq. I suggest that exposure of the Lebombo monocline is similarly related to relatively young uplift associated with the elevation of the Great Swell of southern Africa.

A geophysical model has been constructed (Gilchrist *et al.*, 1994) showing how the Great Escarpment in southwestern Africa could have survived as an old geomorphological feature, but I am not sure whether this is particularly strong evidence that the feature is actually old. Indeed, so great is my faith in this kind of geophysical modelling that I have little doubt that an equally presentable model treating the escarpment as a young feature could also be constructed.

Figure 26 shows a pair of generalized sketch cross-sections across southern Africa to illustrate how I envisage the processes of uplift, erosion and offshore deposition operating over the past ~40 Ma. The upper section (Figure 26a) represents conditions while the African Surface was still forming. The continent was low-lying with some residual hills but there was no Great Escarpment. Sediments were being deposited at the continental margin amongst which were some on both east and west coasts that preserve evidence of deep weathering.

Uplift began at about 30 Ma and the African Surface was lifted higher on the eastern side of the continent than on the western side. Although the elevation has the gross form of a single Great Swell, subswells are superimposed and these, like the main swell itself, may have continued to rise through much of the past 30 My. If there has been such continuing elevation, it is hard to tell whether the process was intermittent or episodic. Evidence of river capture and Quaternary tectonic axes (du Toit, 1910; 1933) indicates possible episodicity. The Great Escarpment has developed all around the coastal margin of the uplifted continent and has continued to retreat through the past 30 My. Erosion surfaces below the Great Escarpment are young. There are residual hills on these younger surfaces below the Great Escarpment just as there are on the African Surface itself. The occurrence of residual hills rising from a surface does not serve to date the surface. The residual hills on the younger surfaces are outlying eroded remnants of the material capped farther inland by the African Surface. Partridge & Maud (1987, their figure 12) show about 20% of the total area of their African Surface as lying below the Great Escarpment. They explain how a surface of the same age, the African Surface, was preserved both above and below the Great Escarpment. I suggest that consideration be given to the idea that those areas assigned to the African Surface below the Great Escarpment are in fact younger surfaces. My conclusion is a reversion to the interpretation of De Swardt and Bennet (1974) as well as to Hollingworth's unpublished suggestion. Obviously dating surfaces is difficult. Criteria used to define surfaces, for instance the occurrence on the surface of both residual hills and deep weathering, are particularly hard to assess. My simple sketch is not meant to indicate a rejection of the very detailed research of many workers on the erosion surfaces of southern Africa. I do not, for example, show the sub-Karroo unconformity which has been shown to have been exhumed and which controls the present surface (de Wit, 1995) in several areas below and close to the Great Escarpment.

Another topic that I do not address is the division by Partridge & Maud (1987, table 1, p. 187) of the post African surface erosional history of South Africa. They interpret their detailed work as showing that the present elevation is the result of three separate episodes of uplift. The first, in the Early Miocene, was relatively small at 150 to 300 m but the next, which took place as recently as 2.5 Ma, reached the astonishing local maximum uplift of 900 m. A third and ongoing episode can be distinguished from the second along river valleys and in the eastern Lowveld. I do not know of any work elsewhere in Africa which has allowed uplift history to be divided in such detail. Neither am I aware of anywhere remote from plate boundary zones and volcanic regions where uplift rates as high as 900 m in a few million years have been estimated. This observation suggests to me the existence of a very unusual phenomenon and a challenging opportunity for further research.

Offshore deposition has resulted in new sediment wedges above the 30 Ma unconformity as sketched in Figure 26b. Rust and Summerfield (1990) have assembled well data from offshore southwestern Africa to show that deposition was greater in that area during the Paleogene (65-34 Ma) than in the Neogene (since 34 Ma) and that sediment deposition was fastest around southern Africa during the Cretaceous when the rifted margins were young. Interpreting this kind of information fully requires abundant seismic reflection data to complement the well data. I suspect that sufficient seismic reflection data have been acquired around the shores of South Africa and that sufficient wells have been drilled to allow calibration of that seismic data in enough detail to permit the compilation of maps of depositional surfaces for closely spaced time intervals from the Early Cretaceous to the Pleistocene (Brown *et al.*, 1996). Only when such maps are publicly available will it be possible to fully relate the offshore depositional history to that of the continent. Meanwhile, accepting Rust and Summerfield's (1990) observation that relatively little sediment has accumulated off southwestern Africa in the past 30 My, I suggest that the onset of desert conditions in much of southwestern Africa began when East Antarctica first became glaciated and the Benguela current began to flow by about 34 Ma (Siesser, 1978). Desert climate in the source areas may have been responsible for the limited Neogene sedimentary deposition offshore. This is a contrast to the great abundance of post 30 Ma sediment deposition, especially in deeper water, off much of the rest of Africa.

Figure 26b also indicates places of local isostatic response, probably with wavelengths of 100 to 200 km, to changing surface loads (McGinnis *et al.*, 1993). These effects have been superimposed on the general, and perhaps continuing or episodic, elevation of the Superswell as well as any local smaller scale subswell effects. Eustatic changes in sea level along the coast, which have been dominated by ice volume variations throughout the past 30 My, have interacted with the tectonic, erosional and isostatic influences. The combination of these factors that has led to local exposure near the coast of older Cenozoic sediments such as those showing signs of deep weathering in places like Luderitz, Need's Camp, Zululand coastal plain, Birbury and Pato's Kop (Partridge & Maud, 1987).

River development in Southern Africa

River development in southern Africa has been studied by many workers and has been analysed as thoroughly as the related erosion surfaces, shoreline terraces and sediments. I have made use of various published sources, especially Dardis *et al.* (1988) and an as yet unpublished thesis (de Wit, 1993; see also de Wit, 1995), to illustrate the way in which I envisage the drainage and the related environment of southern Africa as changing with the development of the Superswell over the past 30 My. In Figure 28a, I have depicted conditions before the great current episode of uplift, relying mainly on figure 3.11 of Dardis *et al.* (1988) in which ideas about ancient drainage systems from at least seven major studies have been incorporated. I have also used the results of de Wit's (1993) study of an area in northern Cape Province. I have drawn the coastline of the continent as it is today except near the mouths of the Zambesi and Limpopo where there is evidence of the existence at this time of a carbonate bank and where there has been great progradation during the past 30 My (Kolla *et al.*, 1991; Mougenot *et al.*, 1986; De Buyl & Flores, 1986). The map is consistent with the idea of a low-lying continent, deeply weathered (Partridge & Maud, 1987) and drained by short rivers with the exception of a dominant system made up of the Trans Tswana and the Kalahari rivers. That river system reached the ocean in the general area of the Orange river delta where Miller (1995) reported thick Cretaceous and Tertiary post-rift accumulations of sediment.

Conditions at the onset of the elevation of the Superswell, which now averages 500 m in oceanic areas and 500 m in continental areas (Nyblade & Robinson, 1994), are shown in Figure 28b. The worldwide lowstand of the sea at about 34 Ma, which is attributable to the onset of Antarctic glaciation, extended the southern African shoreline seaward and the beginning of the uplift of the continent provided sediments to erode submarine canyons. The locations of some of these canyons are indicated but there must have been many more. Following de Wit (1993), I have shown the Karoo and Kalahari rivers as continuing to flow. The Trans Tswana river is shown as shut off by the start of subsidence in the Kalahari Basin. Although Cretaceous sediments locally underlie the Neogene deposits in that basin, I do not interpret that observation as indicating continuous basin subsidence since Cretaceous time in the Kalahari. Estimating the start, or the renewal, of the subsidence of the Kalahari Basin at about 30 Ma has led me to indicate that some of the tectonic axes which mark the subswells within the Great Southern African Swell may have been active by that time. The Kalahari Basin surface is no lower than 500 m above sea level and lies entirely within the Great Swell of southern Africa. The Chad and Zaire Basins farther north in Africa, which are comparable in extent, are a few hundred metres lower.

The best documented evidence for activity on the axes of the subswells is of Quaternary age (du Toit, 1910; 1933; Mayer 1973), but the elevation of some of the subswells is so great that consideration should be given to the idea that uplift has taken place on at least some of the axes over longer intervals during the past 30 My. I have followed Siesser (1978) in showing the Benguela current, and also the onset of aridity in southwestern Africa (Van Zinderen Bakker, 1976), as starting with the onset of Antarctic glaciation about 34 Ma. It is close to this time, about 30 Ma, that I consider that the Great Escarpment began to form by retreat from the continental margin.

In Figure 28c, I have sketched a possible version of the southern African environment at about 15 Ma during the Middle Miocene. I have followed de Wit (1993) in showing the Vaal and Orange rivers as important at that time and I have indicated a possible general location for the Great Escarpment, intermediate between the coast and its present position. The fault beneath the Makarikari Pan may have been active at that time, and is certainly active now (Scholz *et al.*, 1976).

In summary, Surface development in southern Africa over the past 30 My has been dominated by the uplift of the Great Swell of southern Africa, one of five major elements of the African Superswell (Nyblade & Robinson, 1994). This elevation has been accompanied by the development and retreat of the Great Escarpment. Subswells within the Great Swell, some of which are comparable in size to the individual swells in the north and west of the continent, have developed during the past 30 My and appear, mainly from evidence of river capture, to be still active or at least to have been active within the past million years. Drainage system evolution has responded to the development of the Great Swell. The Orange system is not as significant a transporter of sediment as was its ancestor, probably because of the onset of aridity in southwestern Africa at the beginning of the Oligocene at about 34 Ma. The Limpopo and especially the Zambesi have become the dominant sediment transporters of southern Africa, building huge sediment accumulations over low-lying areas near their mouths.

I have not attempted to discuss the later history of southern African river systems with their well-developed and dated terraces (Partridge & Maud, 1987) and neither have I addressed the elevated marine sediments and terraces of the southern African coast (Dingle *et al.*, 1983; Davies, 1972). The vertical distribution of the elevated indicators of former shorelines appears to reflect complex interplay of tectonic and ice-volume controlled sea level changes. Southern African surface development over the past 30 My has contrasted in some ways with developments in the northwest and the northeast of the continent, but the basic process of basin and swell formation seems generally similar, although there is an obvious difference of scale.

Western and Northwestern Africa

This area includes all of continental Africa south of the Atlas Mountains, which form part of the Alpine mountain belt, except for those regions that form parts of the Superswell. The huge region consists of numerous swells clustered around the Zaire and Chad Basins and the Western Sahara Double Basin which is made up of the Taoudeni Basin in the north and the basinal area around the inland delta of the Niger in the south (Figures 6 and 20). Individual swells are elliptical in map view with semi-axes ranging from as much as 1000 km to as little as 100 km. Many of the swells north of the Zaire Basin, which lie on Pan-African reactivated crust, are capped by young volcanoes.

The most prominent erosion surfaces rise gently from the continental margins and from the edges of the interior basins. In the Chad Basin, this surface can be interpreted as underlying sediment accumulations of Late Cenozoic age. Occurrences of bauxite and thick laterite in West Africa are concentrated on a higher surface which generally lies above a prominent escarpment and at a range of elevations (Figure 29). Because of the occurrence of bauxitic material among Eocene sediments of the Senegal Basin, this surface has been considered to be of Early Cenozoic age. The similarity in age, bracketed between Late Cretaceous and about 30 Ma, leads me to suggest that the bauxite-bearing surface of West Africa is the equivalent of the surface called the African Surface in southern Africa and the widespread lateritic surface which lies beneath basalt in areas around the Afar. The distribution of this bauxite-bearing surface in West Africa indicates the distribution of those areas that have been elevated on discrete swells (Figure 29).

In contrast to many areas in southern and eastern Africa, where great uplift has led to the removal of much of the sedimentary cover, some prominent escarpments of northern and western Africa are at the edges of outcrops of gently warped sedimentary deposits of various ages. It has been the development of the basin and swell structure of Africa that has generated the gentle warping forming the "plis de fonds" of the continent to which Argand drew attention (Argand, 1924, his figure 6). Where escarpments have formed on Cretaceous sandstones, as in both southeastern Nigeria and southwestern Nigeria (Figure 29), a sub-Cretaceous erosion surface has been exhumed. A much older surface has been exhumed from beneath the prominent escarpment that surrounds the outcrop of the nearly horizontal older Paleozoic sedimentary rocks of West Africa (Figure 29). In the Sahara, successive escarpments of Early Paleozoic sandstones are prominent around the Ahaggar and the Tibesti (Figures 29 and 30).

Escarpments within the crystalline rock outcrops of western and northwestern Africa appear to be in some cases localized by lithological contrast. For example, the Jos Plateau (Figures 15 and 29), which is bounded by an escarpment, does not lie symmetrically along the crest of its swell but bounds the outcrop area of a group of Jurassic A-type granites (Whiteman, 1982, his figure 9). A similar group of Devonian granites in the Air (Figure 29) is also bounded by an erosional escarpment.

All of these escarpments have formed by erosion during the uplift of the many individual swells of western and northwestern Africa. The evidence for the timing of the uplift of the swells comes from the oldest age of the volcanic rocks on their crests, which extends back to 30 Ma, and also from the occurrence of the great offshore unconformity around the coast of West Africa. The start of rapid progradation of the Niger delta (Doust & Omatsola, 1990) records the same process.

In Cameroon, where intrusions indicate igneous activity and presumably the existence of high ground from 65 to 30 Ma (Lee *et al.*, 1994), there may have been a more complicated topographic history. Unfortunately, because the area has been the site of so much young topographic development (Figures 6, 16 and 29), it has not been possible to separate evidence of a distinct earlier phase.

River Systems of Western and Northwestern Africa

River development in northern and western Africa has been radically perturbed by the onset of Saharan aridity about 3 Ma (deMenocal, 1995) but it is still possible to discern something about the river systems that developed with the establishment of the Basin and Swell Structure about 30 Ma.

Before 30 Ma, West African drainage appears from the sedimentary record to have been dominated by an ancestral Benue river flowing along the Benue Rift that had formed in the Early

Cretaceous (Whiteman, 1982). Drainage into the Senegal Basin may have also been along ancient structures like the Casamance Rift (Burke, 1976a). Drainage of northwestern Africa during the interval before both the formation of the Atlas Mountains and the establishment of the Basin and Swell Structure of Africa is obscure but may have been dominantly toward the Mediterranean. Drainage of the huge area now occupied by the Zaire Basin may have involved numerous rivers similar to the present Oguoe. Sediments deposited from such rivers have contributed to the Later Cretaceous and Earlier Cenozoic clastic sequences in the Gabon and Cuanza Basins (Brognon & Verrier, 1966).

With the establishment of the Basin and Swell Structure, internal drainage may have been as depicted in Figure 30 in the Chad Basin, the Zaire Basin, and the Niger Basin with deposition close to the site of what is now the inland delta of the Niger (Reclus, 1888). When and how the internal drainage of these great basins first cut through to the ocean is not clear. The Zaire has been captured by a short river which has also cut the Zaire submarine canyon. This may have happened relatively soon after uplift began along a north-trending swell close to the coast (cf. Summerfield, 1996), as shown in Figure 30, because little non-marine sediment has been deposited in the Zaire Basin over the past 30 My. The Chad Basin has a thicker but poorly dated section of a few hundred meters. Chad Basin internal drainage has been integrated with that of the Benue episodically. Lake "Megachad" (Figure 1) has drained to the ocean through a natural spillway at Bongor during wet Quaternary intervals (Grove & Warren, 1968). How long ago this first started to happen is unknown.

Drainage during the episodes of throughgoing rivers was from the southwestern slopes of the Ennedi (Figure 30) into Lake Megachad and thence over the Bongor spillway to join the Benue river system and reach the ocean in the Niger delta. That system extended over 2,500 km and must have represented one of Africa's largest catchments. An attempt has been made using misinterpretation of space-borne radar images to suggest that this drainage was at one time even more extensive and was linked to rivers draining the Red Sea Hills. This frequently reiterated suggestion has been shown to be wrong (Burke & Wells, 1989).

Niger drainage, like that of the Chad Basin, is integrated with drainage to the ocean during wetter Quaternary times. At present, the Niger crosses a spillway at Taoussa and joins the Benue before that river reaches the Niger delta. During drier times the Niger does not flow beyond the area of the inland delta.

In summary, Surface features of western and northwestern Africa are dominated by numerous relatively small-scale swells and larger intervening basins. The preservation of extensive areas of flat-lying Phanerozoic sedimentary rocks has led to the common occurrence of cuestas which are absent from much of the rest of the continent where elevation has brought crystalline rocks to the surface. The development of internal and throughgoing drainage in the major basins has fluctuated greatly with Quaternary climatic change. Before 30 Ma, the Benue was the dominant river system of the region, but now that role is shared by the Zaire and the Niger/Benue system.

The Niger/Benue system has experienced changes related to Quaternary climatic change which resemble those displayed by the Nile, and this similarity in the style of the major rivers provides a link between surface developments in western and eastern Africa. Eastern Africa contrasts greatly with both southern Africa and western and northwestern Africa because over the past 30 My, it has been overwhelmingly dominated by the development of the active East African Rift System. In addition, during the past 3 million years, the Sahara has experienced climatic fluctuations linked to northern hemisphere glacial changes which have been superimposed on the rift-dominated structure of northeastern Africa.

Northeastern Africa

The discrete 30 Ma or possibly slightly older swells of Jebel Uweinat and Jebel Marra to the west and the Red Sea and Ethiopian swells, which form the northernmost part of the African Superswell to the east (Figure 6), embrace the north trending, broad, low-lying area occupied today by the Nile river system. Rivers that have occupied this area have a history of intermittent throughgoing flow with interior drainage at times stopping in the Sudd. This history extends well back into Cenozoic times (Berry & Whiteman, 1968; Adamson *et al.*, 1980; Adamson, 1982). A source of confusion is that the older rivers of this system are not customarily referred to as the Nile. The earliest river to which the name Nile is commonly given is the Eonile (Said, 1993, p. 5) which cut a giant canyon through the Nile valley during Messinian times (~6 Ma) when drainage was adjusted to a base level on the dessicated floor of the Mediterranean (Hsü, 1972).

The timing of first development of a Nile drainage system, although not a river to which the name Nile is given, is bracketed by two observations: (1) No great river was reaching the north coast of Africa in the area of the present Nile valley during deposition of the Late Eocene and Early Oligocene richly fossiliferous sediments of Fayum (Kappelmann *et al.*, 1992; Said, 1993, figure 1.15) and (2) A great delta was occupying an area two hundred km south of the present Nile delta by the Early Miocene (Said, 1993, figures 1.15 and 1.16; Elzarka & Radwan, 1986).

Taken together, these observations show that the Nile drainage first developed about 30 Ma, but two considerations have helped to obscure the recognition that there has been an approximately 30 million year history of the Nile. One has been a widely promoted suggestion, based on a misinterpretation of space-borne radar observations, that rivers from the Red Sea hills formerly extended "in Mid-Tertiary times" across Sudan to join the Chad Basin, Benue and Niger river systems reaching the Gulf of Guinea in the Niger delta (McCauley *et al.*, 1982). The error of this interpretation has been exposed by pointing out that both the timing of uplift of the Red Sea hills and the volumes of sediment in the Nile and Niger deltas are not consistent with the Trans-African Drainage System hypothesis (Burke & Wells, 1989). The second source of confusion has been that students of Nile history have commonly extended their analysis back no further than the Late Neogene (Said, 1993). This has perhaps been because of the extraordinary richness of the record of Nile depositional, erosional, climatic, faunal and cultural history over the past five million years. Nevertheless it is important to recognize the much longer history of rivers occupying the basin now occupied by the Nile system as typical products of the development of the basin and swell structure of Africa.

The Nile today rises from lakes of the East African Rift System at the equator. This source has probably been only intermittently linked to the great river. Over much of the past 30 million years the Ethiopian Highlands, a product of uplift and volcanism related to the Afar plume, are more likely to have provided the main headwaters of the Nile system. To see why this has been the case, it is necessary to attempt some analysis of the history of the East African Rift System. Questions about that great rift system that are important for understanding the evolution of the African Plate over the past 30 million years include: When did the present episode of rift activity begin? How are the different parts of the rift system linked to each other? How is the rift related to the Superswell? Is the volcanism of the rift system comparable to the volcanism that has erupted elsewhere on the African Plate over the past 30 million years? How is the development of the East African Rift System linked to the reconstruction of the Indian Plate and to the formation of the Central Indian and Carlsberg Ridges? Do the Red Sea and the Gulf of Aden constitute part of the East African Rift System? These are all topics that have been the subject of much research in recent years and I hope that the following attempt to extract coherent answers from my reading of the results of that research does not do injustice to the efforts of the researchers either from my ignorance or from misunderstanding on my part.

East African Rift System

Overview

The East African Rift System has long been one of the most distinctive and well studied features of the African Plate (Suess, 1891), but I have found it difficult to lay hands on comprehensive analyses that address the history of the whole system over the past 30 My. One reason for this deficiency is that the system is so large and so diverse that it may be unrealistic to attempt to treat it as an entity. I am certainly conscious of that consideration. The geology of the submarine part of the system off the coast of Southern Tanzania and Mozambique is not much like that of the great lacustrine basins of the western rift which themselves differ from the three discrete sectors of the eastern rift in Tanzania, Kenya and Ethiopia. The Red Sea is different again, as is the Gulf of Aden, and finally the rift of the Gulf of Suez is quite distinct.

These rifted areas have two things in common which make me consider them all to be parts of the East African Rift System: First, they are all regions of lithospheric extension involving normal faulting which is restricted to widths of at most a few hundred kilometres, and secondly, extension has taken place in all these areas over at least some part of the past approximately 30 million years.

Rift segments with distinct azimuths typically extend for distances from 100 to 300 km, but the Red Sea and Gulf of Aden are much longer. The rift segments are in some places joined to neighboring segments that trend with different azimuths. These joins have been shown in places in the western rift to be at accomodation zones (Rosendahl, 1987). In other places, rift segments end in areas of diffuse, often poorly mapped faults. Lack of continuity of the rifts that make up the East African Rift System, where it cuts only continental lithosphere, can be readily understood because extension in the system is generally small, being in the 10% to 20% range (Rosendahl, 1987). Accomodation of this small amount of extension by numerous diffuse faults in the areas between distinct rift segments is easily achievable. Igneous activity is widespread and diverse in the East African Rift System although some rift segments, such as those in Lake Malawi, show little evidence of igneous activity.

Comprehensive analyses of the geology and geophysics of the entire East African Rift System may be hard to find, but there has been a great deal of productive research on most of the rift system in recent years. For example: In Kenya there has been a recognition of the important role that the older Anza Rift has played in more recent rift development, especially in Turkana. This recognition has involved many different kinds of investigation including geological exploration for oil both by surface and subsurface methods, especially seismic reflection techniques and the drilling of wells. Seismic refraction and teleseismic studies, potential field studies, field geology, volcanic studies, geochemistry, fission track studies and isotopic age determination have all played an important part in recent research in Kenya (see papers in Prodehl *et al.*, 1994). Elsewhere, it has not yet proved feasible to apply such a broad range of methods. As a result, information about much of the East African Rift System is presently patchy and of very uneven quality. I mention this as a cautionary introduction to my discussion of the history of the rift system in terms of four intervals of about 10 My each. I have rushed in (Pope, 1711) to compile four sketch maps summarizing what happened during each of these four intervals. These maps indicate how I think the area now occupied by the East African Rift System appears to have evolved during the past 45 million years.

Setting the Scene Between 45 and 35 Ma

Conditions sketched in Figure 31 show Africa with Arabia as a promontory on its north-eastern corner. Marine sediments, mainly limestones and sandstones, were being deposited during these Late Eocene times across a rifted continental margin which extended from Libya and Eygpt northward up the Levant coast. From southeastern Turkey to Oman, similar marine sediments were being deposited on top of an island arc that had collided with the Arabian promontory of Africa at about 95 Ma. A few occurrences of the same kinds of sediments lie scattered over the interior of Arabia (McGuire & Bohannon, 1989). Rifting in the Euphrates valley and folding in the Palmyrides and Syrian Arcs, which extends into the western desert of Egypt, had accompanied that collision. Sporadic occurrences of Late Eocene marine sandstones, evaporites, and limestones are reported from the east facing rifted margin of Arabia and Africa from Oman to Somalia. In Somalia, Eocene continental sandstones outcrop.

In Kenya and Tanzania, the rifted continental margin of Africa which marks the place from which Madagascar departed during the Jurassic (~160 Ma) overlies rift structures of Late Permian or Karoo age (~270 Ma). Kent *et al.* (1971) reported the accumulation during the Paleocene and Eocene (65 - 34 Ma) of up to 2 km of marine shales, sandstones and limestones at various localities close to the coast of Tanzania and on the offshore islands of Zanzibar, Pemba and Mafia. The Eocene shoreline lay about 50 to 100 km inshore of the present shoreline (Kent *et al.*, 1971). These Tanzanian sections are quite thick and indicate that a substantial river system may have been draining the area to the west. In Kenya, the Lamu embayment was a major site of continental margin deposition because of its position at the mouth of the then active Anza Rift (Bosworth & Morley, 1994; Walters & Linton, 1972).

The thickness of sediment accumulated during the Late Eocene in the Anza Rift itself (Bosworth & Morley, 1994; Winn *et al.*, 1993) indicates that Anza marginal faults may have been active as late as about 35 Ma. This may also have been the case in the Muglad, Blue Nile and Melut Rifts. Active faulting in the Anza Rift between 65 and 35 Ma is considered by Foster & Gleadow (1996) to be compatible with their interpretation of a pattern of apatite fission track ages whose distribution straddles the currently active Kenya Rift. Foster & Gleadow relate the fission track pattern which they have recognised to eastward dipping faults of the Anza Rift system. Those faults, they suggest, are distributed over a distance of about 300 km and are interpreted as having generated topography that was largely removed during an episode of relatively rapid erosion beginning between 60 and 70 Ma.

Lateritized surfaces are reported from various parts of the huge area sketched in Figure 31. Those in northeastern Saudi Arabia (Collenette & Grainger, 1994) formed in Cretaceous times and that in the Yemen (Geukens, 1966) overlies Late Cretaceous to Early Cenozoic marine sediment. The sub-basalt laterite of Eritrea (Drury *et al.*, 1994) has been considered contemporary with that in Yemen. In Ethiopia, in a roughly 300 x 300 km region of the Omo river valley around latitude 5^{o} North, Davidson (1983) reported outcrops of iron stained conglomeratic sandstones a few meters thick. This unit is older than the 45 Ma hot spot volcanism of the area (Ebinger *et al.*, 1993). Occurrences of laterite in Kenya and Uganda (B. King *et al.*, 1972; B. King, 1978) lie on a surface thought to be of Late Cretaceous to Early Tertiary age (~70 - 50 Ma). Foster & Gleadow (1996) obtained fission track results from rocks high in the Cherangani hills on the western flank of the Kenya Rift, and close beneath a lateritized surface, that are compatible with generation of that laterized surface at about 60 Ma. I interpret all these observations as indicating that the area of Figure 31 was mainly occupied between 45 and 35 Ma by a low-lying surface perturbed by rift shoulders and rift depressions in the Muglad, Blue Nile, Melut and Anza regions.

I have depicted short-lived igneous events at Jebel Uweinat at 45 Ma (Schandelmeir *et al.*, 1983) and in the emplacement of Africa's youngest, the kimberlite at Mwadui south of Lake Victoria, at 41 Ma (Davis, 1967). In contrast volcanism in Ethiopia persisted in an area, later to become occupied by part of the southern main Ethiopian rift, for a 10 million year interval between 45 and 35 Ma. About 500 m of flood basalts and associated tuffs were erupted over an area of about 5,000 square km (Ebinger *et al.*, 1993). By 35 Ma, basaltic eruptions had begun in the Lotikipi area west of Lake Turkana (Figure 31). The volcanic rocks erupted represent the earliest Turkana igneous activity. Morley *et al.* (1992, p. 347) emphasized that the eruption in Lokitipi predated rifting and significant extension and may have been related to a mantle thermal anomaly. The Lokitipi eruptions have been followed in Turkana by a history of half graben development and eruption that persists today (Morley *et al.*, 1992). Because there has been no relative motion between the African Plate and the mantle plume population for the past 30 My, there may still be a discrete plume underlying Turkana. The influence of such a plume may be difficult, if not impossible, to discern because of the large amount of extension and consequent related pressure release volcanism in the area (Morley *et al.*, 1992; papers in Prodehl *et al.*, 1994). Ebinger *et al.* (1993) suggested that the 45 to 35 Ma southern Ethiopian basaltic activity might also have been hot spot volcanism related to an underlying plume because little extension was involved. The southern Ethiopian and Lokitipi volcanism may have been related to the same plume. Africa did not come to rest over the mantle plume population until about 30 Ma. Rotation of Africa over a fixed plume population, using the 45 to 35 Ma track of the Walvis Ridge as control, moves the continent over Ebinger's plume and places it beneath Lokitipi by the Early Oligocene (35 - 30 Ma). Rift activity began during the Late Oligocene (30 - 25 Ma) in Lokitipi and was preceded by ~5 My of, possibly, plume generated hot spot volcanism. Interaction with the underlying older rift structures of the Melut and Anza population has perhaps contributed to the dominance of half-graben with east dipping master faults in western Turkana. Rifting in Turkana beginning at 30 Ma (Morley *et al.*, 1992) is coincident, within current resolution, with the inception of the Red Sea and Gulf of Aden Rifts as well as with the eruption of the Afar plume. This observation helps in demonstrating the approximately simultaneous start of the development of the East African Rift System in widely separated areas at ~30 Ma.

In summary, Compilation of various kinds of geological information from widely scattered sites spread over the huge area on which the East African Rift System was to develop (Figure 31) indicates that between 45 and 35 Ma, much of the area was flat-lying and lateritized, but that rifts of the Anza population were still active; that marine sediments were accumulating around the African and Arabian shores; and that igneous activity was limited in extent and intensity. The scene was set for the spectacular changes at about 30 Ma, which represent the beginning of the present episode of activity in the East African Rift System.

An Abrupt Beginning: 35 to 25 Ma

By 25 million years ago, the East African Rift System was active over a length of about 4,000 km stretching from the Gulf of Suez to the Mozambique Channel (Figure 32). Activity has continued since, in some places episodically, in most of the areas that showed signs of some activity between 35 and 25 Ma. The most conspicuous regional extensions of the rift system since 25 Ma have been into the western rift and along the Dead Sea transform. The latter has mainly developed within the past 5 My, but the time of initiation of the western rifts is poorly known. I choose a time for the western rift system origin of about 15 Ma mainly using evidence cited in Ebinger (1989). Comparable thicknesses of sediment in Lakes Tanganyika and Malawi lead me to conclude that much of the length of the western rift may have developed simultaneously.

Propagation of the East African Rift System over the latter half of Cenozoic times, and even into the future, has been a popular topic of speculation (Scholz *et al.*, 1976; Girdler, 1991, his figure 1b). There is plenty of evidence of local propagation within the system on length scales of hundreds of kilometres (papers in Prodehl *et al.*, 1994), but on the continental scale of a thousand kilometres or more, only the western rift provides any indication of long distance development and it is not easy to see that development as representing the propagation of any pre-existing rift structure.

Rifts in the Gulf of Suez (Montenat, 1988) and the Lacerda Basin (Mougenot *et al.*, 1986), which are 4000 km apart, were both receiving sediment by 30 Ma. By that time, tectonic and igneous events had begun in such scattered areas in between as Eastern Uganda, Turkana, the southern Red Sea and the Afar (Figure 32). Some authors have considered extensional phenomena in remote parts of the African Plate as related to propagation of the East African Rift System. Many of these faulting episodes, such as those in the Benue, Zambesi and Limpopo Rifts, involve reactivation of Mesozoic rifts. I prefer to consider them part of the distinctive general tectonic development of the African Plate over the past 30 My rather than directly related to the East African Rift System. Because of continuity with the Lake Malawi Rift, I do consider young faults in the Shire valley and the Urema graben (De Buyl & Flores, 1986) as representing not only reactivation of a Mesozoic rift, but also part of the East African Rift System.

Just as there is an element of reactivation of the Mesozoic Anza Rift in the Turkana sector of the eastern rift and reactivation of Karroo (~270 Ma) and Jurassic (~160 Ma) rifting in the Tanzanian coastal rifts (Kent *et al.*, 1971), the western rift has reactivated segments of Mesozoic rifts which underlie its active areas in such places as Lake Rukwa (Rosendahl, 1987, figure 1).

Developments in the area presently occupied by the Afar are critical for understanding what happened between 35 and 25 Ma in the East African Rift System. The time of eruption of the first igneous products generated in response to the impingement of the huge Afar plume on the base of the continental lithosphere has in the past been considered to have been as long ago as 50 My, but more recent studies (White & McKenzie, 1989) ascribe the main episode of flood basalt volcanism to 30 to 20 Ma. A current study dedicated, at least in part, to determining the duration of the eruptive episode finds that the earliest Ethiopian trap basalts were erupted from the Afar plume very close to 30 Ma (Hofmann *et al.*, 1995) and 30 Ma is the time preferred for first eruption by Schilling *et al.* (1992) in their exhaustive study.

There is little direct evidence of the timing of initial rift faulting in the Afar simply because so much has happened there since. Barberi *et al.* (1972; 1975) considered that rifting began about 25 Ma and this is consistent with the estimated timing of faulting in the southern section of the Afar (Zanettin *et al.*, 1978). WoldeGabriel *et al.* (1990, figure 11) show rift-faulting both preceding and following, but mainly following, the eruption of the Kella basalts, aged about ~30 Ma, in the Guraghe-Kella area of the central main Ethiopian rift some 400km to the southwest of the central Afar. WoldeGabriel *et al.* (1990) consider the post 30 Ma rifting as more important than the pre 30 Ma faulting.

Coleman (1993), who appears sceptical about the existence of more than a very minor plume at the Afar, has pointed out that the earliest 30 million year old basalts of the Afar plume were erupted on top of lateritized surfaces in Yemen and Eritrea. This, he suggests, shows that uplift above the plume head during the 50 Ma to 34 Ma interval did not precede the basaltic eruption as Gass (1970, p. 287) and some others had conjectured. Because Figure 31 shows that the places where the basalt overlies laterite lie nearly 500 km away from the centre of plume impact Coleman's suggestion cannot be definitive. It would be interesting to know what had happened closer to the centre of plume impact but erosion and extension have removed material from the critical area. WoldeGabriel *et al.* (1990) reported no laterite occurrences from on top of the thin Mesozoic sequence which occurs on the rift shoulder of the main Ethiopian rift. Perhaps this is evidence of limited pre-eruption doming, or at least uplift at some time, in that particular area.

Anyway, it doesn't matter too much whether uplift or eruption happened first because so much happened so quickly. The sequence of events may have been: Close to 30 Ma initial eruption of lavas, mainly basalt, related to the Afar plume. These lavas extended by 25 Ma, and possibly much sooner, over an area about 1,000 km in diameter (Hofmann *et al.*, 1995; Menzies *et al.*, 1990; Drury *et al.*, 1994; White & McKenzie, 1989; Schilling *et al.*, 1992). Basement uplift appears to have begun at the same time, within temporal resolution, as the eruptions. Eruption in the Afar was simultaneous, again within temporal resolution, with ~30 Ma faulting in the Red Sea that by 25 Ma formed a rift between 30 and 75 km wide (Bohannon & Eittreim, 1991). Evidence of rifting includes "initiation of volcanism and first structural tilts of strata" (Bohannon & Eittreim, 1991). This information relates to observations on the Arabian side of the Red Sea that extend as far north as about Jiddah. There is also evidence of the deposition of rift sediments by 25 Ma in Midyan on the Saudi Arabian shore at the northern end of the Red Sea (Bayer *et al.*, 1988) as well as in the Gulf of Suez Rift (Montenat *et al.*, 1988).

It looks as though by 25 Ma the entire length of the Red Sea, almost 2000 km in extent and including the Gulf of Suez area, was occupied by a narrow, newly formed rift system. The rift faults were of almost uniform north-north-westerly trend. Variations in azimuth show best in the aero-magnetic signatures of a spectacular set of basaltic dikes that were emplaced in a zone about 100 km wide along the entire length of the Red Sea Rift around 22 Ma (Bohannon & Eittreim, 1991; Blank, 1977; Coleman, 1993; Qureshi, 1971). There is much less evidence of the timing of onset of rifting in the Gulf of Aden, but for the purposes of Figure 32, I have assumed that it was behaving in the same way as the Red Sea.

A good question is why did these rifts develop? One explanation may be that which Bailey (1993) considered important for the renewal of African igneous activity. Thirty million years ago was a time of change in the convergence of Africa and Eurasia. Although what was then the Arabian promontory of Africa would not finally collide with Turkey, Iraq and Iran until 15 Ma, the Apulian promontory of Africa was beginning to impact Europe and so was the Maghreb far to the west (Dewey *et al.*, 1989).

Another possibility is sketched in Figure 33. Major forces acting on the African Plate at 30 Ma are likely to have included "ridge push" from the spreading centers of the South Atlantic and Indian Ocean and "slab pull" from subduction under Eurasia. By analogy with today, these forces would have been likely to generate within-plate shortening stresses (Zoback, 1992). I suggest that the direction of maximum shortening stress within East Africa and Arabia had an azimuth trending roughly north-eastward in a direction somewhat anti-clockwise from that in

Africa today because of the different arrangement of Indian Ocean spreading centers. By analogy with conditions in the East African Rift System today (Zoback, 1992), impingement of the Afar plume would have modified the general stress distribution in the African Plate and effectively reduced the shortening stresses over the Afar region.

Extending forces could have come to dominate locally and propagation of stresses from that modified stress field in the Afar region through the plate would have interacted most effectively with other localities at which there were unusual local stress distributions. The most obvious anomalous area not too far away was the corner between the Levant coast and the coast of Egypt (Figure 33). Because of the unusual shape of the continental margin in that region, larger than normal extensional corner stresses would already have been in existence. I suggest that the development of the Red Sea Rift System, which is of exceptional length and formed so rapidly with an unusually constant azimuth, is a result of interaction between a pre-existing region of peculiar stress distribution, the Levant corner, and the establishment within the continent of a second peculiar area: that overlying the site of impingement of the Afar plume head. As Steckler & ten Brink (1986) showed, the Red Sea Rift did not propagate into the ocean floor at the Levant corner because of the great age (>200 Ma) and strength of that ocean floor. Once the azimuth of the Red Sea Rift was established, those of the Gulf of Aden and the main Ethiopian rift developed at roughly 120^{o} from the master rift (Burke & Dewey, 1973).

The timing of early East African Rift System events in the main Ethiopian rift has been addressed by WoldeGabriel *et al.* (1990). They found that eruption of the Kella basalts took place at about 30 Ma and was both preceded and followed by episodes of rift faulting. Meanwhile, what was happening farther south in the East African Rift System? Half grabens with master fault downthrows to the east were developing between 30 and 22 Ma in Lotikipi and for about 100 km to the east (Figure 32; Morley *et al.*, 1992). These faults may have been influenced in their trend and inclination by older structures of the underlying Anza Rift System. Basalts which had begun to be erupted in Lotikipi at 35 Ma extended about 100 km to the east in Turkana and the half grabens in which the basalts are preserved also received sediments from the south and southwest. A question is: Were those sediments eroded from a "Kenya Dome" which has often been postulated to have existed?

Smith (1994, pp. 6 and 7) has made an agonizing reappraisal of the evidence for the existence of a Kenya Dome which might represent lithospheric uplift that preceded volcanism on the future site of the Gregory Rift. Smith concluded: ". . .the available evidence supports the presence of an elongate area of limited <1 km crustal uplift which formed prior to the initiation of volcanism in the Gregory Rift. . .the uplift is post-Cretaceous in age and had more or less ceased by Early to mid Miocene times. . ." This conclusion is compatible with the results of Foster & Gleadow (1996) who state that "no area of basement rocks [on either side of the Kenya Rift] has yet been identified where cooling [in the past 25 Ma] exceeds about 50 degrees C." This indicates to me that there can only have been modest denudation (< 2 km) of the Kenya Rift area over the past 25 My.

On Figure 32 I have indicated, following Smith (1994), the existence of an area of higher ground over the site of the future Kenya Rift and I have also shown an area where small carbonatite plugs were emplaced between 31 and 25 Ma in western Kenya and eastern Uganda (B. King *et al.*, 1972; Le Bas, 1987). These bodies are unusual when viewed in the continent-wide perspective because they are emplaced into an area of older crystalline rocks that had not been reactivated during Pan-African times.

Oligocene marine sediments with ages between 34 and 22 Ma are known from various places around the shores of the Afro-Arabian continent from Fayum (Kappelman *et al.*, 1992; Said, 1993), through Iran, on Socotra, in the Lamu embayment and along the coast of Tanzania. Kent *et al.* (1971) reported 485 m of Oligocene sediments from a well on Pemba island. The Oligocene shoreline, which was far offshore in much of Africa, is shown by Kent *et al.* (1971) as roughly coincident with the present coastline in Tanzania and this is perhaps an indication of later regional elevation of that area. The coastal Oligocene sediments of northeastern Africa and locally in the interior of its then Arabian peninsula (Bohannon *et al.*, 1989) do not appear very different from those deposited during the earlier Paleogene between 65 and 35 Ma.

Very different Oligocene marine sediments were deposited in two offshore rift basins, the Kerimbas and Lacerda Basins, which lie about 100 km from the northeastern coast of Mozambique in a depth of about one km of water (Figure 34; Mougenot *et al.*, 1986). Although these rifts are in some ways the poorest known of the East African Rift System, they are in other ways the most fascinating. Earthquake epicenters show that the Kerimbas and Lacerda Rifts are active now. They contain sediments indicative of the timing of initiation of active rifting as old as about latest Eocene (35 Ma). This result comes from tying seismic reflection lines to the nearby DSDP 242 (Figure 35). Two igneous bodies, the St. Lazare and Paisley Seamounts, each several tens of kilometres across at the sea bed, lie at nodal or perhaps accomodation zone positions among the six or seven individual rift segments which together make up the Kerimbas and Lacerda Basins. There is no sign of active igneous eruption in those areas and there have been no obvious major breaks in deposition in the rifts which have accumulated 3 to 5 km of sediment over the past 35 million years. Like the onshore rifts of the East African Rift System, the Kerimbas and Lacerda Basins do not appear to represent sites of great extension.

In summary, the interval of 35 Ma to 25 Ma was dominated by the eruption of the Afar plume and the perhaps related rapid initiation of the Red Sea Rift. The rifts of the Gulf of Aden may have formed at the same time, but their origin is not as well dated. In the interior of the continent, volcanism had perhaps migrated from southern Ethiopia to Lokitipi. The first Turkana Rift structures began to develop about 30 Ma over a possible site of both hot spot volcanism and the older Anza Rift structure. There was no igneous activity farther south in the region of the future Kenya Rift but there is some evidence for limited elevation on the future site of the Gregory Rift. Small carbonatite plugs were emplaced into the craton in western Kenya and eastern Uganda at this time. As in the rest of the African Plate, the spectacular change from a tectonically and volcanically quiet time to one of great activity took place at about 30 Ma. The distinctive features of East Africa at the time of the change are overwhelmingly dominated by the development of the Afar plume, the Red Sea and possibly also the Gulf of Aden Rifts. East Africa was an area where minor rift activity had persisted in the Anza system. That rifting intensified in Turkana at 30 Ma. There was one area in southern Ethiopia where hot spot volcanism had been going on for 10 million years. That volcanism might represent a plume which was later partly responsible for the early eruption of basalt in Lokitipi. Far offshore, in the Indian Ocean, plate structure was being radically reconstructed with the formation of the Central Indian Ridge and Carlsberg Ridge sectors of the Indian Ocean spreading center system.

The new rifts develop from 25 to 15 Ma

The ten million year interval from 25 to 15 Ma was a time of limited innovation in the structural development of the East African Rift System in comparison with the time around 30 Ma which marked the establishment of the system. The newly formed structures evolved without radical changes and innovations were relatively minor when compared with those that had happened at 30 Ma (Figure 36). These innovations included the eruption of the Samburu flood basalts on the site of the future development of the Gregory Rift in Kenya and perhaps the propagation of rifting southward from the Afar into southern Ethiopia.

Along the entire length of the Red Sea, including the Gulf of Suez, an intracontinental rift system typically 30 to 75 km wide had become well established by 25 Ma (Bohannon & Eittreim, 1991, p. 139). Non-marine and locally marine sediments (Dullo *et al.*, 1983) were being deposited in the rifts. Volcanic rocks from Suez have yielded isotopic ages from 26 to 22 Ma. Both flood basalts and numerous individual igneous centers are particularly well described from Yemen and southern Saudi Arabia (Coleman, 1993, pp. 43 - 49 and 59 - 62). The great 100 km-wide dyke swarm which runs the length of the Red Sea (Blank, 1977) was being emplaced. Baldridge *et al.* (1991, p. 188) report ages for the dikes ranging from 26 to 18 Ma with about 20 Ma as the peak emplacement time, but Coleman (1993, p. 62) urges caution in interpreting ages from dikes that may have reacted with potassium rich country rock and emphasises the ages from rheomorphosed contact rocks of 22 Ma.

In the Red Sea, access to the waters of the world ocean over the interval since the initial rifting at about 30 Ma appears to have been mainly from the Tethys to the north. Marine sediments of Early and Middle Miocene age (~22 to ~10 Ma) are best known from the Gulf of Suez where the Belayim, Rudeis and Kareem formations consist of shales and marls with limestone and evaporitic intercalations (Montenat *et al.*, 1988). These rocks grade into sandstones and conglomerates close to the rift margins. Sequences similar to those of the Gulf of Suez may be widespread farther south within such major lithounits as the Maghersum Group of Sudan and the Infra Evaporite Series of Saudi Arabia, although relatively few wells have yet penetrated those rocks (but, see for example: Coleman, 1993, figure 2.9). Tantalizing glimpses of variations in depositional character within the Red Sea Rifts have been gleaned from field studies on both shores (Purser & Hötzl, 1988).

The occurrence of marine rocks in the 2000 km long sub-sea level Red Sea Rift must have depended on the interaction of global sea level changes with the geometry of access through the Mediterranean. It is likely that for part of the time during the 25 to 15 Ma interval there were no ocean waters in the narrow Red Sea Rift valley. From time to time the rift valley was likely to have been flooded by the sea. Episodes of restricted circulation are recorded by the occurrence of evaporites.

Igneous rock ages in the flood basalts of Yemen and in the Arabian Plateau basalts indicate a decline in activity after 25 Ma and none is reported younger than 20 Ma (Coleman, 1993, figure 4.1). The youngest ages reported on the rocks of the great Red Sea dyke swarm paint a similar picture of declining activity at the rift margins. Although younger ages are absent from the Arabian shore of the Red Sea until the younger harrats began to erupt after 15 Ma, it seems likely that igneous activity may have been in progress in a now inaccessible but then active part of the rift which is buried under younger rocks and the oceanic waters of the Red Sea. Indeed sporadic volcanic rock occurrences recorded in the Rudeis, Kareem and Belayim formations of the Gulf of Suez may be representative of a much larger volume of igneous rock deeply buried closer to the rift axis in the Red Sea.

Ages of igneous rocks associated with the Gulf of Aden intracontinental rift also tail off as 20 Ma is approached like those from the Red Sea but there is even less evidence of how the Gulf of Aden Rift evolved between 25 and 15 Ma than there is for the Red Sea. There is certainly no evidence that the two rifts were evolving differently.

In the Afar, where there had been a great burst of volcanism at 30 Ma, there is evidence of well-established rifting by 25 Ma. The sediments of the Afar Red Series, which consist mainly of alluvial and lacustrine deposits and which have been deposited on the floor of the Afar, contain intercalated basalts that have yielded K/Ar ages ranging from 25 Ma to 5.4 Ma (Tiercelin *et al.*, 1980). Because the Afar Red Series seems to me typical of rift-deposited sediment, I suggest that the Series has been deposited episodically in active rifts of the Afar since 25 Ma or slightly earlier times. Faulting episodes in the main Ethiopian rift area to the south include some older than 30 Ma, but major initial rift developments have been assigned ages between 18 and 11 Ma with some concentration around 15 Ma (Mohr, 1983; WoldeGabriel *et al.*, 1990; Ebinger *et al.*, 1993). The main Ethiopian rift has been episodically active and has changed from a pattern of along strike alternating half-grabens to the present more symmetrical style only about 5 Ma (WoldeGabriel *et al.*, 1990).

Rift activity continued between 25 and 15 Ma in the Turkana half-grabens with both volcanic and sedimentary fill, the latter continuing to be derived in part from the south (Figure 36; Morley *et al.*, 1992, their figures 3 and 11). The cratonic area in western Kenya and Uganda, which had been the site of eruption of small carbonatitic bodies between 31 and 25 Ma, became an area in which six or more major, strongly alkaline carbonatitic volcanoes developed between 22 and 12 Ma (B. King *et al.*, 1972; Le Bas, 1987).

Although the Gregory Rift had not yet begun to develop during the interval 25 to 15 Ma its future site which had perhaps been occupied by the modest elevation of the Kenya Dome became, for the first time, a region of igneous activity. Between 20 and 12 Ma about 1 km of the Samburu flood basalts and their correlatives were erupted over an area of perhaps 50,000 square km (Smith, 1994, p. 9). Smith (1994) concluded that a mantle plume underlies the site of the eruption of the Samburu basalts and this seems appropriate because there was no contemporary rifting and the elliptical shape and size (~300 km x 100 km) of the Samburu province resembles those of the many other volcanic provinces overlying mantle plumes which have developed in Africa over the past 30 My (Figure 6). What is distinctive about the Samburu flood basalts is not what happened when they were erupted, so much as what has happened on the site of their eruption during the past 10 My. I know of no analogues in the Africa of the past 30 My to the one km thick Plateau phonolites which were emplaced above the Samburu flood basalts between 16 and 8 Ma (Smith, 1994, p. 9). The still more recent development of the Gregory Rift on the site of the eruption of both the Samburu basalts and the Plateau phonolites is equally a distinctive event.

Between 25 and 15 Ma, during the earlier Miocene, there is evidence of both rapid deposition and normal faulting in coastal and offshore Tanzania (Figure 36; Kent *et al.*, 1971, their figure 44). Approximately 600 m of detrital Miocene limestone accumulated in the Kerimbas Rift Basin and a basaltic sill was intruded into those rocks at about 15 Ma (Mougeneot *et al.*, 1986).

Farther away, the Nile delta first developed at about 25 Ma (Said, 1993, figures 1.15 and 1.16). Its formation was the result of erosion from the newly formed Red Sea and Afar swells as well perhaps as the discrete swells of Darfur and Jebel Uweinat. All these areas may have been sources of Nile delta sediment (Figure 6). In the Indian Ocean, the Central Indian and Carlsberg Ridges, formed at about 30 Ma, were propagating between 25 and 15 Ma toward the Gulf of Aden Rift which would be entered at about 10 Ma.

In summary, Between 25 and 15 Ma the structures of the East African Rift, which had formed at ~30 Ma, continued to develop, but few new structures were formed. Igneous activity persisted in the areas where it had begun at ~30 Ma. The most conspicuous new igneous feature was the Samburu flood basalt eruption which had many of the characteristics of the new population of African intraplate hot spots. The Samburu hot spot, erupting at ~20 Ma, appears to overlie the third identifiable plume involved in the evolution of the East African Rift System. The first had generated the basalts of southern Ethiopia between 45 and 35 Ma and, I suggest, had later contributed to Lokitipi volcanism starting at 35 Ma. The second was the giant Afar plume which appears to have impinged on the bottom of the lithosphere and started to erupt at ~30 Ma. These three plumes, and pressure release volcanism in extending rifts, may serve to account for all of the East African Rift igneous activity. Off-rift igneous provinces that are sometimes treated as integral parts of the East African Rift System such as the Uganda carbonatites and the Marsabit, Chiyulu hills, Nyambeni and Kilimanjaro provinces call for independent plume sources similar to those that have been active over much of the African Plate during the past 30 My (Figure 6). Many of those provinces, in East Africa, are more appropriately thought of as related to the East African swell rather than directly to the rift system.

Did Collision Take Over?: 15 to 5 Ma

The most prominent feature of Figure 37 is the collisional suture of the Bitlis and the Zagros, marking the place where Arabia has collided with Asia. The collision, which has been dated in various ways, appears to have taken place over a few million years ending at ~10 Ma in Turkey (Sengör & Yilmaz, 1981) and at ~15 Ma in Iran. One of the most persuasive indications of the timing of the collision is a change in bottom water characteristics as indicated by benthic faunas in the northern Indian Ocean (Woodruff & Savin, 1989; Smart & Ramsay, 1995). Tethys-derived intermediate and deep ocean water stopped flowing into the northern Indian Ocean at some time between 16.5 and 15 Ma. This abrupt change can be attributed to the collision which closed off the Mediterranean as a remnant of the Tethys.

The new harrats shown in Figure 37, which reach as far north as Karacelidag, just south of the Bitlis suture in Turkey, and occur also in Syria, Jordan and Arabia, can be recognised as a coherent population of volcanic provinces because: (1) their oldest erupted rocks yield ages of less than 15 Ma, marking a renewal of volcanism in Arabia after an interval of about 5 My (Coleman, 1993, figure 4.1); and (2) Their volcanic centers, of which there are over one hundred, are arranged in north-south lines or, locally, in lines that strike up to 20 degrees east or west of north. They differ from alignments of the older harrat centers and dikes of the Red Sea swarm which trend in a direction parallel to the Red Sea.

I relate both the eruption of the new harrats and the east-west azimuth of extension, indicated by their north-south volcanic alignments, to the Zagros collision primarily because of a coincidence in timing. The oldest isotopic age from a newer harrat is ~13.5 Ma (Camp & Roobol, 1989) and the collision was at ~15 Ma. At the collision, a “collision force” was substituted for the “slab-pull force” all along the length of the suture. Interior stresses in the Arabian-African Plate were modified and east-west extension developed over an area of more than one million square kilometres of eastern Arabia. There is no obvious change in tectonic style in the Red Sea itself but the formation of new ocean floor propagated for the first time into the Gulf of Aden by 10 Ma (Cochran, 1981). That may have been, at least in part, a response to the change in regional stress conditions in southwestern Arabia as well as a result of the mechanical weakness of the continental lithosphere in comparison with the 150 Ma oceanic lithosphere of the Somalia Basin (Steckler & ten Brink, 1986). Another possible indication of changed conditions in Arabia lies in the present topographic contrast between the eastern and western Red Sea shores (Dixon *et al.*, 1989). The Sudan and Egyptian shores are typically African with steep slopes reaching to the coast, but there is a wide coastal plain on the Arabian side which may be an indication that the continuing uplift which dominates in Africa is no longer operating in Arabia. Uplift of the coasts would be much more symmetrical if it resulted simply from the thermal influence of the oceanic spreading center in the Red Sea.

At the southern end of the Red Sea in Yemen, there is no such topographic contrast. The huge Afar plume was presumably playing as important a role in generating topography in that area between 15 and 5 Ma as it is today.

Discriminating events and establishing the history of faulting and volcanism in the Afar and in the Ethiopian Rift is rendered difficult by the extent of Pliocene and Quaternary volcanic rocks and by associated rift faulting. Mohr (1983, p. 178) presented a tabulation of events on the eastern and western flanks of the northern main Ethiopian rift at about 9^{o} North. The work of WoldeGabriel *et al.* (1990) about 100 km farther south is compatible with Mohr's analysis, but places more emphasis on the ages of volcanic rocks and the changes in Pliocene and later times. Mohr (1983) discerned flood basalt events with ages between 28 and 22 Ma and episodes of downwarping and faulting at 15 and 10 Ma.

Farther south at about 7.5 degrees north, Mohr (1983) reported ~30 Ma flood basalt eruption and the eruption of flood trachytes and dikes between 19 and 15 Ma. In the southernmost part of the main Ethiopian rift, near six degrees north, Ebinger *et al.* (1993) showed that rift volcanism and faulting had developed in the area where there had been hot spot volcanism between 45 and 35 Ma. Mohr (1983, p. 180) distinguished flood basalt eruption and faulting in this area at about 20 Ma from further eruptive episodes at 17 and 13 Ma, which were followed by a major block faulting and erosional event. Renewed activity in this area is Pliocene and younger.

Mohr (1983, p. 186) summarized the general evolution of the Ethiopian Rift as involving broad crustal downwarping and mainly flood basaltic volcanism between 30 and 22 Ma. Workers on volcanic rocks in the East African Rift System commonly report "downwarping." I interpret the term to mean recognition of a thick sequence of lavas with little sign of faulting.

The interpretation has been that a downwarped basin must underlie the volcanic rocks. This is doubtful. The volcanic rocks have generally been erupted in areas of basement uplift so that the reported basins are non-existent. The eruptions which Mohr reports over the interval 30 Ma to 22 Ma, I interpret as the initial products of the eruption of the Afar plume. The establishment of the rift margins by a faulting episode at 15 Ma was, according to Mohr (1983, p. 186), accompanied by rhyolitic eruption. There was more faulting at 10 Ma. Major and complex faulting and eruptive episodes have taken place throughout the Ethiopian Rift during the past 5 My.

In Turkana, the interval between 15 and 5 Ma was occupied by continuing half graben development with abundant volcanism and sediment deposition (Morley *et al.*, 1992, figures 10 and 11). The Elgeyo fault at about 1^{o} North in Kenya provides a link between the half-grabens of Turkana and the area of the Gregory Rift in which extensive and thick Plateau phonolites were erupted at this time. Eruption of the phonolites took place between 16 and 8 Ma with most of a volume of between 25,000 and 50,000 cubic kilometres having been erupted between 13 and 11 Ma (Smith, 1994, p. 9). Some faulting in the area that would later be occupied by the Gregory Rift structure was in progress while the Plateau phonolites were erupting, but the rift structure itself may not have been established until about 5 Ma. The area of eruption of the Plateau phonolites corresponds quite closely to that of the earlier Samburu basalts and the two are realistically attributable to a plume source comparable in size to those responsible for the other alkali basalt lava provinces that have erupted in Africa over the past 30 My (Smith, 1994).

An intriguing question is whether there are any signs in the East African Rift System at, or soon after, 15 Ma of influence of the Bitlis-Zagros collisional event? The change in Kenya of the composition of eruptives, from basalt to phonolite, comes at 16 Ma which is within resolution of the time of the collision. Fifteen million years ago appears, from Mohr's (1983) work, to have been a significant time of rift development in the Ethiopian Rift. On the other hand, there is no sign of any particularly important change at 15 Ma in Turkana (Morley *et al.*, 1992). Perhaps the western rift might be a better place to look for evidence of influence of the collision.

Unfortunately the western rift (Figure 37) has proved one of the hardest parts of the East African Rift System in which to work out the sequence and timing of events. The reason is that the western rift is quite discontinuous and that much of its record is in sediments on the floors of the basins of the Great Lakes where stratigraphic timing over intervals of millions of years has not yet proved possible.

I have relied heavily on two splendid reviews by Rosendahl (1987) and Ebinger (1989). From them, I conclude that it is extremely hard to infer when the western rift began to form. The oldest Cenozoic volcanic rocks of the western rift formed about 13 Ma in the Virunga province (Figure 37), so activity had begun by then. Because volcanism is commonly one of the earliest manifestations of rift activity in the East African Rift System, I consider that it is quite possible that the western rift was initiated at ~15 Ma which was the time of the Zagros collision. The development of the western rift might have been a response to stress changes induced by that collision. My reason for thinking of that possibility is that collision appears to be the commonest way of starting continental rifts (Molnar & Tapponnier, 1975; Sengör *et al.*, 1978; Sengör, 1976; Burke & Lytwyn, 1993). Starting rifts in association with mantle plumes is another, although less common, mechanism (Burke & Dewey, 1973; Burke & Sengör, 1978; White & McKenzie, 1989). In contrast to the eastern rift, I have not yet discerned evidence in the western rift of rift-associated plume activity such as can be seen in the Afar, at Lokitipi and at Samburu. Ebinger (1989, p. 900) emphasises that four of the largest volcanic provinces of the western rift are localized in nodal areas which may have developed at accomodation zones between sedimentary rift basins. None of the young, plume generated volcanic provinces of Africa lie near the western rift (Figure 6). Indeed the northern one-third of the length of the western rift lies on top of cratonic basement whereas the rest of the East African Rift System lies entirely on continental lithosphere that had been reactivated during Pan-African times.

Once the western rift had begun to form, scattered isotopic ages on volcanic rocks (Ebinger, 1989, table 1) indicate that it remained active at least episodically and sporadically. Ages on sedimentary rocks from the western rift (Ebinger, 1989, table 2) paint a similar picture. These two sets of ages also give an impression that although faulting, deposition in lake basins and volcanic activity persist once a rift segment has formed, there has been a southward progression in the age of segment initiation. Much more evidence will be needed before this can be more than an impression.

Between 15 and 5 Ma, the Kerimbas and Lacerda Rifts continued active and sediments were deposited off the coasts of Kenya and Tanzania, but there is no evidence of any great change in either of these areas. The Nile delta continued to prograde into the Mediterranean. In the easternmost Gulf of Aden, ocean floor began to form at 10 Ma (Cochran, 1981) and by 5 Ma had propagated almost to the Afar.

In summary, During the interval of 15 to 5 Ma, a major change in the East African Rift

System was the development of the new harrats of Arabia in response to the Bitlis collision. At 15 Ma, rifting propagated southward from the Afar into the main Ethiopian rift and the western rift appears to have been initiated. It is intriguing to consider whether the changes in the Afar and initiation of the western rift may, like the development of the new harrats, have been related to the Bitlis collision.

Narrow Rifts and Plate Motion from 5 Ma

Formation of ocean floor for the first time in parts of the Red Sea, transform offset on the Dead Sea system, propagation of new ocean floor formation into the Afar from the Gulf of Aden, and southward propagation of rifting from Turkana to form the Gregory Rift constitute some of the most radical changes in the East African Rift System over the past 5 My. Rigid plate motion of Arabia with respect to Africa began, and Jestin *et al.* (1994) have treated the area east of the East African Rift as a discrete Somalia Plate. That plate is moving with respect to the rest of Africa at about half a centimeter a year or roughly half of what is usually regarded as a minimum resolvable plate speed. If this velocity represents average motion for the past 5 My, then the Somalia Plate has moved 25 km to the east with respect to the rest of Africa over that interval. This is too small a motion to show up on the scale of the phenomena under consideration here but Jestin *et al.* (1994) emphasise the importance of appreciating current motion in assessing earthquake slip vectors and other short-term strain release indicators such as bore-hole breakouts.

Whether the beginning of plate motion or other controls have been important, there have been considerable changes in the operation of the East African Rift System over the past 5 million years. The active Dead Sea transform has a strike-slip offset of about 100 km. The timing of movement along the fault-system has been much disputed but that motion may all have taken place in association with the formation of ocean floor in the Red Sea. The timing and extent of ocean floor formation beneath the Red Sea is itself a difficult problem. Early authors considered that ocean floor formed long ago, 16 Ma or even 42 Ma might underlie much of the evaporitic section near the coastlines, but most modern authors (see discussion in Coleman, 1993, p. 129) favor a beginning to ocean floor formation in sporadic localities only since ~5 to 6 Ma.

Messinian evaporation of the waters of the Mediterranean (Hsü, 1972) took place at a time close to the beginning of the formation of ocean floor in the Red Sea. Messinian time is represented by evaporites in the Gulf of Suez which has evolved in a manner distinct from that of the Red Sea only in the past 6 My. The conclusion that inception of substantial movement on the Dead Sea transform system and the formation of the oldest ocean floor in the Red Sea took place no more than ~6 Ma seems to me compatible with most of the evidence.

At the other end of the Red Sea, in the Afar, 5 Ma marks the onset of a new style of rifting and intense volcanic activity culminating locally, during the Quaternary, in what is almost an oceanic style of rift formation (Hayward & Ebinger, 1996). Those authors have contrasted the development of the Afar during the past 5 My with that of the main Ethiopian rift to the south, discerning evidence of a possible style of sequential rift segment evolution.

"The increase in crustal thinning from south to north observed in the rift system [of Ethiopia] suggests that these may form some parts of an evolutionary sequence of rift segmentation: (a) Early rift forms as a series of half-graben governed by large border fault segments. (b) With increased extension the rift becomes more symmetrical, forming a full graben with some intrabasinal faulting and commonly with felsic volcanism occurring in the graben at the segment terminations. (c) Large basin bounding faults will be abandoned when the stresses required to rupture them exceed that required to form a new fault system in the weakened lithosphere within the rift and we see a new set of smaller graben forming within the original rift depression. (d) The onset of voluminous basaltic volcanism due to adiabatic decompression melting, coupled with further decreases in the length-scales of active faults and basins due to decreasing lithospheric strength will result in rift segmentation dominated by basaltic volcanic ranges with a narrow zone of small faults along the axis of each segment. (e) Increasing extension and magmatism will lead to dike injection as the major method of strain accomodation, producing sea-floor spreading, where we will begin to see the rift valley border faulting characteristic of slow spreading mid-ocean ridges." (Quoted from caption to figure 12 of Hayward & Ebinger, 1996).

In Turkana over the past 5 My (Morley *et al.*, 1992, their figure 10 D, E and F) volcanism developed farther east while lacustrine and fluvial conditions alternated over the site of Lake Turkana. In Quaternary times the long-lived half-graben style of Turkana has been modified as numerous small-throw normal faults developed over volcanic rock in the Kino Sogo fault belt. This change in style has been attributed to the emplacement of magma in the crust and the thinning of the lithosphere. The general migration of Turkana faulting to the east with time is suggestive of developing integration with the southern Ethiopian rift as was long ago pointed out by Cerling & Powers (1977).

Paradoxically, the Gregory Rift to the south of Turkana, which has been perhaps the most famous segment of the rift system, appears to be one of the youngest to form. It has developed on the site at which the Samburu plume erupted first a kilometer thickness of basalts and then a kilometer of phonolite. Normal faults as old as 12 My occur in the area, but the full development of mappable rift segments may have happened as late as 5 Ma. The Pliocene (~5 - 2 Ma) was characterized by the eruption of large volumes of trachytic and phonolitic tuff which cover nearly 30,000 km and reach a maximum thickness of more than 2 km (Smith, 1994, pp. 10 - 14). Overall, the still active Samburu plume seems to have erupted about 100,000 cubic kilometres of volcanic rock at the surface over the past ~20 My in three discrete episodes.

Smith & Mosley (1993) have related the fault pattern of the Gregory Rift to the imposition of present stresses on a volume of lithosphere rendered heterogeneous in Pan-African times. It seems to me that at least two other phenomena may have contributed to the development of the Gregory Rift. One has been pointed out by Smith (1994, p. 12): "the coincidence of volcanism and subsidence with the centre of the Kenya dome suggests that initially the axis of this plume [the Samburu plume of this text] was located beneath the central part of [what was to become] the [Gregory] rift where the E-W-trending Nyanza Rift forms a third arm to a tri-radial junction." The other phenomenon which I suggest may have been important is that, because of the modification of lithospheric stress in response to the thinning of the lithosphere above the Samburu plume, the long-lived rifting of Turkana may have selectively propagated southward into the area of the thinned lithosphere to initiate rift formation in the Gregory area at ~5 Ma. Various authors in Prodehl *et al.* (1994) (see for example: Hendrie *et al.*; Keller *et al.* [a] and [b]) have suggested this propagation on the basis of seismic refraction experiments which have shown that the crust is much thinner beneath Turkana, where it is ~20 km thick, than it is beneath the Gregory Rift where it is 30 km thick.

Between 2^{o} and 3^{o} South, the Gregory Rift opens out into a volcanic area nearly 300 km wide that extends from Kilimanjaro in the east to the Serengeti in the west. Farther south, the Eyasi and Pangani graben extend to the southwest and to the south with much block faulting in the area between. Dawson (1992) reported that major faulting in this area is about 1 million years old and was preceded by volcanism in the form of alkali basalt and phonolite erupted at major centers with minor carbonatite volcanoes. Since 1.2 Ma, nepheline-phonolite and carbonatite volcanoes have dominated. It is not at all clear to what extent the igneous activity in this area, particularly in such large volcanic centers as Kilimanjaro and Mweru, is related to the eastern rift. I prefer to consider those volcanic centers as discrete members of the post 30 Ma plume population of Africa which are localized near to the high part of the East Africa swell that happens to be occupied by the eastern rift.

The question of the origin and variety of East African Rift igneous activity, which is well exemplified in the northern Tanzanian volcanic centers, has been much addressed as an isolated problem. There are, however, reasons for considering that challenging topic in the broader context of its relationship to the igneous activity of the rest of the African Plate over the past 30 My.

Igneous activity in relation to that of the rest of the African Plate

The volcanic rocks of most of the numerous 30 Ma and younger hot spots of the oceanic parts of the African Plate show coherence: (1) in being much less voluminous than those of, for example, Hawaii and other great Pacific island chains; (2) in being associated with a substantial element of basement uplift (Figures 14, 15 and 16); (3) in showing a dominance of derivation from the HIMU source (Figure 38 redrawn from Hart *et al.*, 1992). Because HIMU derived rocks are considered to reflect only a small degree of partial melting (Zindler & Hart, 1986) the small volumes of volcanic rock and the importance of basement uplift which characterise these hot spots are related properties. Lastly (4) these oceanic hot spots show ^{3}He signatures below those of MORB.

Tristan and its track provide the obvious exception to these generalizations but this is readily understood because Tristan alone, among active hot spots on the African Plate, is not part of the 30 Ma and younger hot spot population being the surviving manifestation of a plume which first erupted 130 Ma. For evidence of Tristan's geochemical distinction there is a good discussion in, for example, Wilson (1993, her figure 2b). The Afar is clearly a special case because of its large volume of volcanic products and its ^{3}He signal. Réunion is somewhat distinct isotopically in having a ^{3}He signature greater than MORB (Graham *et al.*, 1992). This distinctive geochemistry and its location have led to Réunion being considered to mark the present site of the Deccan trap mantle plume source (Peng & Mahoney, 1995), but I have shown that there are severe geometric difficulties with that identification. Possibly Réunion's peculiar geochemistry may be related to its being an extremely young hot spot representing, at less than 2 Ma, one of the youngest plume eruption sites in the interior of the African Plate.

The HIMU signature of the hot spots of the oceanic areas of the African Plate can be considered indicative of derivation from plumes concentrated within, if not confined to, that part of the mantle above the base of the transition zone (Zindler & Hart, 1986). The HIMU source appears to represent the same depleted source as the MORB source, a source perhaps dominated by recycled oceanic lithosphere, but HIMU-derived volcanism represents a much smaller degree of partial melting (Blusztajn & Hart, 1995). Restriction of source plumes to that part of the mantle above the base of the transition zone is compatible with the observation of Thiessen *et al.* (1979; Figures 24 and 25) that the spacing of swells on the African continent shows a concentration at about 700 km. If the source of ^{3}He is in the lower mantle, as many have speculated, then the low ^{3}He signature of the 30 Ma and younger population of African hot spots is compatible with their derivation from above the 670 km discontinuity.

The dominance of basement uplift and the eruption of only small volumes of volcanic rock which characterize hot spots on the oceanic part of the African Plate can also be seen in many of the continental hot spots of northern and western Africa (Figure 14 showing the Ahaggar) as well as in hot spots like those of Dakar and Mount Cameroon at the continental margin (Figures 13 and 16a). This structural similarity among the hot spots of the oceanic and continental parts of the African Plate leads me to suggest that the plumes beneath both areas may be geochemically dominated by the HIMU source.

On the continent, it is much harder to be confident from studies of geochemistry and petrology about the asthenospheric source of erupted volcanic rocks. This is simply because reactions with both mantle lithosphere and continental crust are likely to be involved. The observation of the coherent tectonic and volcanic behavior of the African Plate in both oceanic and continental areas which I have emphasised here provides a basis for considering that the igneous rocks of the East African Rift System may, like those on the rest of the plate, be dominated by derivation from the HIMU source. If both continental and oceanic areas have shared such features as discrete basement swells, small degrees of partial melting, small amounts of volcanic rock erupted and long-duration volcanism in the same places, then it seems worth considering that igneous rocks in all the areas of the African Plate may reflect the dominant influence of the same HIMU source.

The idea of an important role for the HIMU source in igneous activity in the East African Rift has, in fact, been widely emphasised, for example, by Hay and his co-workers (Hay & Wendlandt, 1995; Hay *et al.*, 1995) in an experimental and geochemical study of Kenya flood phonolites erupted above the Samburu plume. The study concluded that the phonolite had been generated by partial melting of alkali basaltic material. That material had been emplaced in the lower crust during the earlier history of the Samburu plume which resulted in the eruption of the voluminous Samburu basalts. Hay *et al.* (op cit.) characterized the mantle-derived source of the alkali basalt, which they postulate to have been emplaced at the base of the crust, as describable in terms of the HIMU source and enriched mantle with both EM1 and EM2 components. A degree of contamination by lower crustal continental rock was thought to be appropriate in accounting for some trace element concentrations of the phonolites. Fertile mantle lithosphere is considered to underlie areas that have been involved in Pan-African collision and delamination (Ashwal & Burke, 1989). That material is an appropriate source of EM1 and EM2 characteristics so that the work of Hay *et al.* (op cit.) can be interpreted as indicating that the Samburu plume, one of the three plumes known to be involved in the evolution of the East African Rift System, can be considered to be dominated by the HIMU source.

Volker & Altherr (1995) used a different approach in studying Quaternary volcanic rocks from Kenya. They found lavas from the Quaternary shield of Marsabit to be indicative of HIMU and MORB source derivation with some involvement of EM1. I would suggest that the latter results from reaction with underlying fertile mantle lithosphere so that again the HIMU source appears dominant. According to the same workers, Nyambeni, the Huri hills and Chiyulu appear generally similar to Marsabit with evidence of HIMU source dominance (Volker & Altherr, 1995).

Carbonatites, which are widely represented among the 30 Ma and younger population of African Plate volcanoes (Bailey, 1993), show isotopic signatures indicating the importance of the HIMU source both in Kenya (Kalt *et al.*, 1995) and in the oceanic part of the plate in the Canary islands (Hoernle & Tilton, 1991). Kalt *et al.* (1995) pointed out that carbonatites older than 40 Ma in East Africa showed evidence of a somewhat different mantle source and it is worth noting that no kimberlites younger than 40 Ma have yet been recognised in Africa. The picture for the past 30 My may be one of shallow separation of carbonatite and associated nephelinitic magma from material derived from the HIMU source under thinned lithosphere (Wyllie, 1987, his figure 6).

I have chosen to emphasise the importance of the HIMU source in the volcanism of the East African Rift System, but I do not wish to give the impression that this has been by any means the only or necessarily everywhere the most important control in the evolution of a complex and diverse set of magmas and igneous rocks. Hay *et al.* (1995), for example, emphasised the important role of storage of material near the base of the crust over about a ten million year interval. Paslick *et al.* (1995) showed how isotopic compositions of lavas from northern Tanzania indicate the importance of sub-continental lithospheric mantle in their genesis. Thompson & Gibson (1994) applied a general model of rift magmatism to the East African Rift System. That model places sodic-dominated material from the convecting mantle near rift axes and potassic-dominated material involving the mantle lithosphere closer to the margins of broad swells. Thompson & Gibson (1994) cautioned that this pattern can be obscured by crustal contamination and by magma mixing in open system reservoirs. In the equatorial region of the East African Rift System, they found their model generally applicable. However locally in areas like that east of the Gregory Rift, where the evolution of the Anza Rift between 140 and 30 Ma had (they suggested) led to compositional modification of the underlying mantle, their model did not work so well. MacDonald *et al.* (1994), who brought decades of experience to bear on the magmatism of the Kenya Rift, saw evidence of both convecting and lithospheric mantle involvement. Workers on the volcanic rocks of the East African Rift, for example Bloomer *et al.* (1989), have emphasised the role of shallow processes, dominantly crystal liquid differentiation but also magma mixing, multiple intrusive episodes and the incorporation of earlier lavas into younger eruptives (Karson & Curtis, 1989).

The Afar and the Ethiopian traps, that began to be erupted from its plume source at 30 Ma, are the most challenging of igneous products associated with the evolution of the East African Rift System. This is simply because of their size. Fortunately Afar geochemistry has been the subject of a comprehensive and masterly treatment by Schilling *et al.* (1992) who found that the Afar plume resembled the other plumes of continental Africa in being dominated by the HIMU source and in showing the involvement of both EM1 and EM2. Can the Afar plume then, like the other African Plate plumes, be considered to come from above the base of the transition zone? Like the other African hot spots, the Afar is associated with substantial basement uplift. The ~2 km thick Ethiopian traps lie on basement which has been elevated to as much as 2 km above sea level. Perhaps the Afar plume is just part of the general population but its ^{3}He signal is unique on the African Plate in being considerably greater than the MORB signal (Craig, 1977) and the volume of basalt erupted in the Ethiopian traps and in the Afar greatly exceeds that of the other African Plate hot spots. One possibility seems to be that the Afar plume is just a particularly large member of the 30 Ma and younger African plume population. Another possibility, suggested by the high ^{3}He signal and consistent with the huge volume of volcanic products, is that the Afar alone among young African plumes comes from great depths in the mantle. Réunion and perhaps the Cape Verdes (Gerlach *et al.*, 1988) are the only other 30 Ma and younger African hot spots that seem to me possibly strongly influenced by a source other than the HIMU source and therefore likely to have originated from great depths.

If the Afar plume is indeed unique on the African Plate, then an intriguing possibility is that its eruption may have been the trigger leading to the cessation of motion of the African Plate with respect to the mantle circulation. Morgan & Smith (1992) suggested that plume impingement on the base of the lithosphere could modify the viscosity of the asthenosphere over an extensive region, perhaps by entraining magmatic material or magma fertile material in the plume head. Tarduno & Sager (1995) showed that the Pacific Plate moved slowly with respect to the dipole magnetic field, and therefore the spin axis of the earth, at the time of emplacement of oceanic plateaus in the Pacific Ocean during mid-Cretaceous times. Tarduno & Sager (1995) suggested that the mechanism of Morgan & Smith (1992) might have operated in that process. Control by the eruption of the Afar plume rather than by change in the collision of Africa with respect to Europe (Bailey, 1992; 1993) is at least an interesting alternative for the arrest of Africa. Of course both processes could have operated together and had complementary effects.

Whatever the study of the East African Rift System can teach about the evolution of the African Plate over the past 30 My, it is essential to integrate that information with observations from the rest of the continent and from the oceanic regions of the plate. Ocean floor only began to form in the Gulf of Aden Rift within the past 10 My propagating toward the Afar. It is only within the past 5 My that ocean floor has begun to form sporadically within the Red Sea Rift. What has the tectonic history of the entire African Plate been over the past 30 My?

Tectonics of the plate interior away from the East African Rift

Areas of the African Plate remote from the East African Rift System have been dominated tectonically by the development of swells of various sizes during the past 30 My. Unlike the East African swell, the other swells have not developed crestal rifts. Those swells and their related volcanic areas have been discussed earlier in this paper mainly in relation to their geomorphological and hot spot developments. No attempt is made here to deal with the Atlas Mountains of Morocco, Algeria and Tunisia or their associated foreland basins because the tectonics of those areas relate to the Alpine system and the collisional environment of the Eurasian and African Plate boundary zone rather than to Africa itself. Questions here relate to reactivation of older rift structures within the African continent and faulting within the interior of the African Plate. Brief references to Madagascar and to smaller microcontinents of the African Plate are also included.

Reactivated rifts on the African continent

The map of Africa shows that several of the ancient rifts of the continent have topographic expression today. Some of these ancient rifts have been caught up in the East African Rift System and clearly involve faulting within the past 30 My (Rosendahl, 1987, figure 1). Others, including the Limpopo, the Zambesi and its possible extension under the Kalahari (Scholz *et al.*, 1976), as well as the Benue Rift, are not closely related to the East African structure. The Luangwa Rift is an ancient rift with present topographic expression which links the ancient Zambesi Rift at its southern end to the northern end of the modern Malawi Rift. The present Luangwa, Zambesi and Limpopo valleys are eroded into the Great Swell of Southern Africa. Because I consider the elevation of that feature to be less than 30 million years old, I infer that the three valleys which all lie within ancient rifts have been cut, very likely along the courses of older valleys, within the past 30 My. There are few mapped active faults along these valleys but Brandl (1995) has reported a 170 km long Quaternary fault system on the southern side of the Limpopo valley with escarpments as high as 10 m. Similar fault systems may exist in the Zambesi, Luangwa and Benue valleys.

A good question is: Why should old rifts react in a distinctive way to regional uplift of the swell on which they happen to lie? Possibly the modification of lithospheric structure that took place when the rifts were originally formed, and which has been accomodated during the interval of rift inactivity by the strength of the surrounding lithosphere, may have been destabilised by the new processes accompanying regional uplift.

There is no young igneous activity associated with the reactivated rifts which lie on top of the southern African swell but in the Benue valley there are young volcanic rocks. Their distribution on regional maps (Whiteman, 1982, figure 9) and my own field observations suggest to me that, although these 30 Ma and younger volcanic rocks lie within the reactivated Cretaceous Benue Rift, their distribution in relation to the nearby Jos and Cameroon line volcanic rocks makes them more readily interpretable as part of the general population of post 30 Ma hot spots of Africa rather than as related to the young reactivation of the Benue Rift. This is especially true of the Biu and Longuda Plateaus which, like the Jos Plateau, carry volcanic cones aligned tangential to the periphery of the neighboring Chad Basin.

Active Faults of the African Continent

Within-plate seismicity and seismicity remote from the East African Rift System are widespread on the African Plate and magnitudes as high as M = 7 have been reported. Zoback (1992) showed that earthquakes and other short-term strain release indicators remote from the East African Rift showed that the interior of the African Plate is generally in compression with the direction of greatest shortening trending very roughly east-west, in the direction to be expected if ridge push forces from the north-south trending South Atlantic and Central Indian Ridges dominate. Modification and rotation of stresses as a result of buoyancy in rift and active igneous areas is very local. The implication is that many of the faults within the African Plate are thrust or strike-slip faults. A good question is: Why have those faults developed in the places in which they are known to be concentrated?

The best places to study active faults are in the areas where earthquakes and faults occur together. In some of those places, there is persuasive evidence that the earthquakes have been localised on an old fault within the plate. For example, at Accra, in Ghana, and Kribi in Cameroon, earthquakes with magnitudes as high as M = 6.4 have occurred in places where oceanic fracture zones reach the edge of the continent (Burke, 1969; Sykes, 1978). These are places where abrupt changes from continental to oceanic lithospheric structure occur. Induced stresses generated at the sites of the abrupt structural changes are presumably from time to time exceeded, faults move and the earthquakes represent those ruptures.

In South Africa earthquakes have been much studied and alternative interpretations for their occurrence, such as a link to the East African Rift System or response to underlying lithospheric thinning, have been suggested (Hartnady, 1990). The possible existence of microplates to account for the seismicity of the western Cape has also been considered (Ransome & de Wit, 1992). More recently, the distribution of regional anomalous features has begun to be mapped (Andreoli *et al.*, 1995).

There are great opportunities for mapping active and recent faults on the African Plate and for seeing how they fit in with the basin and swell and rift structures which dominate the active intraplate tectonics of Africa. Frustrating considerations include the difficulty in determining when a fault moved and integrating local considerations into a broad picture. Because there are real seismic hazards at many places in Africa, there are social reasons for much more research on the active faults of the continent.

Microcontinents on the African Plate: Madagascar, Seychelles, Agulhas Plateau and Mozambique Ridge

The island of Madagascar is a relatively small continental object within the oceanic area of the African Plate. Other microcontinents on the plate are the much smaller and wholly submarine Agulhas Plateau and the Mozambique Ridge which appear, like Africa itself, to show evidence of uplift within the past 30 My (Ben-Avraham *et al.*, 1994). The Seychelles microcontinent has a substantial marine carbonate bank cover but the occurrence of outcropping granitic rock of Pan-African age is an indication of recent uplift of at least part of the bank.

Madagascar too has been the site of recent uplift and, in its history over the past 30 My, it is in many ways a smaller version of the African continent. The island constitutes a single discrete swell with two subswells. The high area in the center of the island, which is underlain by Pan-African reactivated continental rocks, is capped by hot spot volcanic rocks (Figure 6). Similar basement in the area of Cap d'Ambre at the northern tip of the island also supports a hot spot volcanic area. High ground in a separate subswell farther south, with basement that was not reactivated in Pan-African events, supports no volcanism.

Relief and structure of the oceanic part of the African Plate

Introduction

Two domains make up the oceanic parts of the African Plate and they have experienced very different histories over the past 30 My. A substantial area (Figure 18) has been generated by sea-floor spreading since the time when Africa came to rest over that part of the mantle circulation represented by the plume population. This area and conjugate areas on adjacent plates (Figure 18) have had quite a different history from the older parts of the plate. Those older areas have responded to the change in underlying mantle circulation in ways that closely resemble the response of continental Africa. Basins and swells have developed on the older ocean floor (Cloos, 1937; Nyblade & Robinson, 1994) and volcanoes have formed on the crests of some, and perhaps on all, of the swells. For example: volcanoes have formed on the Cape Verde swell and on the individual discrete swells of the off shore part of the Cameroon line, but discussion of the 30 Ma and younger ocean floor as a unique environment on earth must come first.

Ocean floor generated within the past 30 My

Ocean floor that has formed on the African and conjugate plates during the past 30 My is in many ways typical ocean floor formed from the MORB source at slow spreading rates. Hot spots that were active on the African Plate at or close to spreading centers which were forming ocean floor at 30 Ma have either died, as the Deccan trap plume source died under Cargados Carajos bank, or have remained active in the same place on ~30 Ma old ocean floor as has the Tristan hot spot. Figures 10 and 18 show that there are no tracks on 30 Ma and younger ocean floor that might indicate the underlying presence of old plumes. Conjugate 30 Ma and younger ocean floor on the North American, South American, Antarctic, Indian and Arabian Plates (Figure 18) must be similarly devoid of the tracks of old plumes. Hot spots such as Ascension, Bouvet and Shona are of special interest because they represent new, very young plumes. Ascension, which lies on the South American plate very close to the spreading center, is geochemically a mixed product of the HIMU and MORB sources (Hart *et al.*, 1992) and Shona appears to be similar (Douglass *et al.*, 1995). I would expect Bouvet to be the same.

Schilling (Schilling *et al.*, 1994) has discerned the neighboring presence of hot spots which lie hundreds of kilometres from the South Atlantic spreading center, such as the Sierra Leone and St. Helena hot spots, in the geochemistry of material that he has dredged from the spreading center itself. This observation has been explained by the existence of a sort of "hose-pipe" connection between the spreading center and the shallow levels of the plume beneath the off-spreading center hot spot (Kincaid *et al.*, 1995). There is no question but that Schilling *et al.* (1994) have seen a clear signal. However, the conventional explanation seems unsatisfying to me because hose-pipes in the mantle would appear hard to create and sustain. The spreading center extracts material from a large volume of underlying and neighboring mantle as McKenzie & Bickle (1988) showed. I can understand (Figure 19) that the catchment volume of a hot spot 500 to 1,000 km away might be affected by drainage to the South Atlantic spreading center, but the reverse seems harder to envisage because the spreading center is so much larger in its magma production rate. I can see the spreading center becoming locally deficient in the kind of material being generated at the plume, but this seems to be the reverse of what Schilling *et al.* (1994) have reported. The possibility that I am left with (Figure 19) is that because the material generated at the spreading center represents the maximum extraction from the MORB source, the nearby presence of a plume which is extracting even the typically small volume of HIMU type products leaves the spreading center generating a less typical MORB and looking somewhat more plume-like. I do not think of this as a very satisfactory explanation but I am forced to it because of unhappiness about "hose-pipes."

The 30 Ma and younger history of the Central Indian Ridge is important to the idea of Africa being at rest over at least part of the mantle circulation. The Central Indian Ridge (Figure 18) is a relatively young structure that, when formed, accomodated an abrupt change of roughly 20 degrees of clockwise rotation of the azimuth of the African/Indo-Australian spreading center plate boundary. Norton & Sclater (1979) ascribed an age of 38 Ma to the formation of this new boundary which was to propagate northward and turn westward into the Gulf of Aden by 10 Ma (Cochran, 1981; Steckler & ten Brink 1986). Patriat & Segoufin (1988), in a more recent analysis which used much new data, also obtained an age of 38 Ma for the initiation of the Central Indian Ridge. The model being tested here predicts a younger age close to 30 Ma. Newly available satellite altimetry (Sandwell *et al.*, 1994; Smith & Sandwell, 1994) should allow reinterpretation of existing magnetic data from the critical area.

The past 30 My history of older African Plate ocean floor

It was suggested, in the introduction to this section, that the older parts of the African Plate ocean floor had behaved over the past 30 My in ways which resembled the behavior of the continent by developing basins and swells (Cloos, 1937; Nyblade & Robinson, 1994) with, in many cases, volcanoes on the swell crests. There is, however, an important difference from the way in which the continental surface behaves in the topographic evolution of the sub-sea level parts of the plate. Erosion and the development of deep weathering and soils had left the continent with relatively little relief prior to ~30 Ma, but on the oceanic part of the plate, erosion plays a small part after elevated features have been eroded to sea level. Old topography declines as the square root of its age until thermally perturbed in one way or another. Without perturbations, this process produces only small changes on very old sea-floor (Nagihara *et al.*, 1996). Areas that were formed relatively high remain clearly elevated.

The Tristan hot spot track, stretching back to Walvis Bay, is the most prominent elevated feature on the old ocean floor of the African Plate and the Seychelles microcontinent, together with the several parts of the Mascarene Plateau built by the Deccan trap source mantle plume, are quite prominent because carbonate banks have been built upon them (Purdy & Bertram, 1993). The Madagascar Plateau extends from the southern tip of that island and appears to have been generated by a plume that died at about 60 Ma [see Storey *et al.* (1995) for another interpretation].

Continental fragments make up at least part of the Agulhas Plateau and the Mozambique Ridge (Ben-Avraham *et al.*, 1995) although as Ben-Avraham and his colleagues suggest, evidence of active tectonism and extremely young basaltic activity indicate a possible role for "buoyancy-related forces originating in the underlying upper mantle" (Ben-Avraham *et al.*, 1995, p. 6199). It therefore seems likely that the northern parts of both the Agulhas Plateau and the Mozambique Ridge may be underlain by plumes of the 30 Ma and younger population.

An extensive region stretching from the South Atlantic spreading center to the coast of southern Africa, between Walvis Bay and the Agulhas Plateau, was identified by Nyblade & Robinson (1994) as ocean floor which showed anomalously shallow residual bathymetry. They suggested the name African Superswell for this region taken together with the contiguous African continental high areas of Southern Africa, East Africa, Ethiopia and the Red Sea Hills.

The oceanic part of the African Superswell contains a number of hot spots from which isotopic ages determined on volcanic rocks have yielded values of less than ~30 Ma. These include the Vema and Discovery seamounts on old ocean floor and the Shona and Bouvet hot spots close to the spreading center as well as the 600 x 300 km elliptical area beneath which the Tristan plume currently lies (Figure 11). Perhaps the whole area is underlain by a large regional plume with local excresences. A large plume with numerous excresences rising from its upper surface may account for much of the anomalous elevation. Attempts have been made to plot tracks between hot spots in this area of the South Atlantic and various Cenozoic and Mesozoic igneous rocks on the African continent. However, recent studies (O'Connor & le Roex, 1992) and application of the rigorous criteria used in this text have shown that, with the exception of Tristan and its plume parents the Etendeka flood basalts and the intrusives of Namibia, this practice does not appear to be justified.

Although shallow elevations of basement are known at several hot spots on the oceanic part of the African Plate, it is hard to be sure whether basement elevations underlie all the active hot spot volcanoes. The reason for this uncertainty is the general paucity of suitable observations, particularly seismic reflection lines. It is clearly the case that there is a basement uplift in the island of Maio, in the Cape Verde archipelago, where old ocean floor outcrops beneath the volcanic edifice (Figure 39; Akhmet'yev *et al.*, 1985). Regional seismic lines (Figure 16 redrawn from Lehner & De Ruiter, 1977) and outcrops of ocean floor in Fuerteventura (Schmincke, 1982) indicate that this is also the case for the Canary islands, although some cross-sections and gravity models (Figure 40) interpret the islands as underlain by huge volcanic volumes which weigh down the underlying lithosphere, as does Hawaii. Seismic lines across the Cameroon volcanic line show that there the oceanic basement has been elevated (Rosendahl *et al.*, 1991). Seismic data from Réunion in the Indian Ocean (Driad *et al.*, 1995) do not clearly show whether the oceanic basement is elevated beneath the island or not. This is compatible with the geochemical observations, including ^{3}He results, which suggest that Réunion is not a typical African Plate, less than 30 million year old, HIMU source dominated, hot spot.

Nyblade & Robinson (1994) made use of generalized data in preparing their map of residual bathymetry around Africa. With higher resolution data sets of sediment thickness, magnetic anomalies and bathymetry as well as with the use of satellite altimetric observations, it should be possible to extend the map of Thiessen *et al.* (1979) of hot spots and high spots to the oceanic part of the African Plate and in that way to provide a simple test of the conclusion that African swell crests tend to be separated by about 700 km.

Nyblade & Robinson (1994, p. 765) found that residual bathymetric elevation was high along all the coastlines of Africa. They attributed this to transitional structure between oceanic and continental lithosphere or to flexural support of the oceanic seafloor as a result of mechanical coupling of the oceanic lithosphere with that of the neighboring continent. The area within a few hundred km of the boundary between continental and oceanic lithosphere is commonly a zone in which tectonic, erosional and depositional activity are concentrated and in which the record of that activity is well-preserved. Not surprisingly, given the tectonic history of Africa over the past 30 My, the African continental margin has responded in striking ways to change. A great regressive episode is perhaps the most spectacular of those responses.

Regression around Africa at ~30 Ma and consequent deltaic progradation

Introduction

The changes that happened close to the margin of the African continent at about 30 Ma are diverse, complex and interlinked. They show themselves in such varied ways as by: (1) the development of unconformities; (2) the progradation of river systems especially of deltas; (3) on-shore, near-shore and deep water erosion; (4) deep off-shore deposition; (5) an episode of salt-tectonics; (6) elevation of shorelines with respect to sea level; (7) scarp retreat; and (8) flexural responses to both the application of loads to the lithosphere by deposition and the removal of loads from the lithosphere by erosion. Many of these topics have been addressed already and in this section, unconformity development at ~30 Ma and deltaic progradation form the main topics of consideration.

There is a serious difficulty with treating the events at and following ~30 Ma around Africa as distinctive and unusual. The problem is simply that there is a well-known mid-Oligocene, or ~30 Ma, unconformity that has been recognised around all the shorelines of the world's oceans (Haq *et al.*, 1987). The deltas of many of the world's great rivers have also generally prograded substantially during the past 30 My. Bartek *et al.* (1991) showed that this familiar global mid-Oligocene unconformity can be plausibly linked to the removal of a large volume of sea water from the world ocean by the initial formation of the East Antarctic ice sheet. If the volume of the East Antarctic ice sheet rose rapidly to an amount close to its present volume and if that volume has remained roughly constant ever since, then sea level world-wide would have declined by ~50 m and that deficit would have been sustained. Although there is no consensus about the early size, stability and even the persistence of the East Antarctic ice sheet, I am impressed by the arguments and by the models which indicate that the isolation of Antarctica by the circum-Antarctic Ocean current would have served to stabilize the East Antarctic ice sheet once it had formed.

What then can be said about special conditions around Africa? My initial answer is that having looked at regional seismic lines from many parts of the world, I find myself impressed with the intensity of the mid-Oligocene unconformity, and especially the developments of submarine canyons, that are clearly displayed on African offshore seismic lines. This observation is anecdotal and not a readily quantifiable statement, but I hasten to point out that I am not alone. McGinnis *et al.* (1993), for example, have emphasised the importance of the ~30 Ma unconformity around the shores of Africa. My suggestion is that the mid-Oligocene event achieves a general intensity around the shores of Africa that is reached only locally around other shores. In the future I hope that it may prove possible, with better temporal resolution, to distinguish the circum-African phenomena from the world-wide phenomena. Here I will simply draw attention to some examples of the occurrence of the mid-Oligocene unconformity around Africa and point out how those occurrences fit in with onshore observations related to Africa having come to rest over the mantle circulation.

Mid-Oligocene unconformity around Africa

Some of the earliest publications which showed the importance of a major unconformity that formed around Africa in mid-Oligocene times at "~29 Ma" were by the Exxon Production Research Laboratory pioneer researchers into sequence stratigraphy. Parts of offshore seismic lines, both normal and parallel to the African coast, are shown in such papers as Vail *et al.* (1977), Schuepbach & Vail (1980) and Hardenbol *et al.* (1982). The illustrated lines are from off the coast of Morocco, perhaps near the Essouira Basin. An example is reproduced in Figure 41. Not only is the ~30 Ma unconformity clearly displayed, but so is a major submarine canyon. These canyons began to be filled after ~22 Ma, that is early in Miocene times. I have drawn Figure 42 using data from Brognon & Verrier (1966) to illustrate in a sketch map the appearance of some other members of this widely distributed population of submarine canyons. The canyons I illustrate form part of the Kwanza Basin of Angola. These canyons too were cut during the Oligocene and filled in the Early Miocene. A salt tectonic episode beginning at ~30 Ma off the coast of Angola is attributable to the new uplift of the continent and was probably initiated as a result of salt moving into newly cut submarine canyons (Duval *et al.*, 1992; Lundin, 1992). Other places where the circum-African mid-Oligocene unconformity has been illustrated include figures in Clifford's (1986) review article on African petroleum geology, his figures 14, 26, 33 and 38. More recent papers that address the unconformity include Teisserenc & Villemin (1990) and Seranne *et al.* (1992), who both cite examples from Gabon, and McGinnis *et al.* (1993) who show an unlocated line "from the Congo margin."

Many published seismic lines also show that a great volume of sediment has been deposited in deep water above the mid Oligocene unconformity off the coast of much of Africa. Deposition above the unconformity began perhaps as early as the later Oligocene at ~25 Ma. The sediments above the unconformity have buried older deep water sequences that were deposited at the Atlantic-type margins of Africa between the termination of initial rifting and the end of the Eocene. These continental margins are places of low heat flow and not much sediment had been accumulated in the post-rift sequences of the margins by the end of the Eocene. For those two reasons potential oil and gas source rocks in the sequences below the mid-Oligocene unconformity are likely to have remained thermally immature. Burial by great thicknesses of Late Oligocene and younger sediments may well have driven suitable material through the oil window to produce oil that has since been able to migrate up dip toward the continent.

McGinnis *et al.* (1993) have suggested an important role for deep-sea erosion off West Africa, such as has been well documented off the east coast of North America since the Early Oligocene at ~34 Ma. Similar deep sea erosion has been recognised at continental margins in many parts of the world and has been interpreted to be a consequence of the onset of glacial conditions. The development of polar pack ice leads to the formation from immediately below of flowing, cold, bottom-water masses which travel toward the equator with considerable erosional power. The development of cold water masses and deep-sea erosion certainly happened at about the same time as world-wide sea level lowering in Oligocene times (Kennett & Shackleton, 1976) but I am reluctant to attribute to them responsibility for much of what has happened around the coasts of Africa for two reasons. First, the evidence of elevation of the continent from a generally low-lying level, coupled with the beginning of scarp-retreat and its flexural responses (Figure 26), suggest another control for the development of both the erosional and the depositional processes that developed around Africa at this time. Secondly, in places where there is strong evidence that there has been deep water erosion, such as the eastern coast of the United States off Cape Hatteras or the coast of West Africa southeast of Abidjan, eroded escarpments persist in the deep water today. To attain the condition illustrated in the African coastal seismic lines (McGinnis *et al.*, 1993, their figure 4), which show 2 seconds or more of sediment deposited in deep water above the 30 Ma unconformity, would require deep water erosion first to switch on and later to switch off. This is not what has happened. Glacial conditions persist today in Antarctica. They have not turned off so that any bodies of deep bottom water that have been set up persist. If such deep water masses were eroding the coast of Africa at 30 Ma they would still be eroding it today and the thick sediments in deep water that overlie the deep water unconformity could not have accumulated.

The question of what currents were flowing around the coasts of Africa about 30 million years ago is an interesting one. Smart & Ramsay (1995) have shown that a salt-rich deep current was flowing out of the Tethys into the Indian Ocean until about 16 Ma. I suggest that this current operated as a western boundary current that flowed south, hugging the eastern coast of the African continent. According to Smart & Ramsay (1995), waters from the same North Atlantic source flowed south at depth along the coast of northwestern Africa. Along the Guinea coast the Guinea current, and its possible counter-current (Burke, 1971), are likely to have continued to flow as they had done for nearly 100 My. At the southern end of the continent, the Benguela current has been suggested to have been initiated by ~34 Ma (Siesser, 1978) when the East Antarctic ice sheet first formed. This cold current, which flows from south to north on the western side of southern Africa, is the most likely candidate to have initiated deep water erosion along the west coast of Africa during the mid Oligocene but I am not aware of any evidence that it has played such a role.

Ben-Avraham *et al.* (1994) have reported a change in deepwater circulation off the east coast of South Africa in the relevant time interval. Antarctic bottom water flows eastward through the Agulhas passage and formerly continued to hug the continental margin flowing northward into the Natal valley. This water now flows in a more easterly direction into the Transkei Basin. Ben-Avraham *et al.* (1994) attribute the change to the uplift within the past 30 million years of the Mozambique Ridge microcontinent. There have certainly been fluctuations in the relative roles of Antarctic and Arabian Sea waters close to the east coast of southern Africa (see Caralp *et al.*, 1993, for Late Pleistocene and Holocene fluctuations) but evidence of deep sea erosion by currents has not been emphasised.

Sea level lowering, such as happened as a result of the formation of the East Antarctic icecap, has a variety of consequences in marine sedimentary environments which have come to be very well understood through the development of sequence stratigraphic analysis (Vail, 1987). The occurrence of the kind of changes that have come to be expected have been well documented around the shores of Africa (Brown *et al.*, 1996). Only one such change concerns us here. That is the progradation of the deltas of the great rivers of Africa. These rivers have received increased sediment supplies as a result of the elevation of the continent. Some rivers show the consequences in particularly spectacular ways.

Deltas of the great rivers of Africa

The Niger delta is the largest of African deltas and its history is particularly well known because of the oil industry that it supports (Doust & Omatsola, 1990; Whiteman, 1982). From ~80 Ma until ~35 Ma, a major river system flowed southwestward along a valley that occupied the site of the Benue Rift. The mouth of that river lay in the general area of the present apex of the Niger delta. The Benue Rift had been formed as part of the widely distributed African rift population which originated ~140 Ma, but its tectonic and igneous activity had long ceased by the end of Eocene times at ~35 Ma. Since the beginning of the Oligocene at ~34 Ma, the Niger delta has prograded from its location at the then mouth of the Benue valley for more than 200 km to the southwest. The delta has also broadened so that it now has a width of ~500 km (Figure 43 modified after Whiteman, 1982). Throughout the past 30 My, the Benue valley has been the main source of water and sediment supply to this growing "Niger" delta. During the Late Quaternary, the Niger river has frequently died out in the swamps of the "inland delta" (Reclus, 1888, p. 282) region of Mali. At those times the Niger does not contribute at all to the great delta at the coast which bears its name. It is not known when the Niger first began to intermittently spill water and some sediment into the Benue system and contribute at all to the growth of the great delta. It may well have been only in the past few million years.

The sediments which have reached the Niger delta from the Benue drainage system are dominantly derived from the basement uplifts beneath the volcanic areas of the onshore part of the Cameroon line (Figures 6 and 16), although right bank tributaries draining such discrete elevated areas as the Jos Plateau (Figure 15) also contribute. I suspect that the increase in sediment supply to the delta, recorded in its forward surge and its broadening in width, has come mainly from the generation of newly elevated areas associated with the continent-wide upsurge in volcanism at 30 Ma. However, it is worth remembering that northern Cameroon was the site of emplacement of 17 discrete granitic bodies between 65 and 30 Ma (Lee *et al.*, 1994). Some elevated ground, which would have been likely to supply sediment to the older delta, must have been associated with those bodies.

Filled submarine canyons associated with Quaternary low stands of sea level have long been recognised as elements in the architecture of the Niger delta (Burke, 1972). Filled canyons within the Niger delta sediment pile are now well-known (Doust & Omatsola, 1990, their figure 4). The oldest, the Opuama canyon complex, is associated with the low-stand of the great mid-Oligocene unconformity.

The Niger delta represents an enhancement during the past 30 My of an already existing delta at the mouth of a long-lived river system. The Nile delta is different because there was no pre-30 Ma major river draining to the north and reaching the Mediterranean close to the present site of the Nile delta. The origin of the Nile and the establishment of its delta some time between 34 Ma and 25 Ma were discussed earlier under the heading "The Continental Surface: Northeastern Africa." Figures 44 and 45 (after Said, 1993) and Figure 46 (from Elzarka & Radwan, 1986) show how the delta has prograded since 30 Ma over the Late Cretaceous and Paleogene carbonate and sandstone shelf sediments of the northern margin of the continent. Canyon development in the Nile delta is best known from the Messinian record (at ~6 Ma) when the Mediterranean had evaporated to dryness and a colossal subaerial canyon was cut (Hsü, 1972).

Large deltas have developed during the past 30 My at the mouths of the Limpopo and Zambesi rivers (Figure 47) which, like the Benue valley, are located over old rifts. The Zambesi valley was an active rift at about 200 Ma and, although evidence for Cretaceous igneous and fault activity is not strong in the Zambesi Rift itself, alkaline intrusives were emplaced during the Early Cretaceous in the associated Shire (or Chire) Rift of southern Malawi which is continuous along strike with the Urewa Rift in Mozambique close to the Zambesi river mouth (Figure 47).

Up to 3 km of deltaic and delta fan sediments forming the "paleo Zambesi delta" of De Buyl & Flores (1986, figure 6 ane p. 417) were deposited in a Zambesi delta depocenter close to 36^{o} East and 20^{o} South during Cretaceous times between ~140 and 65 Ma (De Buyl & Flores, 1986, figure 18). How important elevations on either flank of the present Zambesi were as sediment sources for the Cretaceous "paleo Zambesi delta" is unclear because sediment could also have been derived from rivers draining elevated areas farther north related to the Shire Rift. Perhaps source areas in both regions were important, but wherever the sediment came from, the Cretaceous "Paleo Zambesi" was an important river in southern Africa's Cretaceous drainage.

That appears to have changed during the interval 65 to 34 Ma when much of Africa appears to have been low-lying. Rocks of the shallow water part of the offshore Zambesi mouth depositional system were eroded away at the time of formation of the mid Oligocene unconformity but deeper water resedimented carbonate rocks, some of which show graded bedding and are of Paleocene age (65 - 53 Ma), are interpreted by De Buyl & Flores (1986, their figures 6, 18 and 19) as indicating the up slope existence of a contemporary reef tract where the "paleo-Zambesi" delta had formerly been. Comparable reef deposits are known from farther north in offshore Mozambique and from the Bengurea high, farther south, at about 22^{o} South (De Buyl & Flores, 1986, figure 19).

The mid-Oligocene unconformity, the prograding Miocene and younger Zambesi delta and the deep water fan deposits in front of that delta are magnificently displayed in the regional seismic lines published by De Buyl & Flores (1986, figures 6 and 10). The "paleo Zambesi," which had been inactive or quiescent between 65 and 34 Ma, surged into a new life. Kolla *et al.* (1991) reported that the Zambesi river has built a broad delta-platform between 75 and 100 km into the Indian Ocean and that the first significant and rapid progradation occurred since the mid-Miocene. The earliest of the Neogene sequences are thicker in the south but that has changed as the Zambesi river shifted from south to north in time. Figure 48 (modified from Droz & Mouqenot, 1987) shows that the upper Mozambique fan at the foot of the Zambesi delta was fed from the north in Oligocene to Early Miocene times at the time when the Kerimbas and Lacerda Rift Basins were newly activated (Figures 32 and 34). The area of the Zambesi delta became dominant as a sediment source area in Middle Miocene to Pliocene times (~14 - 3 Ma). During the Pleistocene, a new submarine-fan valley has become active. Droz & Mougenot (1987), making use of unpublished work by Lafourcade (1984), depict the progradation of the continental shelf in the area of the Zambesi delta as exceeding 100 km and show the shelf buildup extending over more than 400 km along shore to reach the Inhambane hinge line of De Buyl & Flores (1986, figure 2). That hinge line separates the domain of the Zambesi delta from that of the much smaller Limpopo delta.

In its higher reaches, the Limpopo flows along a rift that was active in Jurassic times and, in its lower reaches, it flows close to the foot of the Lebombo monocline which was active at the time of the separation of Africa from Antarctica. De Buyl & Flores (1986, figures 9, 13 and 14) show, from offshore seismic reflection lines, that rifting was active into Cretaceous times beneath the present site of the Limpopo delta. A great unconformity, presumably the mid-Oligocene unconformity, separates Miocene limestones and sandstones from Campanian shales in a well on the offshore southeastern flank of the Limpopo delta (De Buyl & Flores, 1986, figure 17). Two seismic lines (De Buyl & Flores, 1986, figures 13 and 14) suggest to me that there has been no preservation of earlier Cenozoic sediments in the Limpopo delta area. A seismic line in the region of the delta axis, oriented parallel to the coast and only about 10 km offshore, shows about one second of Miocene and younger sediment while a parallel line about 50 km farther offshore shows about half a second of accumulation. Altogether the total mapped thickness of sediments above the top of the Karroo strata in the area of the Limpopo delta amounts to only about two seconds (De Buyl & Flores, 1986, figure 9). By contrast, the same map shows more than five seconds of sediment above the Karroo reflector in the main depocenter of the Zambesi delta. The Zambesi drainage appears to have been more important to southern Africa than that of the Limpopo, both in Cretaceous times and in the interval since ~30 Ma. This difference may be tectonically controlled but could be a response to the fact that the Limpopo, particularly during the past ~30 My, has drained a more arid region.

The young deltaic system of the Limpopo is typical of those of many of the rivers of Africa which reach the ocean in showing progradation in Miocene and younger times. The nearby Tugela river (Dingle *et al.*, 1983) is another such example. Perhaps the rivers that behave in a different style are more interesting because they are unusual. The Orange river and the related Cape Basin have shown a substantial decline in sediment accumulation in the past 34 My since the beginning of the Oligocene (Rust & Summerfield, 1990). The distinctive character of the area draining to the Cape Basin has been pointed out by Siesser (1978) who suggested that the Benguela current contributed to the onset of aridity within the catchment area of the Cape Basin. This control appears to have been radical enough to overwhelm any increased sediment transport related to the erosion of newly uplifted southern Africa (Partridge & Maud, 1987).

The Zaire river is anomalous in having no delta. This is generally recognised to be a consequence of Zaire Basin internal drainage having been captured at some time within the past 30 My by a short stream draining to the coast. The captured great river cuts through the high ground at the continental margin on the western rim of the basin (Figure 30). Sediment from the newly formed steep river eroded the Congo (Zaire) submarine canyon at times of low sea level and supplied material to the huge deep sea fan at the foot of the continental slope. The estuary at the mouth of the Zaire river is itself a relatively unusual feature around the coasts of Africa. Its persistence until today shows that the Zaire has not been able in the past ~10 ka to transport enough sediment to fill an estuary which occupied the head of the Zaire submarine canyon and was most recently active during latest sea level lowstand at ~20 ka.

Fluctuations in sea level have a strong influence on delta and delta fan structure and these are among the most prominent of a wide range of environmental changes which climatic and related variations have induced in Africa over the past 30 My. These environmental changes have been much studied and interest here is restricted to posing the question: How have the peculiar structural and topographic features of young Africa, a continent of basins and swells and of rifts, influenced the way in which environmental changes have operated on both the continental and oceanic parts of the African Plate?

Environmental change on the African Plate: The past 30 My

Scientists in Africa were early in appreciating the need for a continent-wide perspective in understanding environmental change. I remember participating in a meeting on "The evolution of the West African environment" at which this broad perspective was emphasised in Accra as early as the late 1950s. The early consciousness of the importance of the paleoenvironment was due to the influence of three remarkable scientists, all of whom took a broad approach to the Earth Sciences and particularly to environmental change. These scientists, who lived near three corners of the continent, were: Louis Leakey in Nairobi, Theodor Monod in Dakar and E.M. Van Zinderen Bakker in Bloemfontein. Although we were early aware of the importance for Africa of the development of the polar ice sheets (Van Zinderen Bakker, 1976), changes in oceanic circulation were not so greatly emphasised. Here I will briefly consider the oceanic, and afterward the continental, environment especially as they relate to Africa's distinctive structure with swells and basins of inland drainage as well as its great rift system.

Environmental change in the oceans around Africa over the past 30 My

The East Antarctic ice sheet appears to have formed by 30 Ma (Bartek *et al.*, 1991) and as a consequence, the Benguela current had started to flow (Siesser, 1978). It seems probable that aridification of the Namib desert had set in as a further consequence but the age of the oldest Namib sediments, which record the onset of aridification, has proved difficult to determine (Siesser, 1978). One thing that does appear clear is that aridification began in the desert areas of southern Africa much earlier than it did in the Saharan region in the north where the onset of desertification appears well recorded by an increase in the windblown dust content of ocean drilling project cores at ~3 Ma (Tiedemann *et al.*, 1989) lately refined to ~2.8 Ma (deMenocal, 1995).

Before ~3 Ma, which was roughly the time of onset of northern hemisphere glaciation, the circum-African oceanic environment was much influenced by the Antarctic ice sheet. Southern Component deep water dominated in the Atlantic (Woodruff & Savin, 1989) in the Early Neogene (~22 to 15 Ma). Evidence of a short-lived role for Northern Component water has been discerned in the South Atlantic for the interval between 19.5 and 16.5 Ma. Smart & Ramsay (1995) and Woodruff & Savin (1989) identified Atlantic benthic foraminiferal faunas with ages between 14 and 8.5 Ma which they associated with a water mass similar to modern North Atlantic deep water. The same two studies showed that warm saline bottom waters flowed out of the Tethys into the Arabian Sea until about 15 Ma. Southern Component deep water may have dominated deep water farther south in the Indian ocean (Ben-Avraham *et al.*, 1994).

The closure of the Tethys in the Zagros Mountains of Iran at about 15 Ma (Figure 37) may have been an important influence in helping to establish the present conveyor belt circulation of the world ocean (Woodruff & Savin, 1989) with surface flow from the northwestern Pacific and an important role around much of Africa for both Antarctic bottom water and North Atlantic deep water.

Van Zinderen Bakker (1976) introduced a simple conceptual model of how winds, ocean currents and onshore rainfall patterns might have differed in Africa between glacial and interglacial times. A map depicting an updated version of his model is shown in Figure 49 (from

Kroepelin, 1994, after Shaw, 1985). The model is based primarily on the idea that, when there are ice sheets at both northern and southern poles during northern hemisphere glacial times, the northern hemisphere zonal wind belts occupy more southerly latitudes. In interglacial times, when there is only an ice sheet at the South Pole, the same wind belts lie farther north. I have plotted the positions of seven interior basins among the swells of Africa on Kroepelin's map to show how climatic variations influence deposition and erosion in the various basins in different ways.

Conditions between ~34 Ma and ~3 Ma, when northern hemisphere ice sheets first formed, resembled the interglacial state shown in Figure 49. Monsoonal conditions were important but a good question is: When did the Asian monsoon become as important for the Arabian Sea and the climate of East Africa as it is today? An influence on intensity has been considered to have been the elevation of the Himalaya and the Tibetan Plateau (Ruddiman & Kutzbach, 1991). Important changes in the style of the Indian tectonic collision, which may have led to greater elevation of those mountain belts, are known at ~20 Ma. Events related to a further increase in elevation are recorded between ~10 Ma and ~5 Ma, but there is no single tectonic event within the past 10 My which could be linked to the onset or to the intensification of the Indian Ocean monsoon (Harrison *et al.*, 1992). A recent study (deMenocal, 1995) has used temporal variations in windblown dust concentration from Ocean Drilling Project cores obtained in the Arabian Sea to identify controls on climatic variability in the region (deMenocal, 1995). To quote from deMenocal (1995, p. 55): "Before 2.8 Ma, the spectra of the dust records indicate that subtropical African climate varied primarily at precessional (23- to 19-ky) periodicities; this mode of variability extended [back] at least into the mid-Miocene (~12 Ma)." This evidence from the Arabian sea apparently shows that the Asian monsoon was a significant factor in the climate of Africa from at least as far back as ~12 Ma. It is tempting to consider that the ~15 Ma changes associated with the closure of the Tethys in Iran might also have influenced the Arabian Sea monsoon.

Figure 49 shows that a second monsoonal system is likely to have been important on the western side of the continent. Evidence from a site offshore of Dakar (Leroy & Dupont, 1994), to which deMenocal (1995, p. 57) draws attention, indicates that humid taxa of rain forest affinity were replaced in that region by arid-adapted taxa between 3.2 and 2.6 Ma. This event perhaps corresponded to a reduction in the extent of influence of the West African monsoonal system at the time of inception of northern hemisphere glaciation.

Other environmental changes in the oceans around Africa during the 30 My before the onset of northern hemisphere glaciation may have included an increase in hemipelagic contributions from the African continent. The development of very numerous submarine canyons and the large accumulation of sediments at the foot of the continental slope, especially in such great deep sea fans as those of the Zaire river, the Niger, and the Mozambique fan, would surely have led to an increase in fine material dispersed over the ocean floor.

Changes on the ocean floor from ~2.8 Ma record fluctuations related to those of the northern hemisphere ice sheets. African climate variability, as indicated by concentrations of eolian dust and other indicator materials in Ocean Drilling Project cores (deMenocal, 1995), shows step-like increases in aridity and shifts in variability near 2.8, 1.7 and 1.0 Ma. The African climate became periodically cooler and drier after 2.8 Ma "as a result of dynamical effects related to the development of cold glacial North Atlantic Sea surface temperatures. This effect was further amplified after 1 Ma, following the increase in the duration and magnitude of high-latitude glacial cycles" (deMenocal, 1995, p. 57). Because paleoclimatic evidence shows that precessional forcing of monsoonal climate persisted throughout the entire Plio-Pleistocene (about the past 5 My), deMenocal considers that the high latitude influence on the African climate was a "superimposed but primary factor" (1995, p. 57).

Environmental change on the African continent over the past 30 My

At ~30 Ma, uplift, volcanic activity and rifting broke out over what had been a low-lying, tectonically and volcanically quiescent Africa. Climatic conditions approximated to those shown in the interglacial condition of Figure 49. The Benguela current had begun to flow and aridification had begun in the southwesternmost part of the continent.

Quantifying the amount of uplift is hard. Fission track studies (Foster & Gleadow, 1996) indicate that less than 2 km of material has been eroded from most of the continent over the past 30 My. Nyblade & Robinson (1994) concluded that the average excess elevation of Africa is now about 500 m. Few high areas exceed more than 3 km in elevation but a large area reaches 2km. There is little evidence as to how fast elevation happened after it had begun at about 30 Ma. On the other hand, the extent of the areas involved in uplift are better known. In total they amount to about half the area of the continent.

Limited elevation over wide areas probably led to a widespread decrease in surface humidity and perhaps, because of an increase in seasonal pressure variation over elevated areas, to an intensification of monsoonal conditions. The newly formed basins among the swells are likely to have responded to climatic change in different ways (compare Figure 49 with Figure 6). The Kalahari Basin would likely have been dry with perhaps episodic lakes. The Zaire Basin would have been wet for most if not all of the year and this could have helped in the early generation of a throughgoing drainage system that reached the ocean perhaps by Miocene times. The Chad Basin would have been seasonally wet and dry and this might also have been the case in the inland delta region of the Niger drainage system. The northwestern Sahara region, which presently centers on the Taoudeni Basin, may well have been seasonally moist and have drained to the north coast of Africa.

Northeastern Africa and its then Arabian promontory were probably dry but the east coast of the continent north of the equator would have been influenced by the Asian Monsoon with January rains and July aridity. The East African Rift System would have been likely, as now, to have generated some highly elevated areas which, again as now, would have generated local, unusually moist climates.

Dating the oldest sediments in African lacustrine basins has proved difficult, but lakes were probably widely developed from about 30 Ma both in the large interior drainage basins and within the rift system. The benign character of the lakeshore environment for hominid evolution with perennial water and abundant vegetation is clear.

Nevertheless the climatic change at ~2.8 Ma, which is so well documented from evidence on the ocean floor (deMenocal, 1995), has been deemed of critical importance in hominid evolution. Several authors have had the idea that climatic change has been important at various times in human evolution (Vrba *et al.*, 1996) and the onset of aridity as an influence on evolution has been a common theme. For example, Stanley (1992) developed an argument which linked the pronounced encephalization of *Homo* to adaptations "shortly after the onset of the modern ice age about 2.5 Ma." He recognised this as a time of contraction of forests in Africa and suggested that the contraction caused a crisis that led to the evolution of Homo by compelling some australopithecine populations to adopt a fully terrestrial existence. Caution in interpreting a fossil hominid record which is extraordinarily sparse is urged by deMenocal (1995, p. 57) who was, however, able to show that major changes in hominid evolution appear to have coincided with the three steps in increased aridification that he has recognised from deep sea drill cores. These steps took place at ~2.8 Ma, ~1.7 Ma and ~1.0 Ma (deMenocal, 1995, his figure 6).

Bad though the record of fossil hominids is, especially over the critical interval since 5 Ma, it is better in Africa than elsewhere. Although this seems likely to be primarily because *Homo* evolved in Africa, there is another reason why the record is as good as it is. The tectonics of the East African Rift System, which is itself a critical element in the distinctive tectonics that have developed in Africa since 30 Ma, have produced faults which have exposed the fossiliferous strata from which the fossil hominids have been extracted. A further happy circumstance is that rift volcanism has erupted tuffs that are interbedded with fossiliferous strata. These tuffs frequently contain potash feldspars which have contributed ages to high resolution chronologies. One other factor which has contributed to the rapid improvement of the still very sparse record has been an increase in effort in recent years. WoldeGabriel *et al.* (1994) provide a splendid example of the kind of collecting, environmental interpretation and age determination which are becoming typical.

Although many of the critical hominid fossils have been collected within the East African Rift System, modern collecting has yielded the first australopithecine from outside the rift in northern Africa. Brunet *et al.* (1995) have found material at Korotoro, on the eastern flank of the Chad Basin. A good question is: How does such a fossil appear at outcrop. In the lake basin, fossils as old as ~2 Ma might be expected to lie at great depth. The reason is likely to be either that continuing uplift on the Ennedi and Tibesti flanks of the Chad Basin has exposed the older rocks to erosion or that climatic change has led to downcutting in an early deposited basin flank sediment wedge. Some combination of both processes could also have operated.

Northern African basins, of which the Chad Basin is the best example, are likely to have responded to the onset of northern hemisphere glaciation by becoming dry (Figure 49). In Chad, and in the area immediately south of the Sahara, over the past 2.8 My, alternations between the two states depicted as glacial and interglacial in Figure 49 are expected. Only the latest half-cycle extending back 20 ky is at all well known (Burke *et al.*, 1971; Talbot & Johannessen, 1992). There was a dry phase south of the Sahara about 20 ka at the height of the glacial maximum but the changes between then and now do not represent just a simple increase in rainfall. The Chad Basin filled to its brim and spilled over the Bongor spillway about 10 ka. The Niger stopped flowing at the inland delta at about 20 ka. On the model of Figure 49, the Taoudeni Basin would have been expected to be wet about 20 ka as westerly winds crossed the continent south of the Atlas Mountains. The head waters of the Nile would have received less moisture and there is ample evidence that it was not at that time a through flowing river. The Zaire Basin would have remained wet as always. Figure 49 indicates that during northern hemisphere glacial maxima, the Kalahari Basin may have received more moisture, both in January and July, than at any time since the Benguela current started to flow.

In summary, environmental conditions over the African Plate were changed somewhat from those before 30 Ma by the new tectonism, especially by the establishment of the broad basins and swells, but the establishment of the East Antarctic ice sheet by 34 Ma probably had a much greater influence. The Zagros collision at ~15 Ma may also have had some effect, but the big environmental changes in Africa came at ~2.8 Ma with the onset of northern hemisphere glacial cycles. Hominid evolution may well have responded directly to changes related to the onset of northern hemisphere glaciation. Tectonics in the East African Rift System have exposed the rocks containing the critical fossils.

Much interest focuses today on the most recent changes in the African environment. In contrast to the relatively constant temperatures for the past 10,000 years indicated by the records of two recent Greenland ice-cores, Africa has experienced great climatic variations over the past 10,000 years. Cooperative studies have, for example, shown that between 8,500 and 6,000 BP the Saharo-Sahelian boundary separating desert from relatively wetter environments lay ~500 km north of its present position (Petit-Maire, 1995).

Conclusion

I have repeatedly suggested in the course of this text that the African Plate has been at rest for the past 30 My with respect to the underlying mantle circulation, or at least with respect to that part of the circulation represented by the population of mantle plumes below the plate. One excuse for writing at such great length with much repetition is that last time that idea was enunciated a quarter of a century ago, Wilson and I used less than 1,000 words and nobody took any notice (Burke & Wilson, 1972). I have used the idea of "Africa at rest over the mantle circulation" as a basis for a discursive review of phenomena that have affected the African Plate during the past 30 My. I have adopted the approach which I have used for two reasons. One is that the school in which I was brought up to write research papers (in which evidence is gradually marshalled and finally, at the very end of the paper, a hypothesis emerges) seems to me to constitute both a dreary baconian approach and cruel and unusual punishment to the reader (Oliver, 1988).

A more serious reason is that the approach I have used puts my hypothesis at risk. All hypotheses ultimately experience the fate of being proved wrong. Fortunately, this is sometimes achieved by being accomodated into a broader picture as the hypothesis of continental drift was accomodated within plate tectonics. One advantage of a specifically expressed hypothesis is that it can be tested with a view of being rejected in favor of something better. Of course we all hope that our hypotheses will have a long life but we have to be ready to let them "die for us" (Popper, 1980). With that kind of mortality in mind, I will first summarize the properties of the African Plate that seem to me best accomodated within the model that I have constructed. I put no references in this list because they are liberally distributed throughout the foregoing text. I then explain what I think happened at 30 Ma and subsequently, and finally, I suggest some ways that have occurred to me for testing the model.

Properties that a model of African Plate behavior over the past 30 My must satisfy

Properties that establish the prior condition of the plate

1) The African Plate has moved little with respect to the spin axis of the earth since ~200 Ma.

2) What motion there has been mainly happened between ~100 Ma and ~30 Ma. That motion consisted largely of an ~10 degree clockwise rotation about a pole at ~0^{o} East and ~0^{o} North.

3) During the past ~200 Ma interval, other continents have moved away from Africa so that new oceans have formed around the continent and convergent boundaries have commonly developed on the sides of the departing continents that are remote from Africa. For that reason no subducted slabs have been emplaced into the region of the mantle deep beneath the area now occupied by the African Plate for the past ~200 My. The occurrence of a large low-velocity volume in the deep mantle far beneath the surface of Africa may be related to the absence from that region of ancient subducted slabs.

4) Africa was a low-lying continent with a humid climate over most of its surface between ~ 65 Ma and 30 Ma.

5) The plume that constructed the Walvis Ridge, which is also known as the Tristan and Gough plume, and the Deccan trap source plume were the only two long lived mantle plumes producing large volumes of hot spot volcanic rock on the oceanic parts of the African Plate between ~60 Ma and ~30 Ma. The Walvis Ridge is the only hot spot track on the African Plate that ends at a currently active hot spot.

6) There was not much igneous activity on the continent of Africa between ~65 Ma and 30 Ma. The small amount of volcanism that did occur was sporadic and episodic. An area in northern Cameroon with continuous activity throughout the whole 65 Ma to 30 Ma interval

constitutes the exception to this observation. It is perhaps significant that the Cameroon area, which is one of continuous igneous activity throughout the past 140 My, is in the general region of the poorly defined pole that describes the rotation of Africa with respect to the spin axis during the ~100 Ma to 30 Ma interval.

7) The only indication from occurrences of hot spot volcanism on the continent of Africa of a possible track generated by an underlying plume is that provided by volcanic activity in southern Ethiopia between 45 and 35 Ma which may be linked to activity at Lokitipi, in Turkana, starting at 35 Ma and continuing in that area ever since.

8) Relatively little clastic sediment was eroded from the low-lying African continent between 65 Ma and 30 Ma. The Benue and the Trans Tswana were among the main rivers of the continent. There was no Nile during that interval.

9) The southern part of the African Plate experienced a change in peripheral oceanic circulation between 40 Ma and 34 Ma with the opening of the Drake Passage and the establishment of the East Antarctic ice sheet. The Benguela current began to flow up the west coast of the continent and Antarctic bottom water flowed eastward through the Agulhas passage and up the east coast.

Properties indicating the arrest of the African Plate over mantle plumes as expressed by the surface expression of mantle plumes at ~30 Ma and the subsequent development of plume lithosphere interaction.

1) Volcanism broke out roughly simultaneously over widely dispersed areas on the African Plate at ~30 Ma.

2) Volcanism from the Tristan plume, which had been generating the lengthening Walvis Ridge for the previous 100 My, has persisted in the same restricted 500 km x 250 km area since ~30 Ma without further progression.

3) Volcanism from the Deccan trap mantle plume source ceased at ~30 Ma at the site of Nazareth bank and Cargados Carajos bank.

4) Volcanism at many of the individual intraplate hot spots on the African Plate has continued episodically since ~30 Ma indicating the persistence of mantle plumes beneath those hot spots.

5) The persistence of volcanism in the same places for up to 30 My shows that the plumes beneath the hot spots have not moved with respect to each other.

6) Volcanism has propagated over some or all of the past 30 My at several of the individual hot spots on the African Plate. These propagations cover distances of less than a maximum of 400 km and show no consistent azimuth. Therefore, not only have the plumes remained

fixed with respect to each other, they have also remained fixed with respect to the overlying African Plate at the spatial resolution of the size of a typical hot spot, which is about 200 km in diameter.

7) Hot spots that have developed on the African Plate within the past 30 My are in many cases, and perhaps in all cases, associated with substantial basement uplifts. This is in complete contrast to such hot spots as Hawaii and Iceland which are associated with huge volumes of basaltic volcanic rocks that form a load that bows down the underlying lithosphere.

8) Thirty Ma and younger African Plate hot spots are associated with relatively small volumes of erupted volcanic rock. Exceptions are: the Afar and associated Ethiopian traps, both erupted from the Afar plume, and to a lesser extent Samburu, beneath the Kenya Rift and perhaps the island of Réunion. Tristan is also different but this is expected because it is not part of the 30 Ma and younger population.

9) On the oceanic part of the African Plate, basalts that have been erupted as part of the 30 Ma and younger population are dominantly of HIMU derivation. This may also be the case for the hot spots on the continental part of the plate and even for the hot spot volcanic rocks associated with the East African Rift System. The volcanic rocks of the 30 Ma and younger population that have been erupted onto the continent also show evidence of the involvement of the EM1 and EM2 sources. This may reflect contributions from reaction with the between plumes of HIMU source material and overlying fertile mantle lithosphere.

10) Young volcanic rocks of the African Plate associated with the 30 Ma and younger hot spot population have ^{3}He signatures below that of MORB. The Afar constitutes a spectacular exception and Réunion a less extreme exception.

11) The low ^{3}He signature, HIMU source dominance and the small degree of partial melting indicated by the small volumes of volcanic rock are compatible with the idea that the plumes responsible for the 30 Ma and younger population of African hot spots arise from no deeper than the base of the transition zone at 670 km. The Afar is the likely exception.

12) Much of the African Plate is anomalously elevated compared with the rest of the earth away from plate boundary zones. This elevation takes the form of elliptical swells of various sizes in both continental and oceanic areas.

13) Swells on the African continent that overlie continental lithosphere which had been reactivated in Pan-African times are generally crested by hot spot volcanic rocks of the 30 Ma and younger population, although this volcanism is not restricted to the crests of the swells.

14) Swells on the African continent that overlie cratonic continental lithosphere, typically one thousand million years in age or older, have no volcanism on their crests. The only place where 30 Ma and younger volcanic rocks have been erupted through rocks this old is in the northern third of the western rift section of the East African Rift System.

15) Swells, elevated areas elliptical in plan, are identifiable on both the continental surface and the ocean bottom surface. These swells began to be elevated at ~30 Ma. Longer axes of swells range from >1000 km to <200 km in length. The larger swells merge with one another and carry smaller elevated areas, called subswells, on their surfaces.

16) Actively subsiding basins within the African continent are simply lower lying areas among swells.

17) The separation of swell crests and therefore the size of intervening basins is varied but there is some concentration, on the continent, at a length scale of about ~700 km.

18) The low-lying surface that formed between ~60 Ma and 30 Ma is widely recognisable as a now elevated surface on the swells of the African continent.

19) Escarpments, including the Great Escarpment, separate this elevated surface from younger erosion surfaces closer to sea level.

20) Erosion of the areas of the continent which began to be elevated at ~30 Ma, coupled with a contemporary low-stand of global sea level, led to development of a great offshore "mid-Oligocene" unconformity around much of the African continent at about 30 Ma.

21) That event was coupled with the development of spectacular and numerous submarine canyons that soon began to be filled with sediment. An episode of salt tectonism began as a result of the canyon cutting.

22) Thick sequences of sediment have accumulated at the base of the continental slope since about 22 Ma. This sediment represents material eroded from the rising swells of the continent, much of which has been carried down the newly cut submarine canyons.

23) The Nile first formed at about 30 Ma in response to erosion from the newly uplifted Ethiopian highlands. The deltas of the Nile, the Niger and the Zambesi have prograded rapidly during the past 30 My as newly elevated areas have supplied much sediment to those rivers.

24) The Orange river has become a less important supplier of sediment to the continental margin since arid conditions set in in southwestern Africa when the Benguela current began to flow at about 34 Ma.

25) The Sahara desert formed at only ~3 Ma and since that time the drainage of the Nile, Niger

and Benue systems has reached the sea only during wetter intervals. During dry times internal drainage systems terminate in the inter-swell basins, especially in those of the Sudd, inland delta and Chad.

26) The East African Rift System sprang to life at about 30 Ma with the eruption of the Afar plume and the simultaneous development of rifts in the Red Sea and Gulf of Suez, the Gulf of Aden, Lokitipi and western Turkana, as well as off the east coast of Mozambique in the offshore Kerimbas and Lacerda Rift Basins.

27) Three plumes may be all that are needed to account for volcanism in different areas of the East African Rift System: These are the Afar plume, the Lokitipi plume and the Samburu plume. All other volcanism within the rift system can be attributed to pressure release volcanism related to rift extension. The Afar plume alone, because of its high ^{3}He signature, is considered a likely candidate to have arisen from deep in the mantle below the 670 km discontinuity.

Volcanic areas close to, but not actually in, the rift system, such as the Miocene carbonatites of Uganda and western Kenya, Quaternary Marsabit, Chiyulu and Mt. Kenya in Kenya, as well as Kilimanjaro and Meru in Tanzania, are part of the 30 Ma and younger African plume-derived volcanic population on a crestal area of the East African swell. They are neither products of plumes directly beneath rifts nor are they products of pressure release volcanism related to rift extension.

28) At ~15 Ma, when the collision of Arabia and Eurasia took place, the style of the East African Rift System changed with the development of the newer harrats of Arabia, the propagation of ocean floor formation into the Gulf of Aden and perhaps also the initiation of the western rift.

29) At ~5 Ma further changes are apparent. These included the start of nearly all of the development of the Dead Sea transform system, the beginnings of ocean floor formation in the Red Sea, a change in the style of rifting in the Afar and the propagation of Turkana rifting southward onto the site overlying the Samburu plume to form the Gregory Rift of Kenya.

What happened to the African Plate ~30 Ma?

Two possible causes, perhaps "triggers" would be a better word, for change in the behavior of the African Plate have been suggested. One developed in this paper, following a suggestion of Bailey (1992; 1993), is that the intensified collision of the African Plate with the Eurasian Plate at 30 Ma sufficed to arrest the motion of the Africa Plate with respect to the underlying mantle circulation. The other, suggested I think for the first time in this paper, is that eruption of the

Afar plume changed the motion of the plate with respect to the underlying circulation.

A possible mechanism (Morgan & Smith, 1992) is that the impingement of the Afar plume on the base of the lithosphere was associated with the entrainment of a large mass of fertile or potential magma source material from a large neighboring volume of the upper mantle. As a result of this entrainment, the viscosity of the asthenosphere beneath an extensive region of the African Plate might have been increased so that movement of the plate with respect to the underlying mantle stopped. Other possibilities abound. For example, more than one plume might have simultaneously reached the bottom of the African lithosphere and produced the kinds of effect here suggested for the Afar plume alone or the two processes of collision with Eurasia and plume impingement could both have happened at the same time and acted together. There is also the possibility that some quite different process arrested the plate motion over the plume population.

Whatever process induced the change, subsequent events are clear. I suggest that many of the phenomena that have characterized Africa over the past 30 My have been induced by the cessation of plate motion, that is a reduction to below a velocity of ~10 mm a year with respect to the mantle circulation. All the changes in the behavior of the African Plate can be explained by the simple idea that, if the plate is not moving over the plume population, then the plumes have time to interact with the overlying plate. If the plate is moving, then interaction is limited because the plate moves away from the site of interaction too fast for the consequences, whether they be thermal, magmatic or topographic, to develop (Figure 50). Erosional and depositional changes on the African Plate over the past 30 My have been dominated by responses to changes in topography induced by the action of underlying plumes.

An explanation for the possible dominance of the Afar plume in changing the behavior of the African Plate lies in the idea that it may be the only one of the 30 Ma and younger plume population of the African Plate which has arisen from great depths in the mantle. This is suggested to me by (1) the huge size of the Afar and the large volume of igneous products erupted from it over the past 30 My and (2) its high ^{3}He signature. I consider its overall geochemistry to be ambiguous as to the depth of origin of the plume.

If the impingement of the Afar plume arising from great depth arrested the motion of the African Plate, then all of the other plumes and swells that have developed on and under the African Plate during the past 30 My can be interpreted to be products of convection confined to the shallow mantle above the 670 km discontinuity. This seems to me compatible with (1) the wavelength of the swells which clusters around ~700 km, (2) the importance of basement elevation and the small volume of basalt at the hot spots, (3) the dominance of HIMU geochemistry and (4) the general low ^{3}He signature.

The kind of upper mantle convective circulation which can be discerned under the African Plate, on the evidence of the variety of phenomena that I have described, is not perceptible under the other plates of the earth because the motion of those plates with respect to comparable convective circulation smears out the pattern and makes it impossible to discern. The "Richter rolls," which have been suggested to underlie the Pacific Plate, perhaps form part of a similar pattern under a moving plate.

African hot spots that do not fit the shallow mantle derivation pattern suggested for most of the population are few. They include: (1) Tristan, which is recognised to be part of another population because its track goes back to a site of impingement at 130 Ma; (2) the Afar; (3) possibly the Cape Verde islands, which Gerlach *et al.* (1988) suggested from the geochemistry of some of the northern islands were not erupted from the HIMU source. Courtney & White (1986) cited unpublished work by Knill indicating that the thermal effects of the Cape Verde plume might have been first felt at as much as 50 Ma; and (4) Réunion which, although only 2 million years old, has produced an unusually large volume of basalt by African Plate standards and perhaps has no associated basement swell. Réunion also has a high ^{3}He signature and distinctive isotopic composition.

Perhaps the kind of pattern of the circulation in the mantle shallower than 670 km that is suggested to me by the topography of the African Plate is one that obtains over much of the earth, but is obscured everywhere else by the motion of the overlying plates with respect to the underlying plumes (Figure 24).

On the African continent, the difference between swells with volcanic hot spots on or close to their crests and those without such volcanism corresponds rather well with the distribution of continental crust reactivated during Pan-African times. One unanswered question is whether the occurrence of a huge volume of seismically slow, and therefore perhaps hot and low-density, material in the lower mantle below southernmost Africa is playing a role in the elevation of southern Africa. This is one, but perhaps the most striking aspect, of the whole question of how much of the newly developed surface structure of the African Plate relates to the upper, and how much to the lower, mantle

The age of the first volcanic rocks erupted at individual hot spots of the 30 Ma and younger population on the African Plate ranges from as old as 31 Ma to as young as 2 Ma or less at, for example, Réunion or Marsabit. It seems possible that this range of ages might indicate changes in the time dependent upward flux of convecting material at different sites within the past 30 My. Perhaps some of the shallow source plumes have impinged on the base of the lithosphere only recently. Réunion is a strong candidate for recent impingement but, with its relatively high ^{3}He signal, it could have come from deeper than the rest of the population. Areas on the African Plate, with a large amount of volcanism early in the past 30 My and recent inactivity, appear to be few but this could represent a lack of recognition consequent upon erosion of the older material from areas which have not remained high through the entire period.

Elevation of parts of the swells seems to have happened at different times as has been inferred from evidence of Quaternary river capture. This is perhaps also a reflection of time dependent behavior in a rising plume. The extremely widespread distribution of the offshore ~30 Ma unconformity is a strong indication that substantial uplift began in many regions at the same time. One of the few regions in which evidence of deposition of different amounts of since 30 Ma has been analysed is at the Orange river mouth where the change has been attributed to climatic change rather than to tectonic variation.

The East African Rift System is clearly part of the story of what has happened since the African Plate came to rest over the plume population because the rift system, in its present incarnation, began to form at 30 Ma. In the Red Sea, Gulf of Aden and Afar, the rift system was dominated by the eruption of the Afar plume and the simultaneous formation of the initial rifts, perhaps in response to the modification of within-plate stresses (Figure 33). Farther south the development of the eastern rift has been much more complex and I would relate it to the influence of discrete hot spots at Lokitipi and Samburu, as well as to the influence of the structure of the underlying Anza Rift. The formation of the western rift is a much younger event, perhaps related, as are the new harrats of Arabia, to the 15 Ma collision of Africa with Eurasia in the Zagros.

Although the gross structure of the African Plate was established very rapidly soon after the plate appears to have come to rest over the underlying plume population ~30 Ma, there has been continuing evolution, especially in the rift system and in such processes as erosion and deposition. The great prograding deltas provide evidence of this kind of continuing change.

On the global environmental scale, the development of the East Antarctic ice sheet, with accompanying sea level fall worldwide at ~34 Ma and the near contemporary initiation of the

Benguela current, have had a strong influence on developments in the area occupied by the African Plate. The much later initiation of northern hemisphere glaciation and of the Sahara desert at about 3 million years ago has also played a role. These later phenomena appear to have been important influences on human evolution. Whatever the process of hominid evolution might have been, the record would not be so good were it not for the distinctive tectonic environment of the East African Rift System with repeated faulting and frequent dateable volcanic eruptions.

In summary, a history of how the African Plate came to rest over the mantle plume population at ~30 Ma, and about what has happened as a consequence, can be constructed. A good question now might be: How can this idea be tested?

Testing the Africa-at-rest hypothesis

Africa is a vast continent and the African Plate is even larger. Much of the information that I have tried to use in this text is known with very poor resolution. The biggest single need is not just to determine what has happened and is happening on the African Plate, but also to determine what happened and when it happened. Two kinds of critical information can probably be greatly refined. Many ages determined on volcanic rocks in early work before the availability of $^{40}Ar/^{39}Ar$ ages may be misleading. A program of reassessment and new age determination, especially for the older volcanic rocks with ages around 30 Ma, is clearly needed. It would be exciting, for example, to know if indeed the Afar plume was the first of the post 30 Ma population to erupt.

A second kind of information is locked up in offshore regional seismic lines, especially where local sequence stratigraphy is controlled by well data. It is surely possible to determine not just when the "mid-Oligocene" unconformity was cut, but also how soon deposition began above that unconformity and what volumes of sediments were deposited above the unconformity in what areas over what intervals of time.

This kind of study will show how much sediment was deposited offshore, in what areas and at what times. River systems on shore must have supplied much of that sediment and on-shore river, escarpment and erosion surface studies can be integrated with information from offshore. Rather complete models of erosional and basin systems will be constructable. Information about timing from onshore volcanic rock ages may help even further.

Geophysics, including mantle tomography, satellite altimetry, seismic, gravity and magnetic studies, will all help to clarify the structure of the African Plate and of the mantle beneath it.

In the deep water far offshore, integrated models will need to embody satellite altimetry which can yield free air gravity maps, magnetic anomaly data, bathymetry and sediment thickness maps. A beginning in this kind of study has been made by Nyblade & Robinson (1994) but the opportunities are great for integration also of volcanic rock ages, Ocean Drilling Project holes and magnetic anomaly data into analyses. A special environment is represented by ocean floor that has formed within the past 30 My. Here the opportunities for characterizing a distinctive environment are unique. One special place is the Central Indian Ridge where the new satellite altimetric data will allow a complete reassessment of the time of reconstruction. The present favored time is 38 Ma (Patriat & Segoufin, 1988) but I would favor an age five or more million years younger. The geochemistry of 30 Ma and younger volcanic rocks on the continent and on the ocean floor will reveal much about their underlying mantle sources.

I do not wish to pursue further the topic of how best to test my hypothesis. I hope that many people are interested in proving me wrong and that they have thought of good ways of doing so. I would end by strongly emphasising, from the example of Alex L. du Toit, that an integrated approach to problem solving for the the understanding of the past 30 Ma history of the African Plate is essential. Clearly the scientists and engineers who deploy broad-band seismic instruments to address the deep structure beneath the continent have special skills very different from those of the scientists who interpret the microfossil records of Ocean Drilling Project cores. Understanding the evolution of the African Plate will call for integrating the results of studies as diverse as these with many others equally diverse.

Du Toit (1937) did just that kind of integration in the research which he embodied into "Our Wandering Continents." We owe it to his memory not to attempt less in an integrated study of the past 30 million year history of the African Plate.

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FIGURES

Figure 1 Basins and swells of Africa. Modified after Holmes (1944; 1965, his figure 763). Dashed lines = the extent of Lake Megachad. French and German language writers had emphasised the location of the East African Rift on the crest of the East African swell and Krenkel in particular (1922; 1957) had emphasised the Basin and Swell structure of the continent.

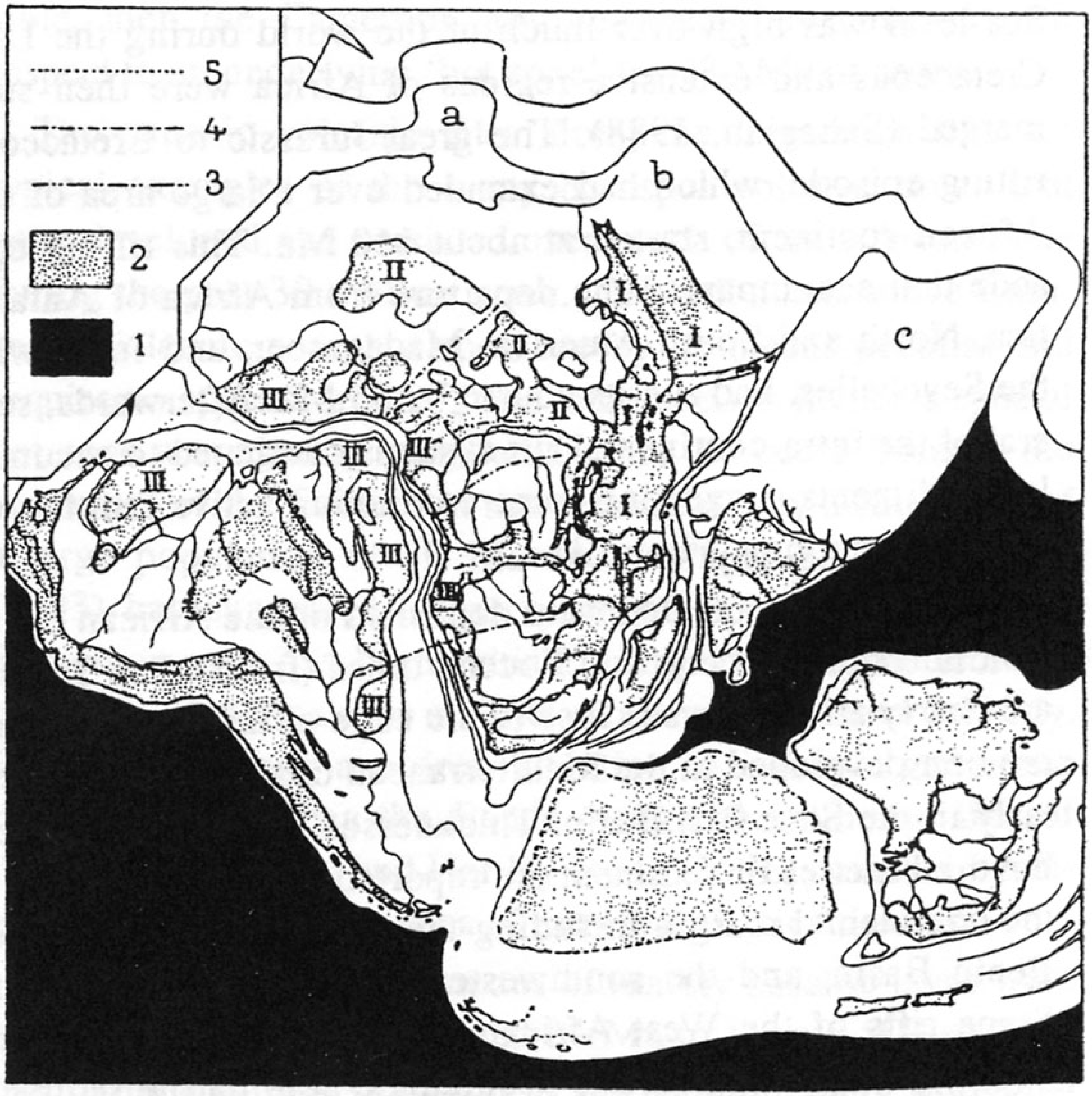


Figure 2 Argand (1924, his figure 6) had also recognised the basin and swell structure of Africa which he attributed to compressional folding. He applied the term *plis de fond* to these structures.

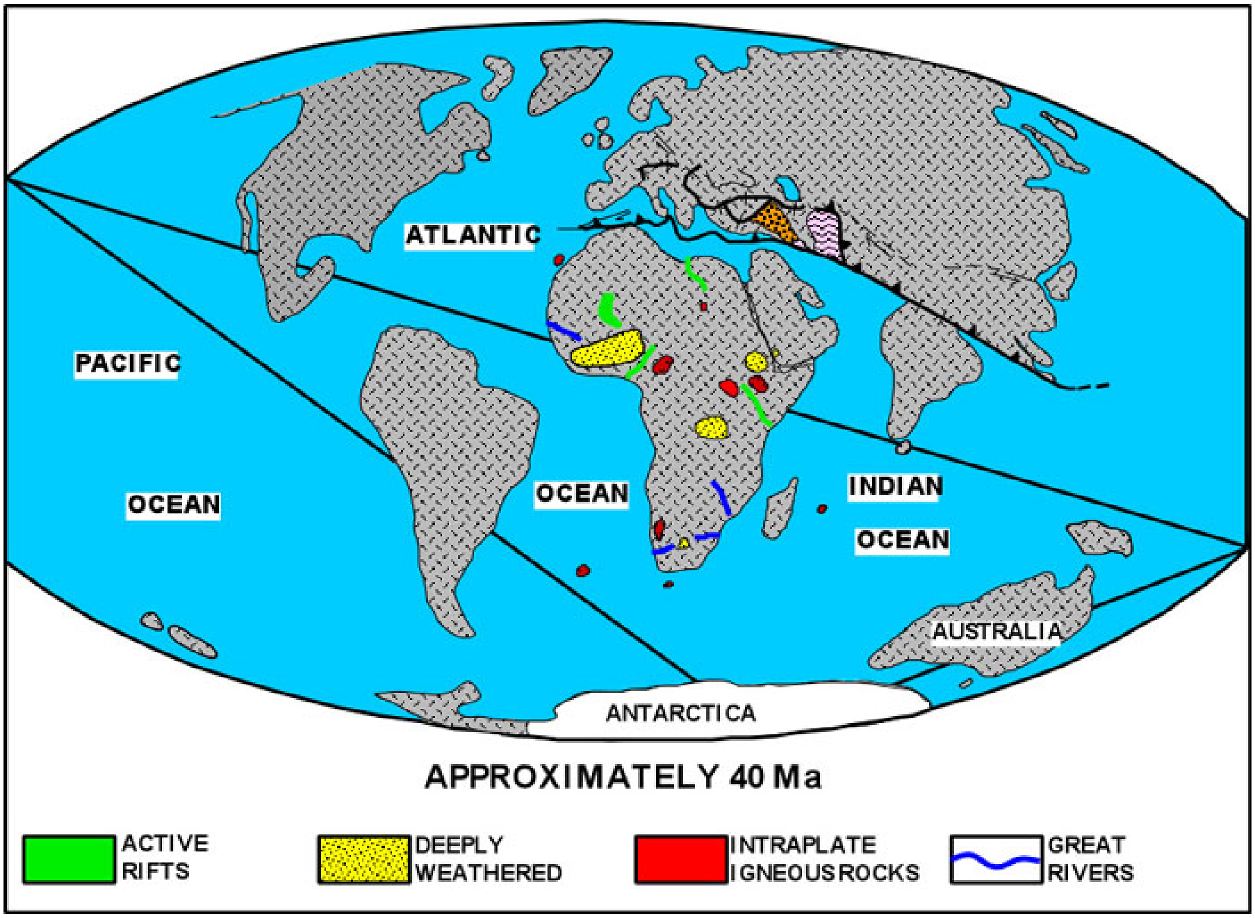


Figure 3 Africa about 40 Ma, before it collided with Eurasia, was a low-lying continent with widespread deep weathering. The Benue, Sirte and Anza Rifts which had formed near the beginning of the Cretaceous (~140 Ma) were continuing to receive sediment. Igneous activity on the African Plate was restricted to the Tristan and Deccan plume hot spots and to sporadic localities on the continent. Continental igneous activity between 65 and 45 Ma was short-lived everywhere except in northern Cameroon where there was continuous granitic activity. Figure modified from Sengör (1989).

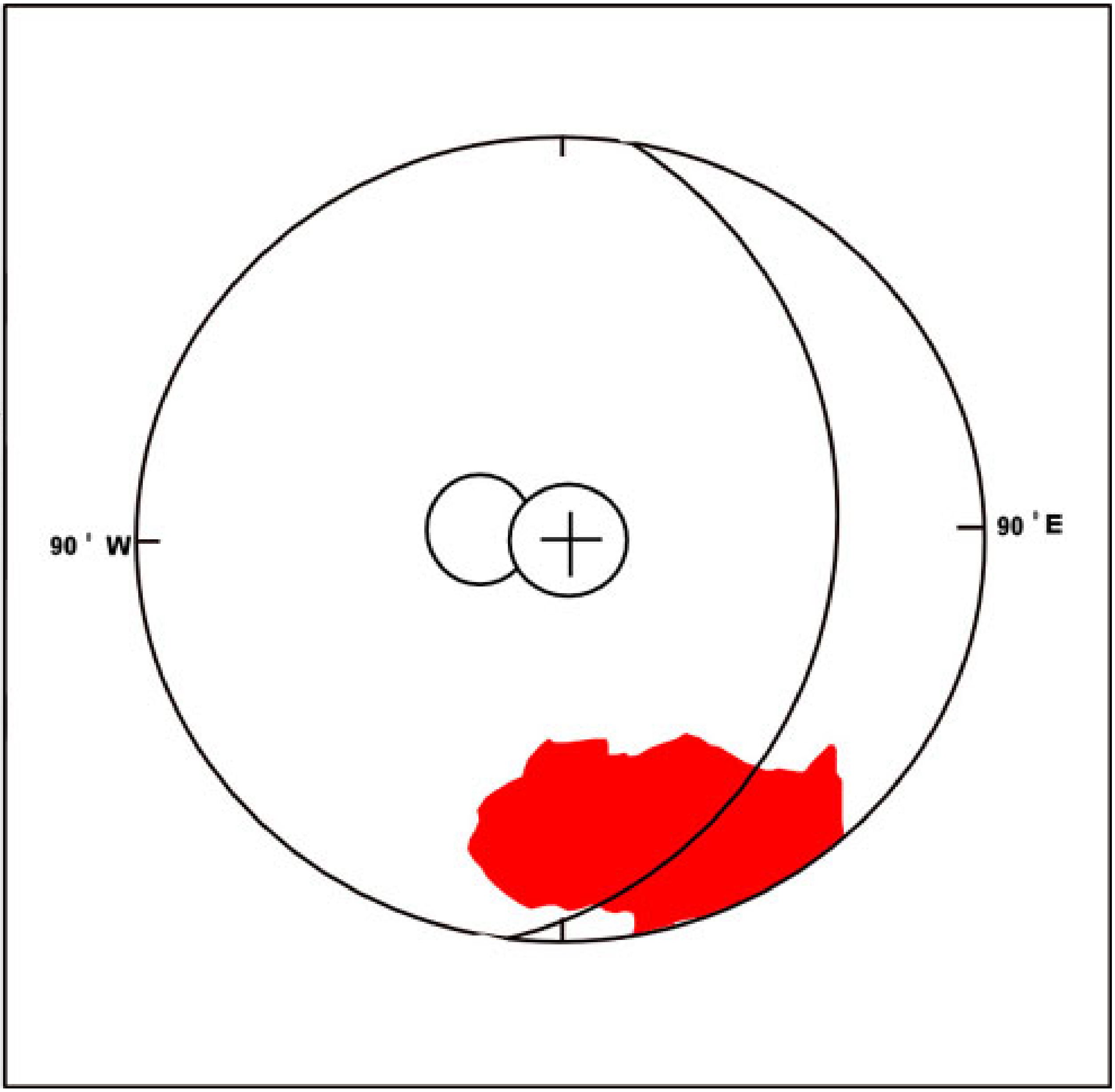


Figure 4 Lower hemisphere stereo projection showing small circles marking envelopes to the locations of paleomagnetic poles for volcanic rocks of Africa. Poles for rocks with ages between 200 Ma and 100 Ma lie within the open circle and poles for volcanic rocks with ages of < 30 Ma lie clustered about the present pole which is marked with crosslines. Between about 100 Ma and 30 Ma, the African Plate rotated a few degrees clockwise about a pole close to zero degrees of both latitude and longitude. Few igneous rocks were erupted on the plate during the 100 Ma to 30 Ma interval. Figure modified from Burke & Dewey (1974).

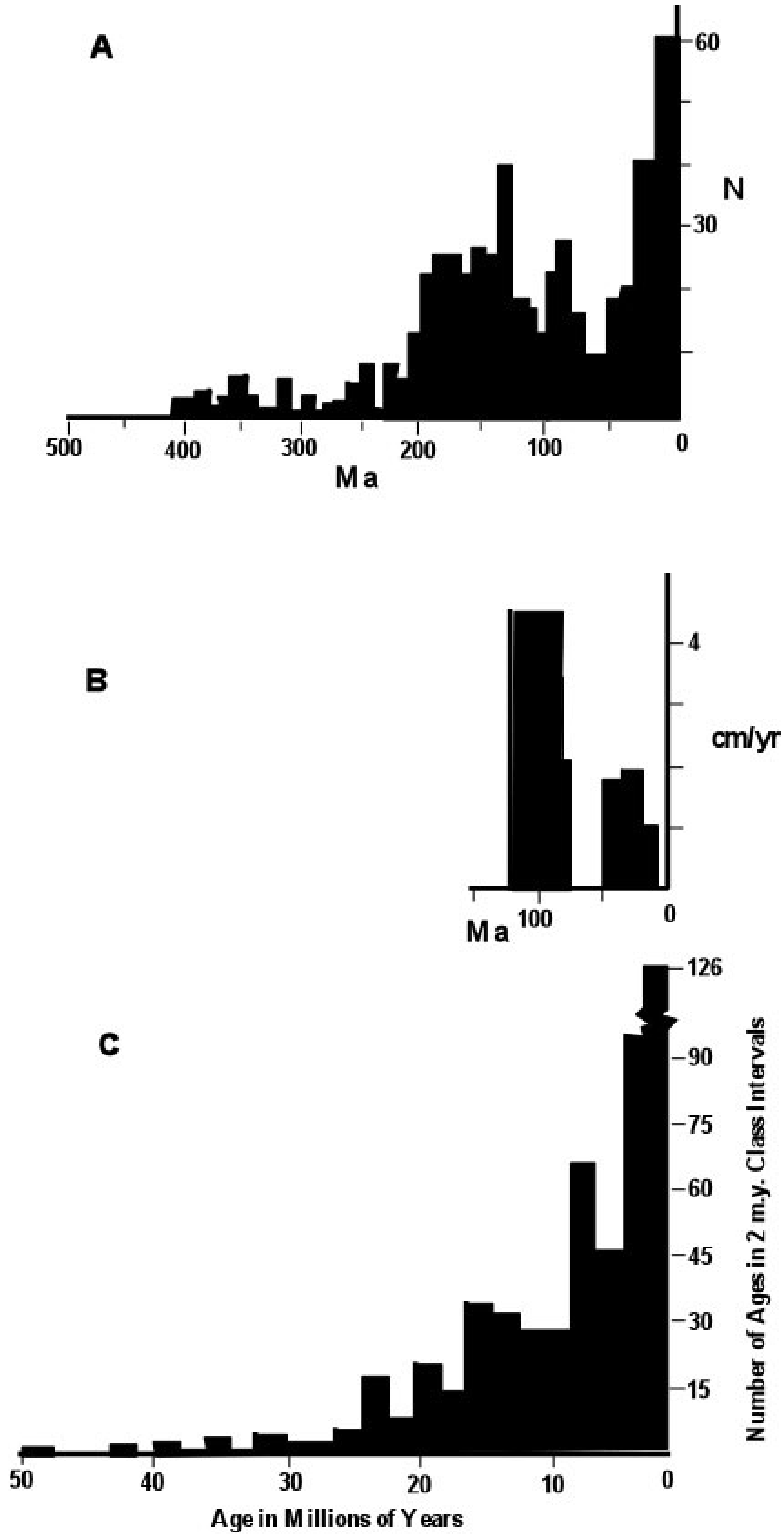


Figure 5 Histograms showing: (A) compilation of isotopic ages for African igneous rocks from Bailey (1993); (B) compilation, also from Bailey (1993), indicating how the velocity of Africa with respect to Eurasia fell during the 80 Ma to 50 Ma interval when few igneous rocks were erupted in Africa; (C) compilation illustrating how isotopic ages for igneous rocks of the African continent have become increasingly abundant over the past 30 Ma. From Burke (1976b).

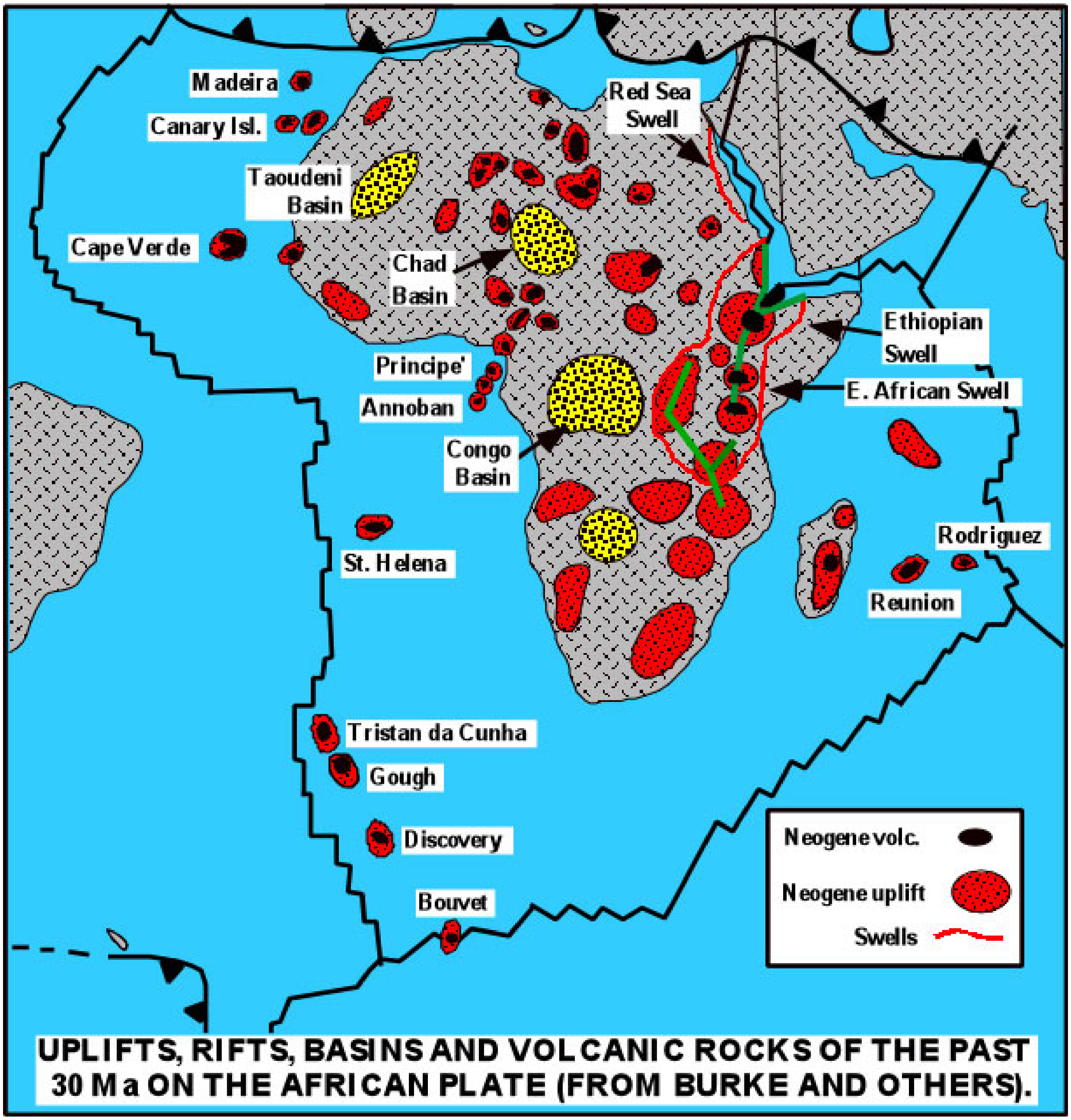


Figure 6 Map illustrating the distribution of volcanic rocks erupted over the past 30 Ma on the African Plate. Swells are enclosed by dashed lines. Note that much, but by no means all, of the volcanic activity occupies the crests of topographic or basement swells. The locations of four interior basins, bounded by swells, are indicated. From a figure drawn by Bill Kidd (figure 6.2.19) in *Basaltic Volcanism on the Terrestrial Planets* (1981).

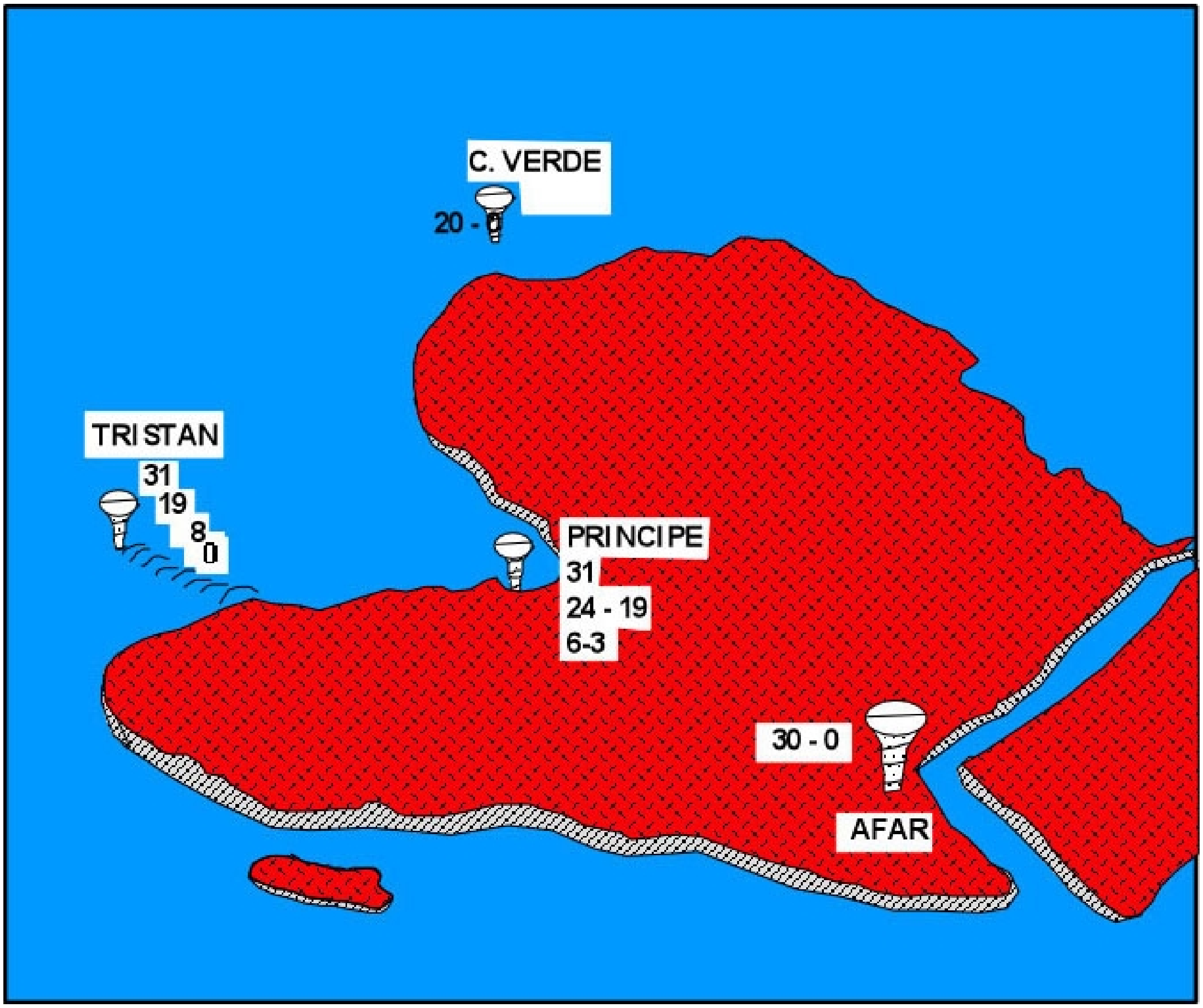


Figure 7 Oblique view of Africa showing it as screwed down. Episodic volcanic activity at the same place has been recorded for several areas on the African Plate through the past 20 to 30 My. This I interpret as showing: (1) that the African Plate has not moved with respect to underlying plumes which are the source of that volcanism, and (2) that those plumes have not moved with respect to each other over that interval.

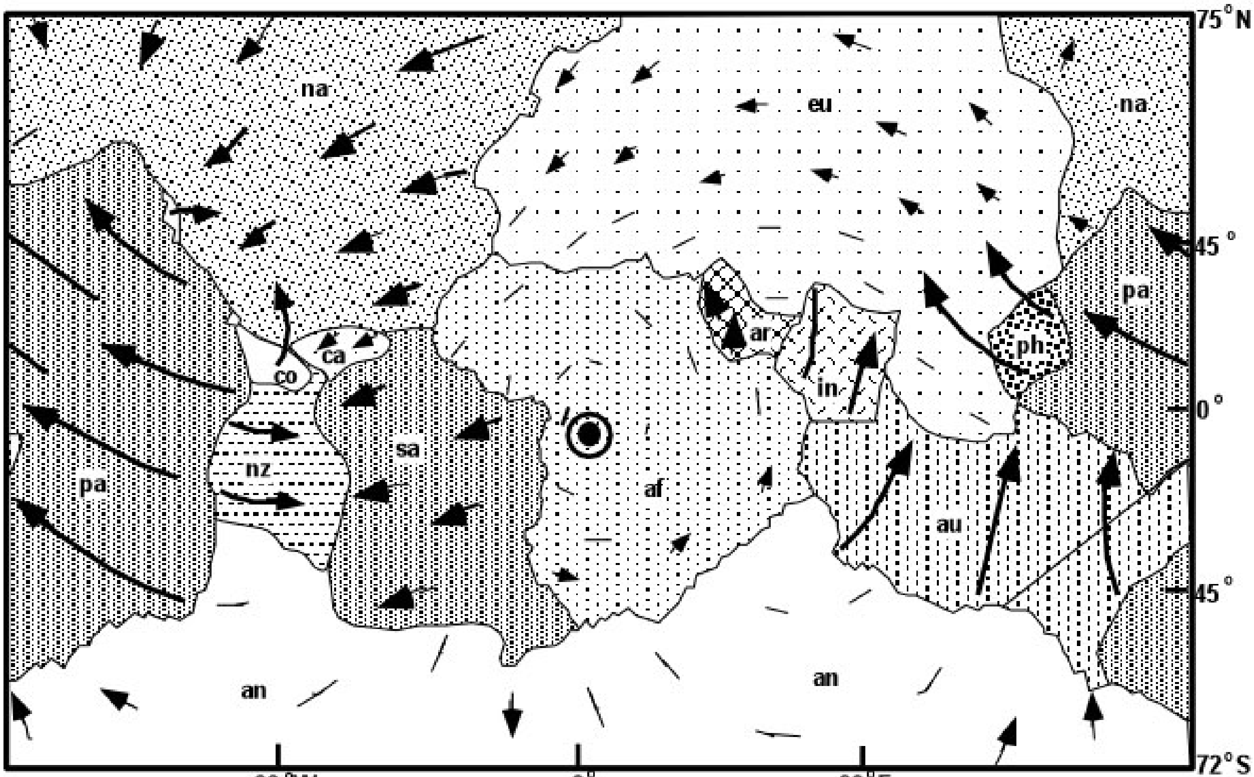


Figure 8 Sketch redrawn from Gripp & Gordon (1990) showing how far the individual plates that make up the earth's lithosphere would move with respect to a fixed hot-spot population over the next 45 My. It is assumed that the present motion of the plates derived from the Nuvel-1 study (DeMets *et al.*, 1990) and representing perhaps the past 3.7 My will persist. Note that the African Plate appears to be rotating very slowly about an internal pole at zero degrees latitude and longitude. It is important to emphasise that the directions of movement and the very slow velocities depicted by the short arrows shown on the African Plate are very poorly determined. Perhaps all that can be said is that the results of this study are not incompatible with those reported here which show that Africa has been at rest with respect to the underlying plumes for the past 30 My.

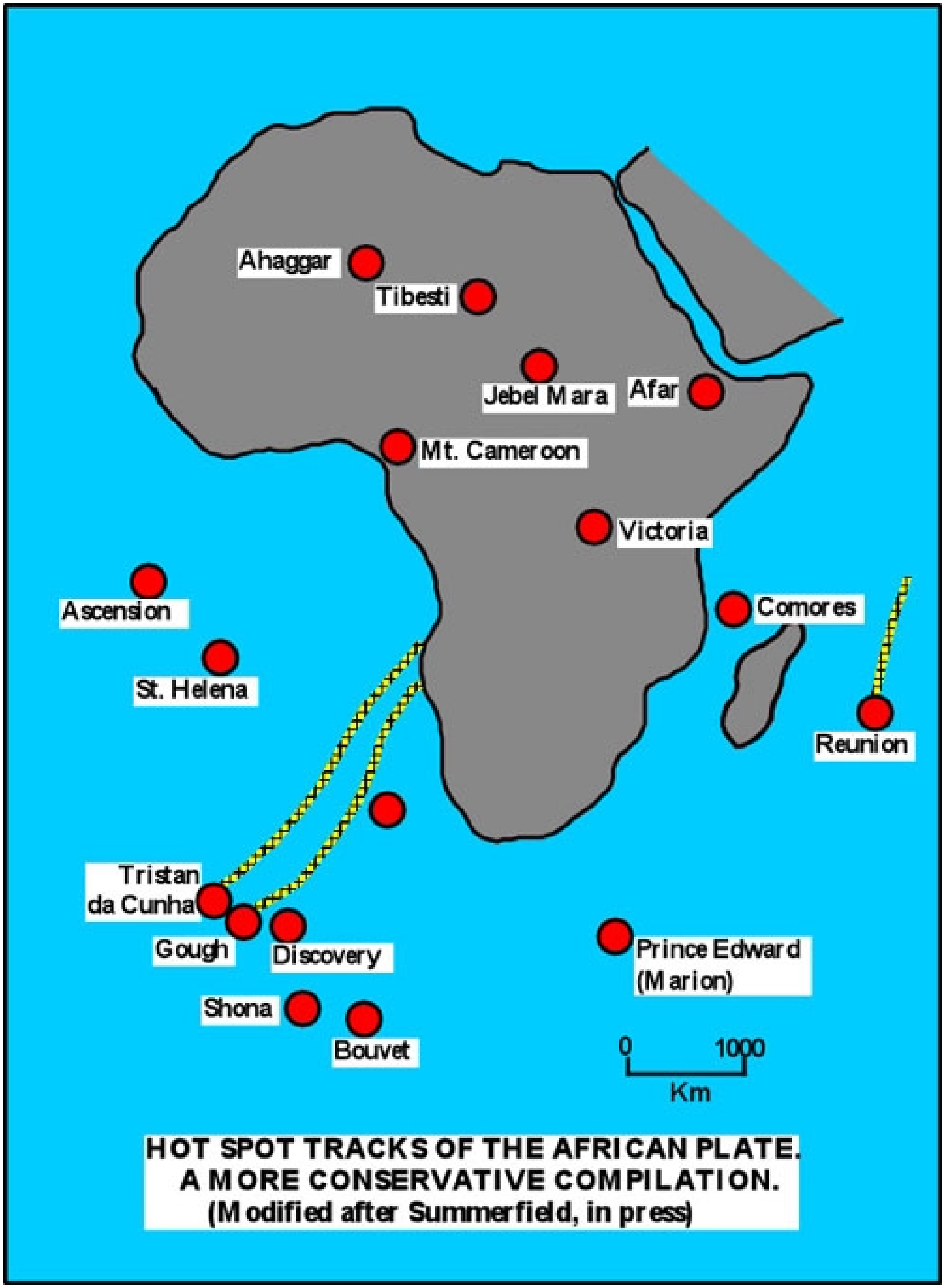


Figure 9 A compilation of hot-spot tracks for the African Plate based on the ideas (1) that all the volcanic areas represented by small open circles are generated by underlying plumes which have existed for more than 150 My, and (2) that the plumes have remained fixed with respect to one another throughout that 150 My. Based on a map in Summerfield (1996).

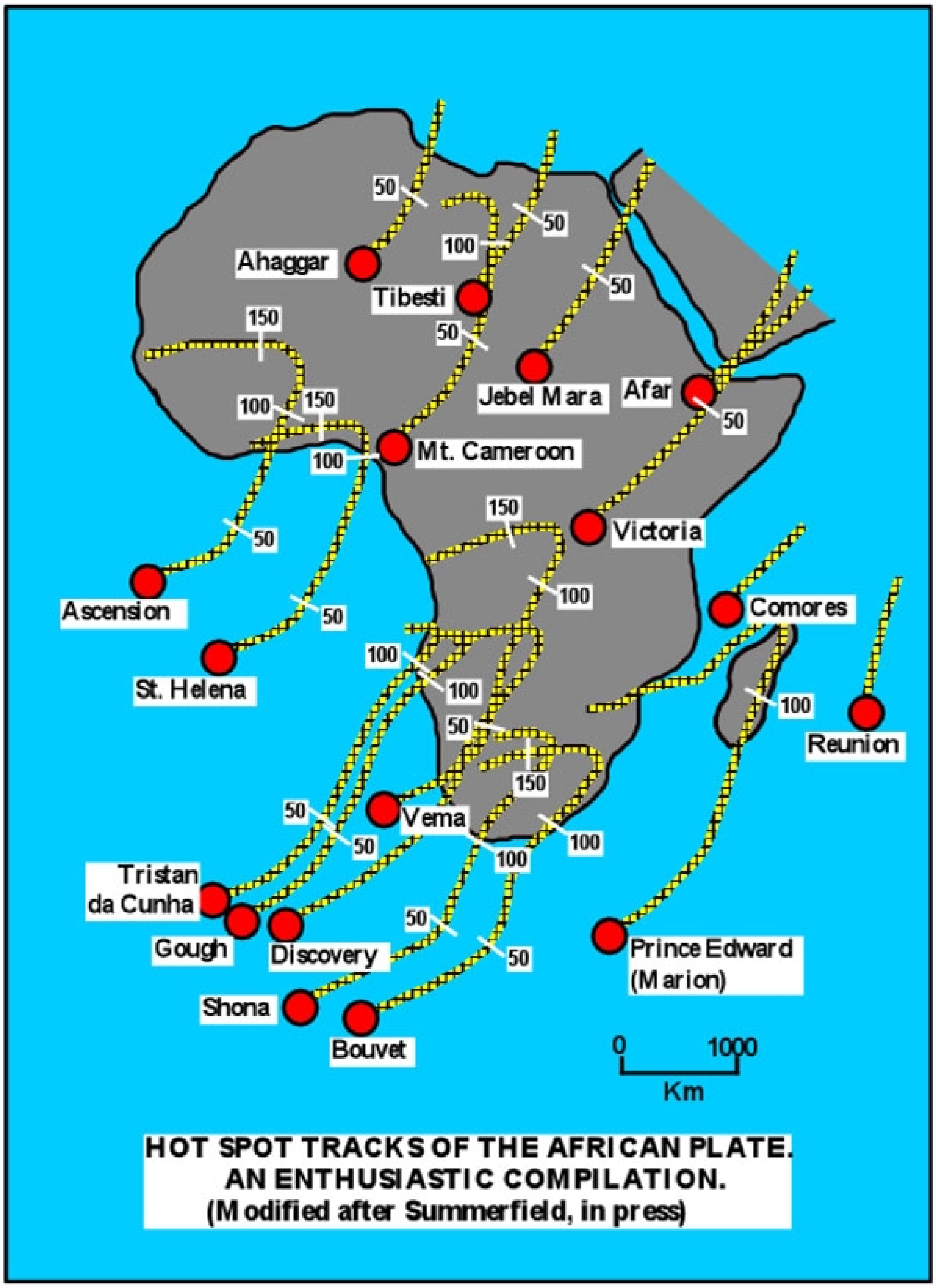


Figure 10 A compilation of hot-spot tracks for the African Plate based on the results of the present study. Only Tristan and Gough, shown here as two separate tracks from a source at the Etendeka basalt outcrop, meet the criteria used here. It is possible that there is a track between southern Ethiopia and Lokitipi representing the interval ~40 to ~30 Ma and that other tracks on the oceanic part of the plate such as that of the Sierra Leone rise (Schilling *et al.*, 1994) may become well enough established to meet the criteria used here at some time in the future.

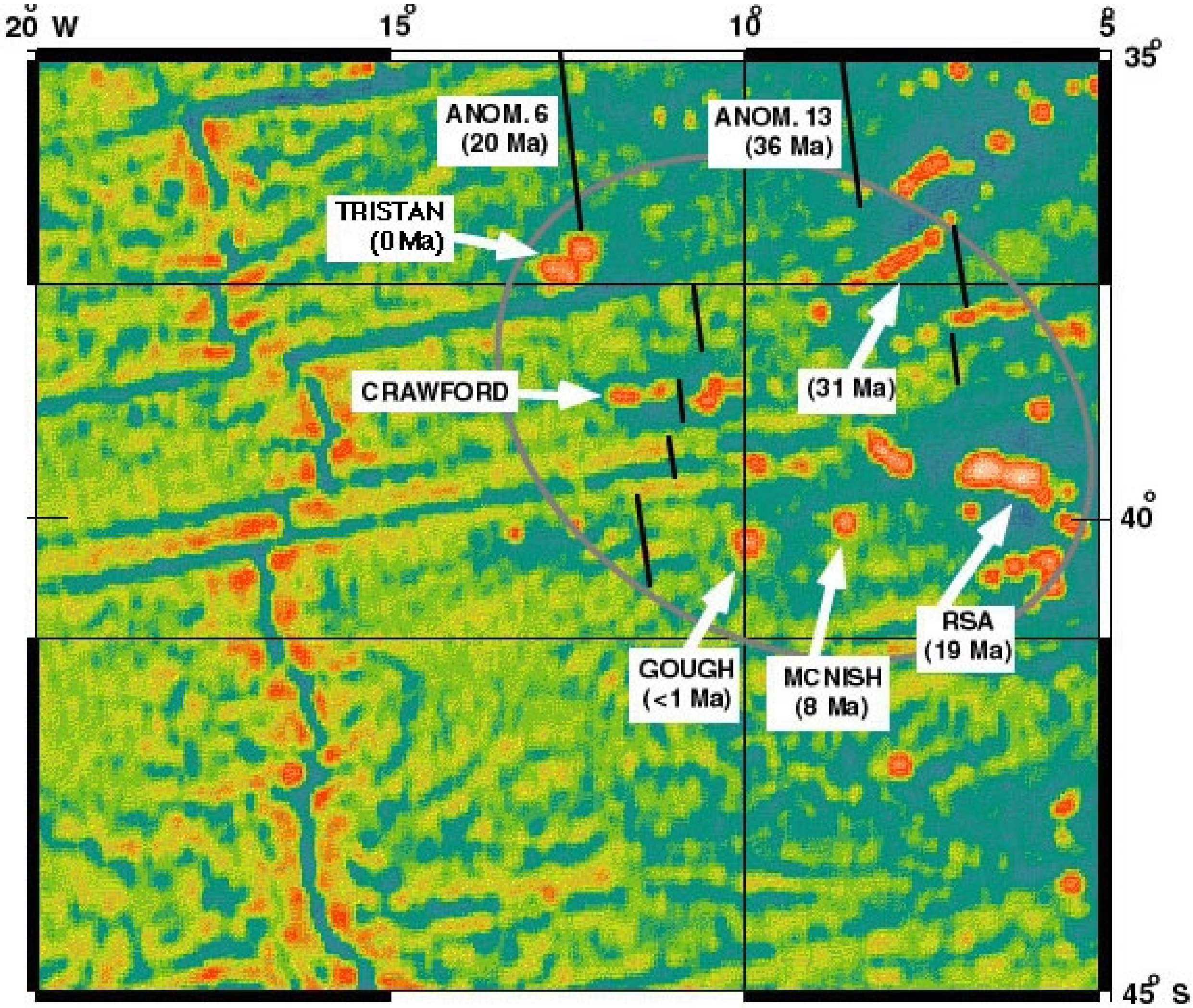


Figure 11 The South Atlantic spreading center to the west, and the present site of hot-spot volcanism located over the Tristan and Gough or Walvis Ridge mantle plume. The young volcanic islands of Tristan and Gough, as well as three dated seamounts all less than 31 Ma, lie within an elliptical area which straddles ocean floor generated at 30 Ma (O'Conner & le Roex, 1992). The underlying plume is interpreted as having remained fixed with respect to the African Plate for the past ~30 My, while the South Atlantic spreading center has migrated westward for a distance of about 650 km. Map is based on satellite altimetry from Sandwell *et al.* (1994).

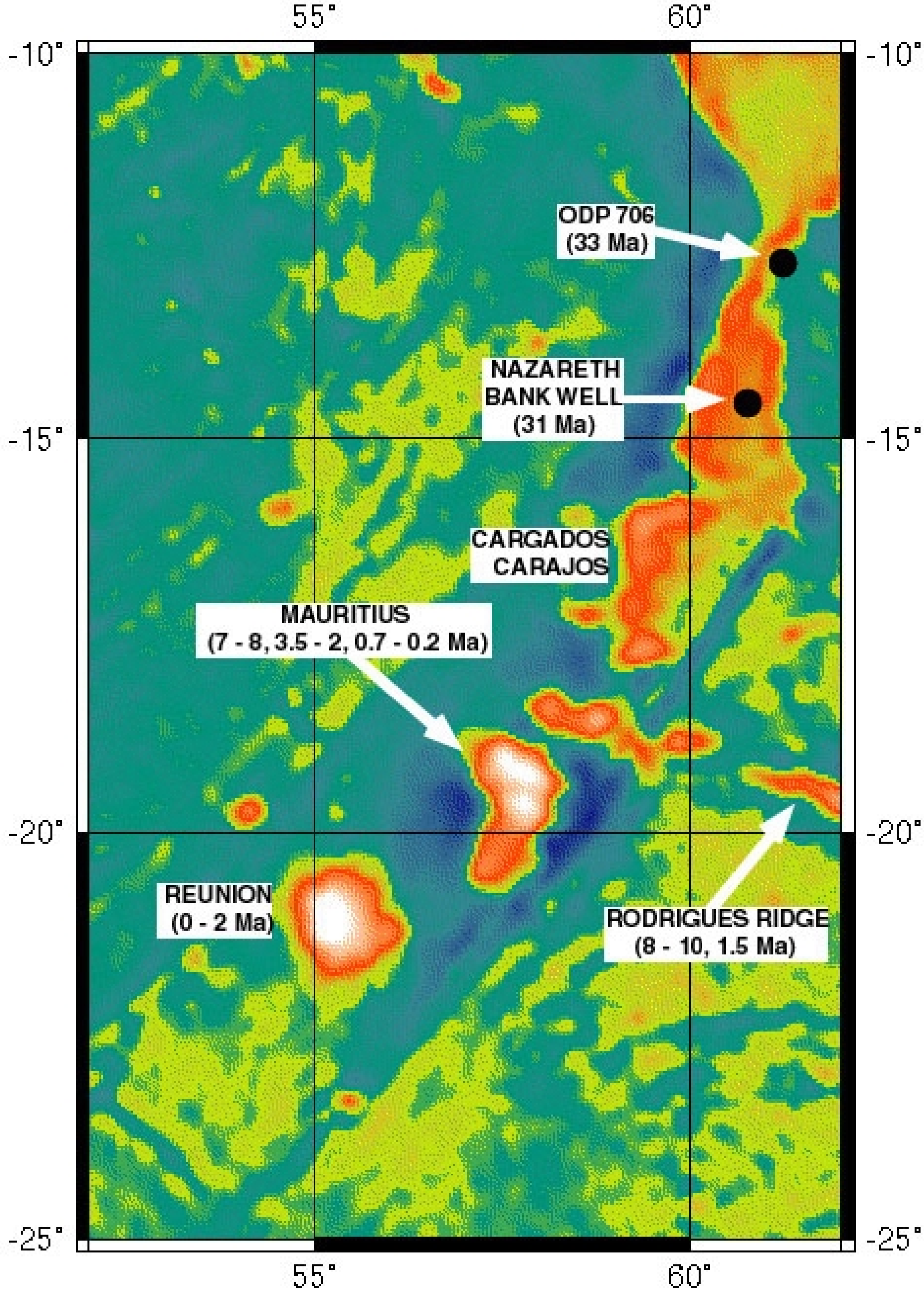


Figure 12 Map based on satellite altimetry showing the positions of the islands of Réunion and Mauritius as well as the Rodrigues Ridge and the carbonate covered Nazareth bank. The bank to the north of Nazareth bank is the Saya de Malha bank and that to the south is the Cargados Carajos bank. Réunion is a very young island yielding isotopic ages no older than 2 My. Only if the plume source beneath Réunion has moved independently of the rest of the sub-African plume population for a distance of more than 700 km between 31 Ma and 2 Ma can Réunion be a product of the plume whose products were encountered at the bottom of ODP hole 706 and the Nazareth bank well. Because I feel uncomfortable about one plume out of a population of about forty careening off independently on its own, I prefer the hypothesis that the plume that made the basaltic substrata of the Saya de Malha, Nazareth and Cargados Carajos banks, which was the Deccan trap source plume, died at about 30 Ma when the African Plate came to rest. Mauritius, Rodrigues Ridge, Rodrigues island (which is just off the map to the east) and Réunion are four separate members of the youthful population of African hot spots each of which is underlain by its own discrete young plume.

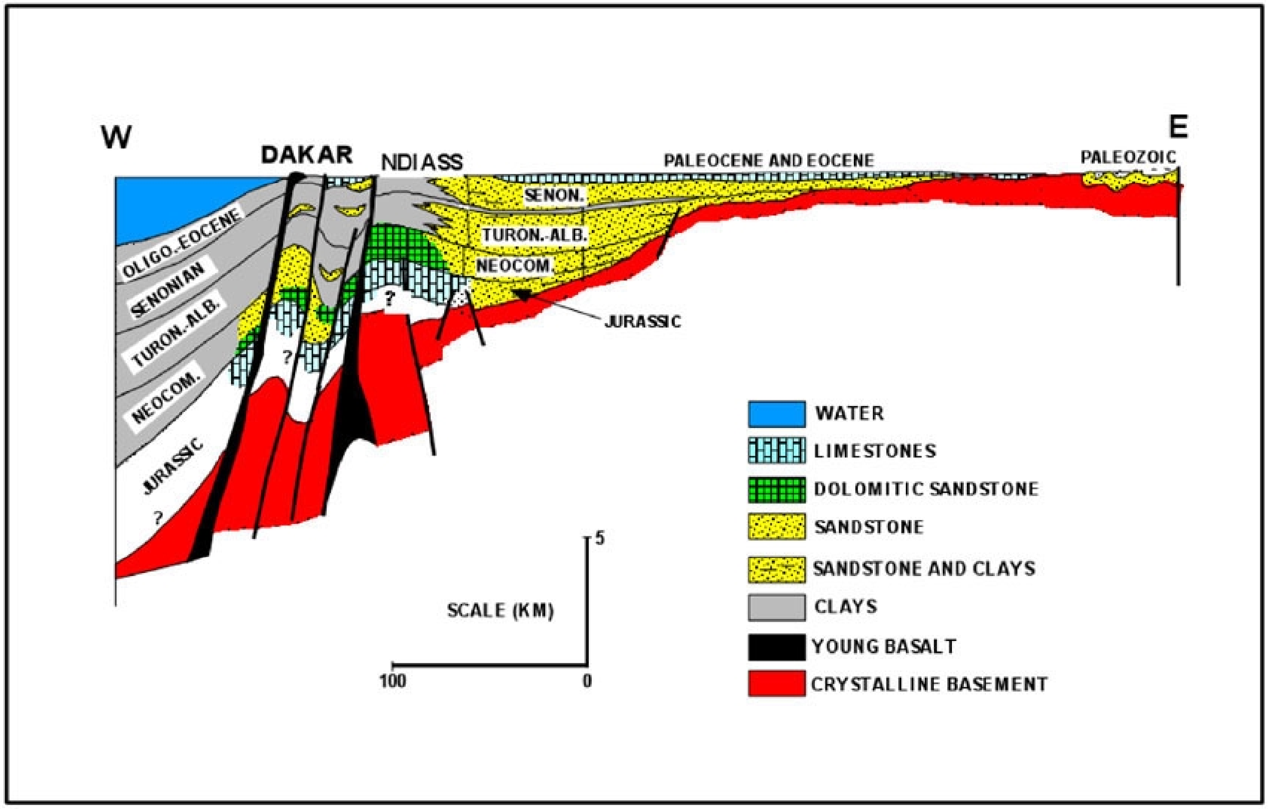
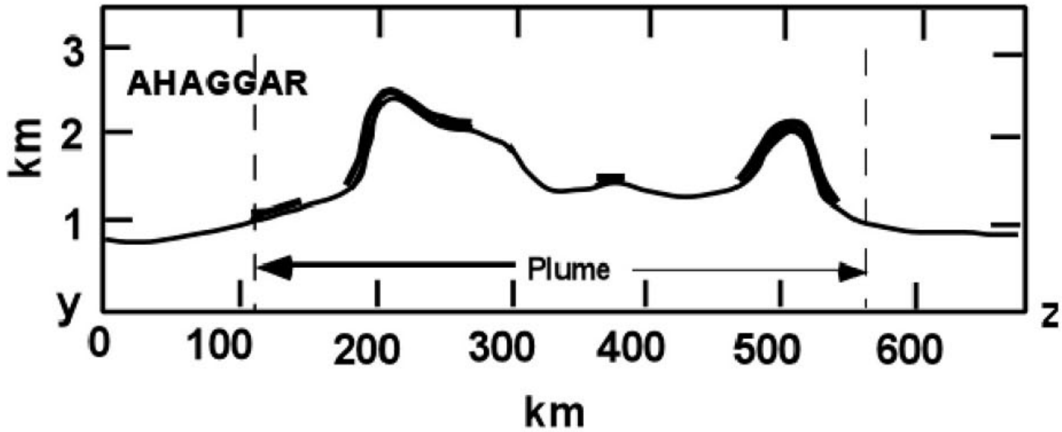


Figure 13 Cross-section through the Dakar hot spot at the westernmost point on the African continent showing that although the two volcanoes of Dakar (known as the *Mammelles*) have little topographic expression, they are associated with one or more kilometres of basement uplift. Redrawn from Spengler *et al.* (1966).



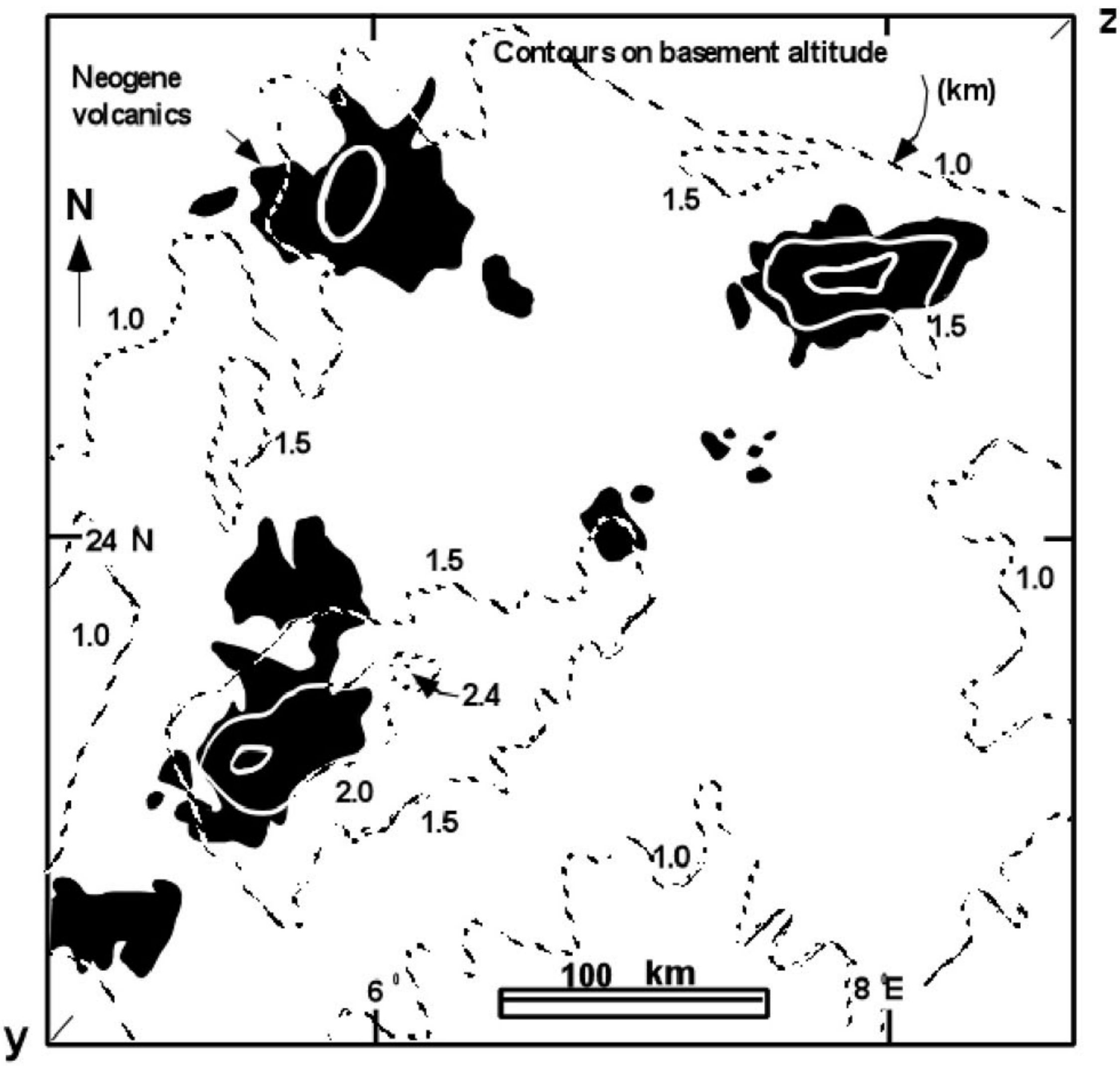


Figure 14 Map and cross-section through the Ahaggar massif in the central Sahara to show hot spot volcanism associated with an elliptical swell of about 300 km by 500 km. The three largest volcanic areas within the Ahaggar occupy separate smaller subswells roughly 100 km in longer axis. The cross-section indicates that the area may be underlain by a large plume with three smaller structures within it. Note that the hot spot volcanism consists of relatively thin outpourings of basalt on the top of the basement uplift. This is the rule among both oceanic and continental hot spots on the African Plate. The figure is redrawn from one in *Basaltic Volcanism on the Terrestrial Planets* (1981), which is itself based on one by Black & Girod (1970).

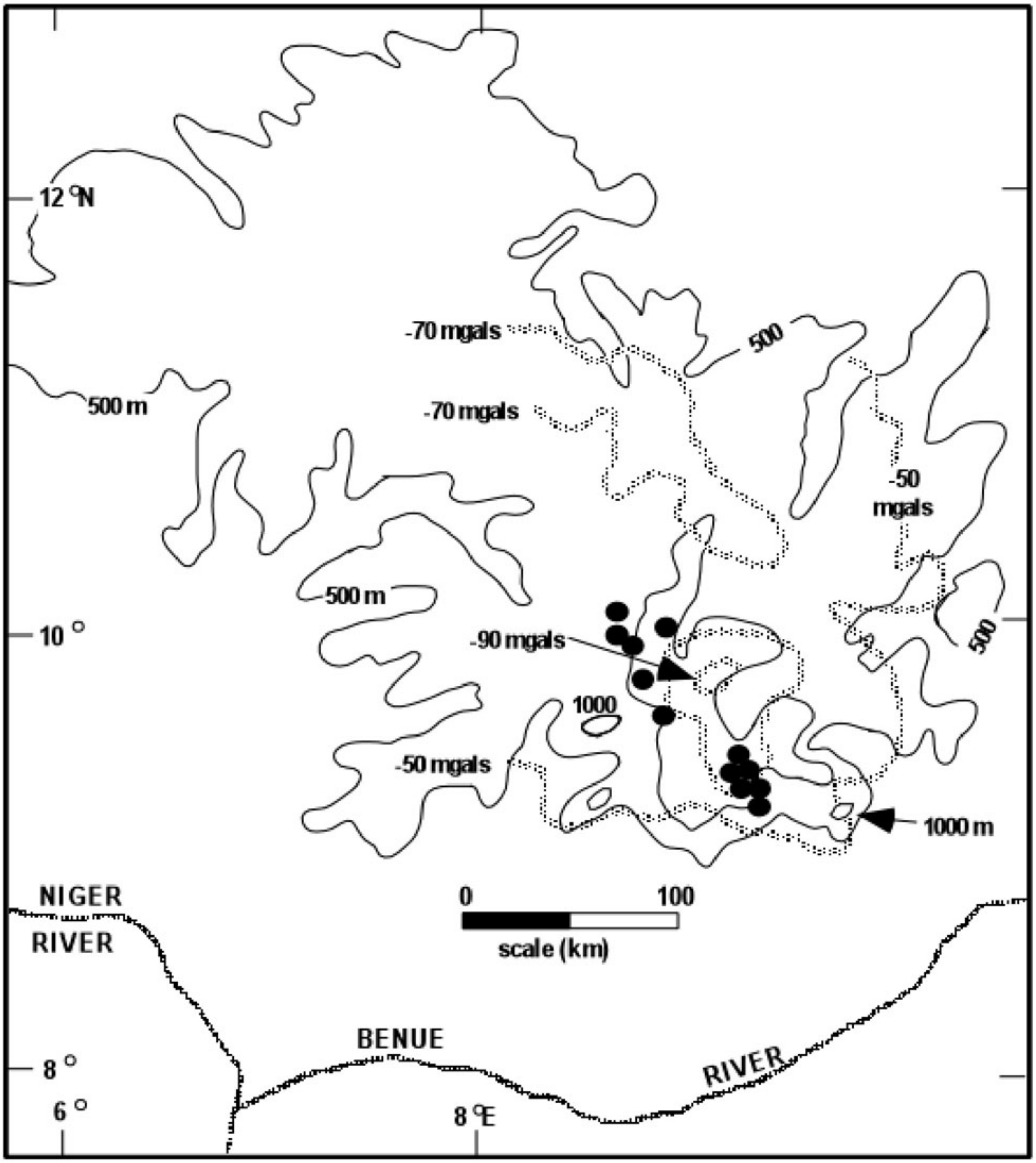


Figure 15 The Jos Plateau in northern Nigeria is one of the smaller swells of the African continent. The plateau consists of an elliptical area ~400 km by 200 km. To indicate elevation, 1000 m and 500 m topographic contours have been drawn. Round dots show the sites of a line of twelve Quaternary cinder cones which parallels both the crest of topographic swell and the trend of the negative Bouguer anomaly (Ajakaiye & Burke, 1973).

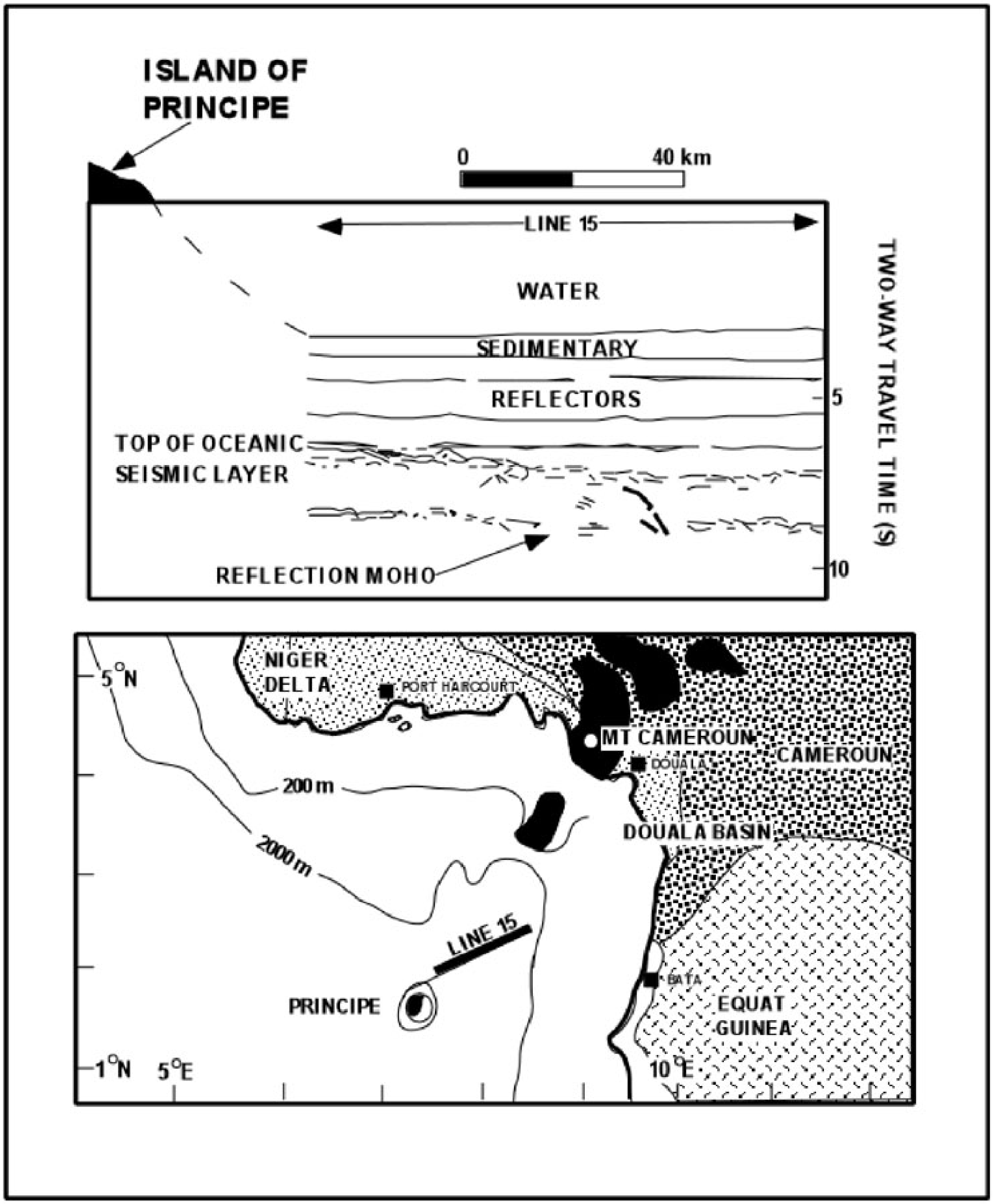


Figure 16 A A deep penetration seismic reflection profile northeast of the island of Principe in the Cameroon volcanic line shows that both the top of the oceanic seismic layer and the reflection Moho rise toward the island indicating basement uplift beneath one of Africa's oceanic hot spots. Mt. Cameroon at the coastline is similarly underlain by a basement uplift. Figure redrawn from Rosendahl *et al.* (1991).

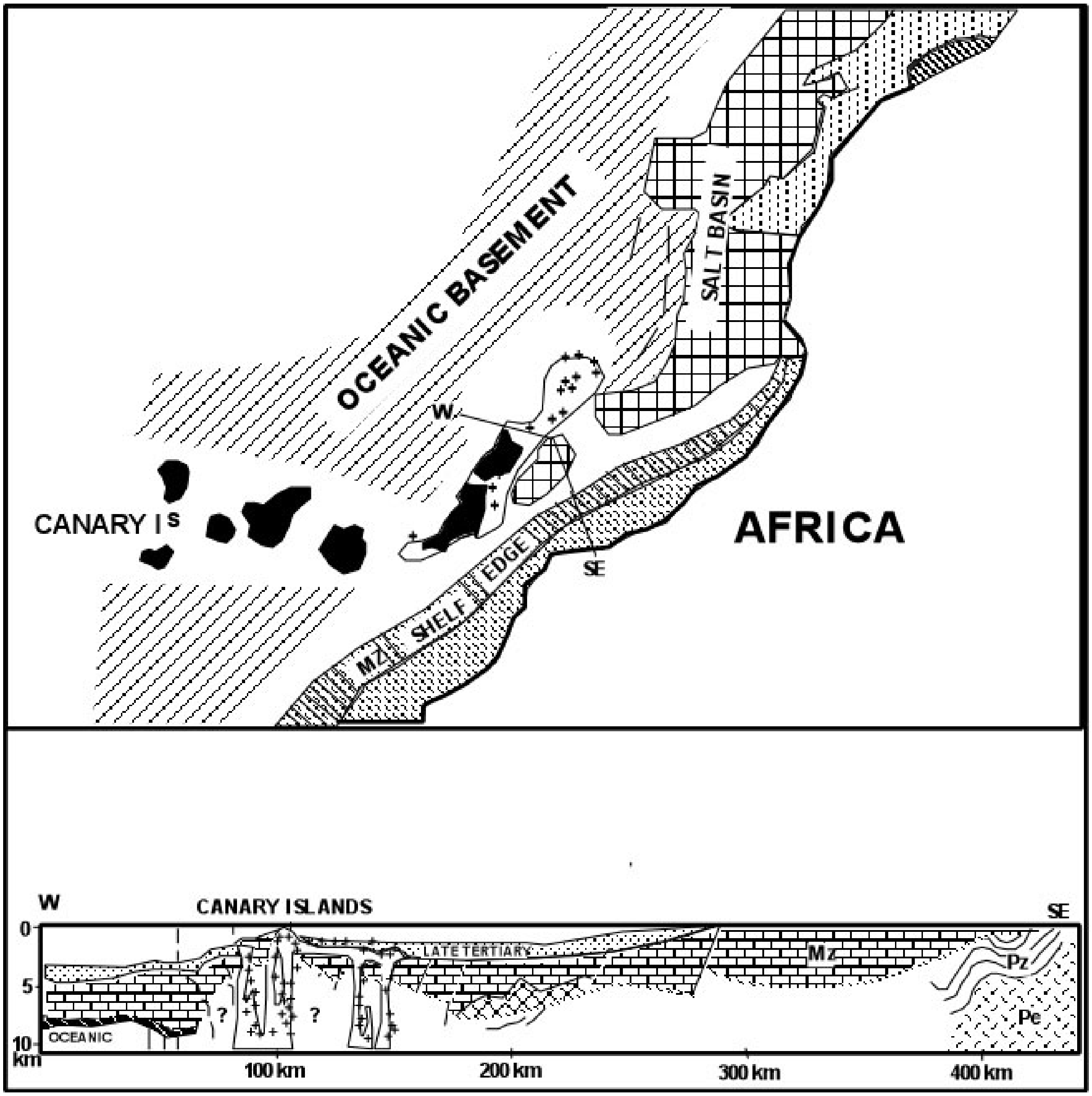


Figure 16 B Sketch of a regional seismic reflection profile across the innermost Canary islands of Fuenteventura and Lanzarote indicating basement uplift beneath the 30 Ma and younger hot-spot volcanic islands. This uplift accounts for the outcrop of Mesozoic ocean floor on the island of Lanzarote which is similar to that in Maio (see Figure 39). Redrawn from Lehner & De Ruiter (1977).

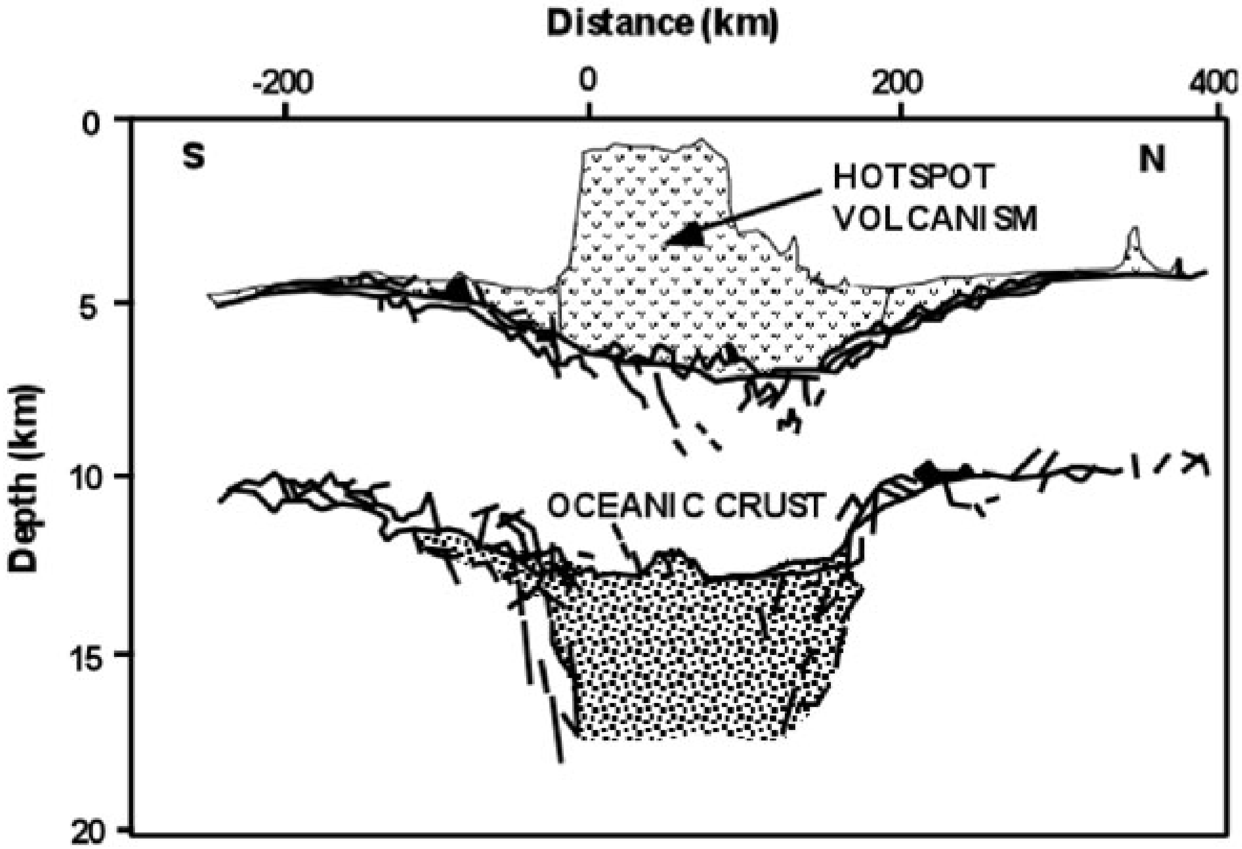


Figure 17 Section across the axis of the Hawaiian swell to illustrate a contrast with hot spots on the African Plate. The crustal thickness in this section reaches an average of ~18.5 km in the central region where a huge load of basalt bows down the lithosphere. African hot spots, both on the continent and on the ocean floor, are associated with basement uplifts and with the eruption of only small volumes of basalt. Some workers have interpreted the gravity field over hot-spot volcanoes on the African Plate in terms of a Hawaiian type structure but this seems inappropriate. Figure simplified from one in Morgan *et al.* (1995), which is itself based on the work of ten Brink & Brocher (1987) and Wessel (1993).

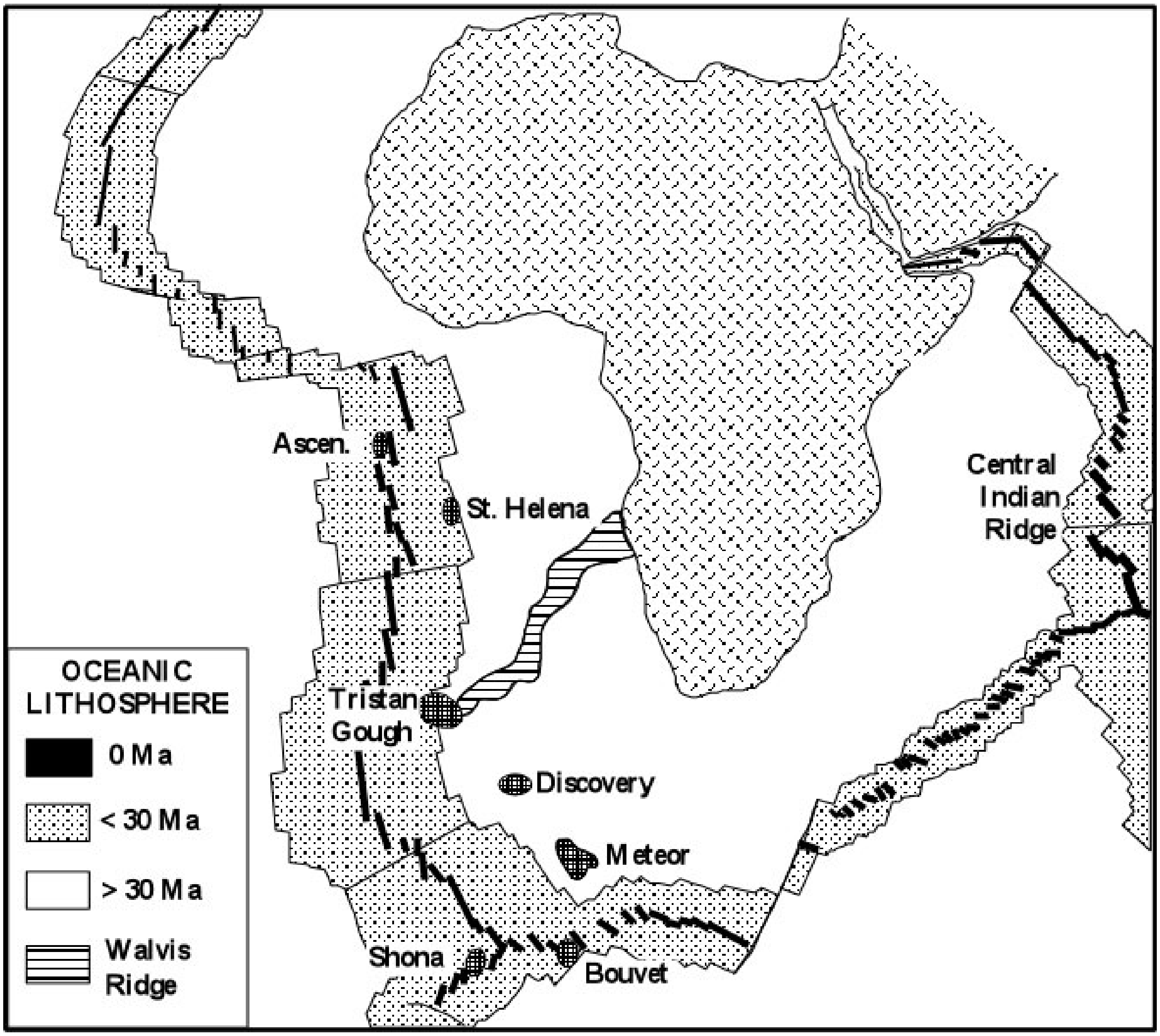


Figure 18 African and conjugate plates showing the distribution of ocean floor formed in the past 30 My. During that interval, the African Plate has not moved with respect to its underlying plume population. St. Helena, Tristan and Gough, Discovery and Meteor form part of the stationary plume population. Ascension, Shona and Bouvet, which lie on oceanic lithosphere generated within the past 30 My on the South American and Antarctic Plates, are new young hot spots. Their compositions are all likely to be similar to that of Ascension which is dominated by the HIMU and MORB sources (Hart *et al.*, 1992).

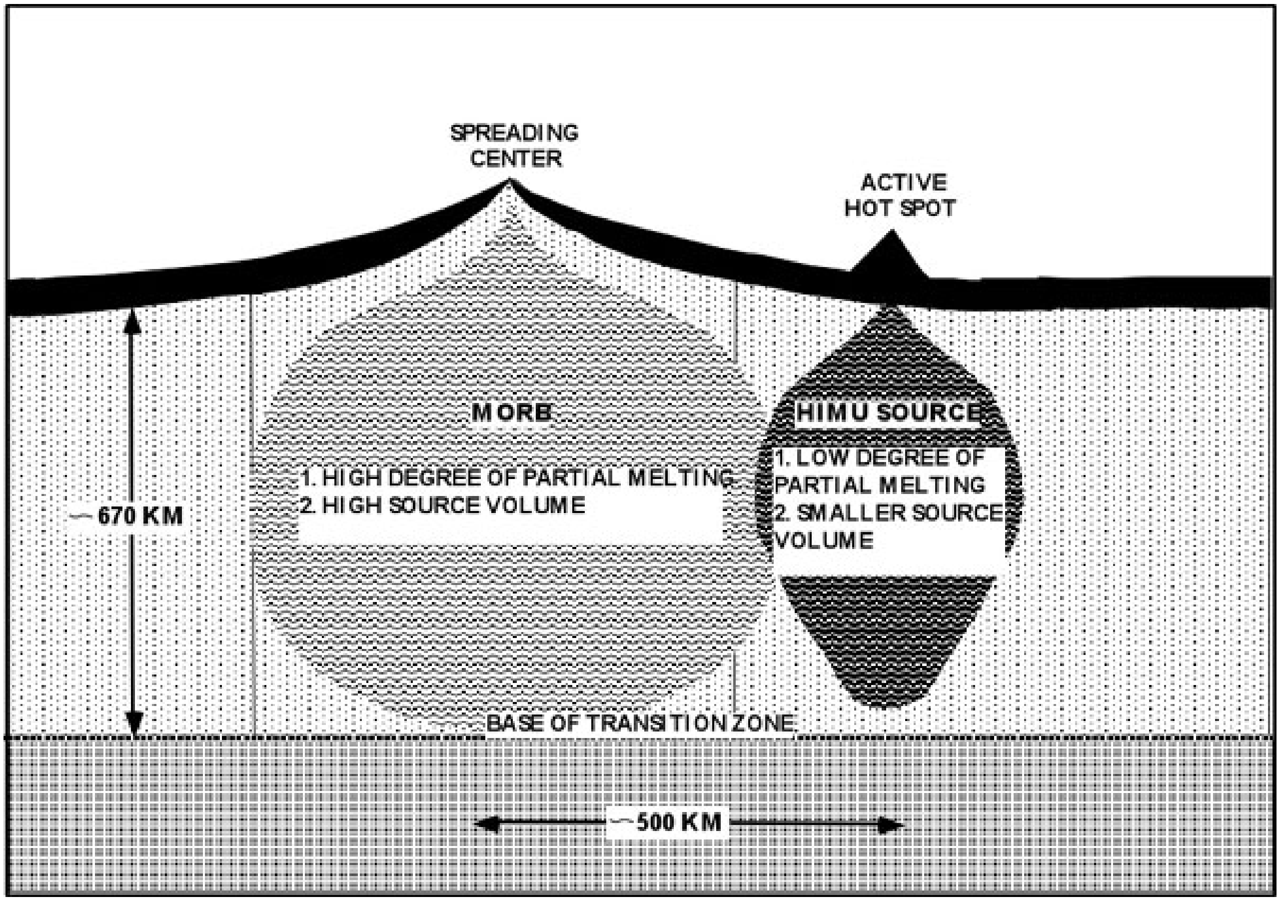


Figure 19 A possible explanation of Schilling *et al.*'s observation (1994) that spreading centre geochemistry recognises the presence of off-axis hot spots hundreds of kilometers away. The spreading center extracts MORB from a huge volume with a relatively high degree of partial melting. A neighboring plume robs that volume to a small extent so that the MORB generated differs from normal MORB in being a product of a smaller volume and therefore appearing more like a plume product.

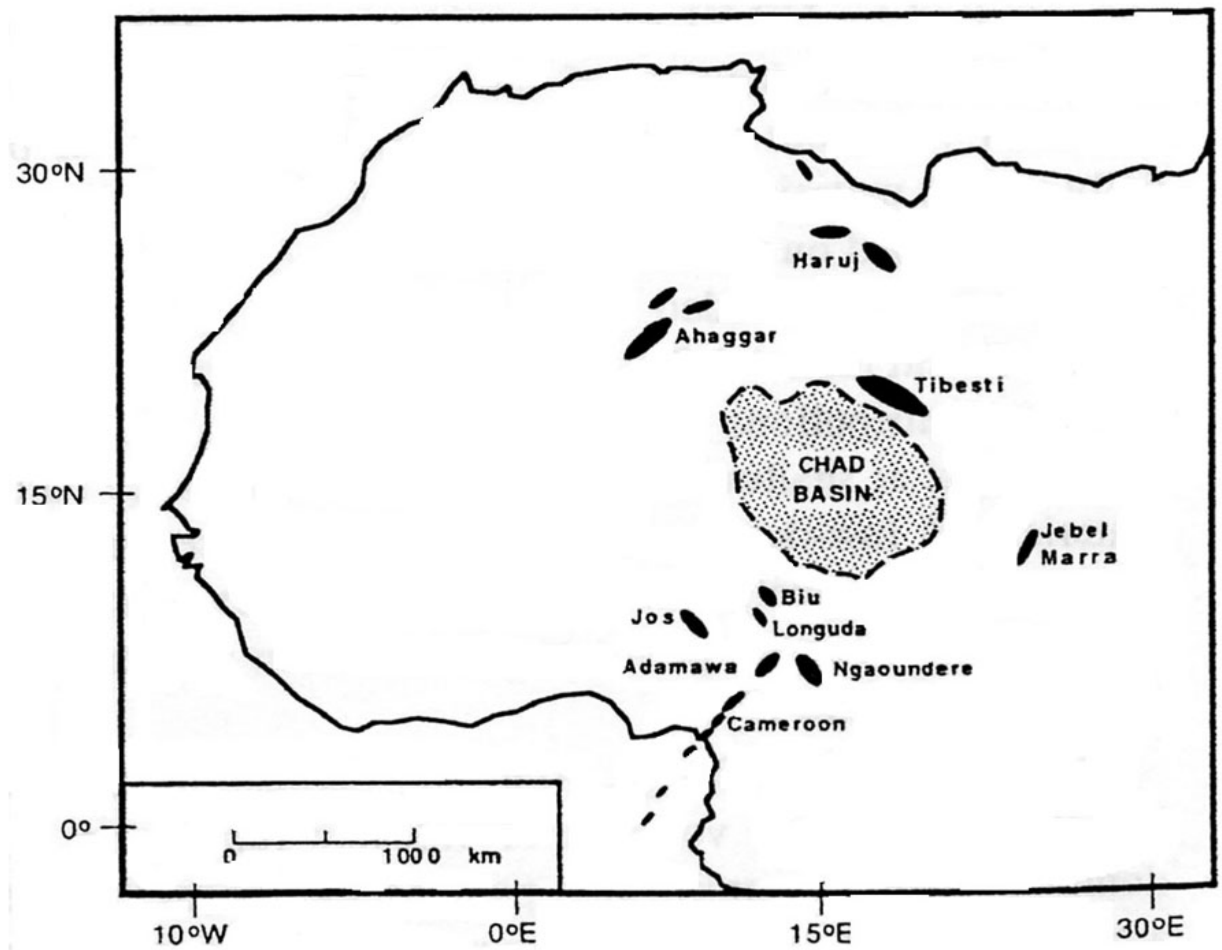


Figure 20 Map showing the distribution of some of the 30 Ma and younger topographic swells of North Africa to illustrate that the Chad Basin is defined by the swells on its periphery. These swells individually resemble the Jos Plateau and the Ahaggar (Figures 14 and 15). Redrawn from Burke & Whiteman (1972).

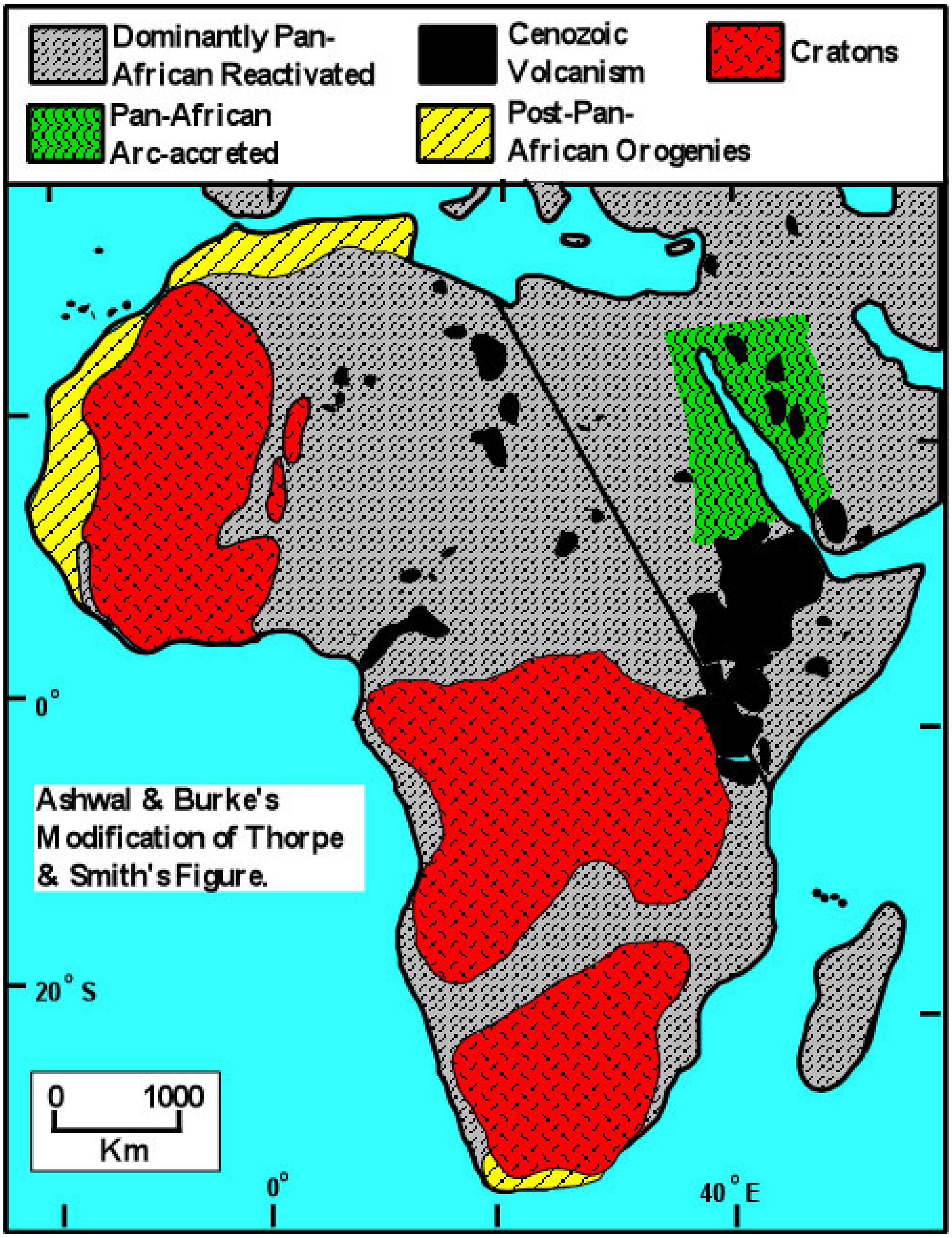


Figure 21 Map illustrating that the 30 Ma and younger volcanic areas of Africa are strongly concentrated in areas involved in Pan-African tectonism. This sketch does not adequately show that in the northern third of the western rift of the East African Rift System volcanism occurs in areas of older cratonic rock. Redrawn from Ashwal & Burke (1989) who based their map largely on Thorpe & Smith (1974).

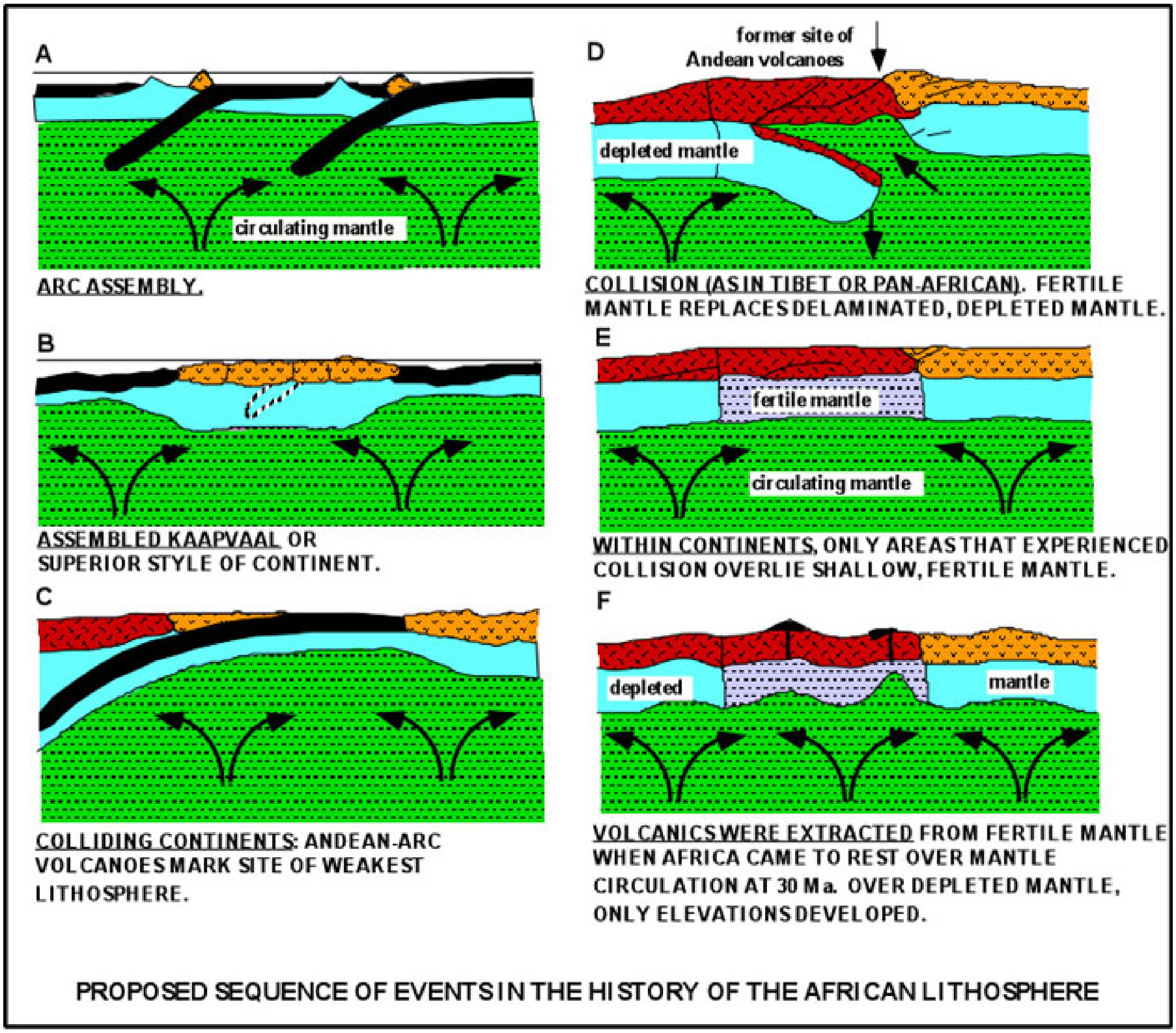


Figure 22 **(A)** Ancient continental masses such as the Kaapvaal Craton consist dominantly of material assembled by the collision of island arcs and microcontinents bearing Andean-type arc margins. **(B)** These continental objects are underlain by deep keels of highly depleted mantle lithosphere as a result of the generation of komatiites. **(C)** Two such ancient continents are likely to collide from time to time as the Earth evolves. **(D)** At continental collision the depleted mantle may be delaminated. Although depleted mantle is buoyant it can be made effectively denser by attached deep crustal rocks which, in a thickened continent, exist as garnet rich assemblages. Fertile mantle lithosphere comes to occupy the place of the delaminated depleted mantle lithosphere. **(E)** Within an old continent such as Africa, areas that have experienced continental collision and delamination, such as those involved in the Pan-African orogeny, are underlain by fertile mantle lithosphere. Areas of arc assembly are underlain by depleted mantle. **(F)** When the African Plate came to rest over the plume population at 30 Ma, interaction with the mantle lithosphere beneath areas of Pan-African crust generated volcanic rocks. Interaction over the same interval beneath areas of arc assembly, such as the Kaapvaal Craton, has generated only topographic elevations. Figure based on Ashwal & Burke (1989).

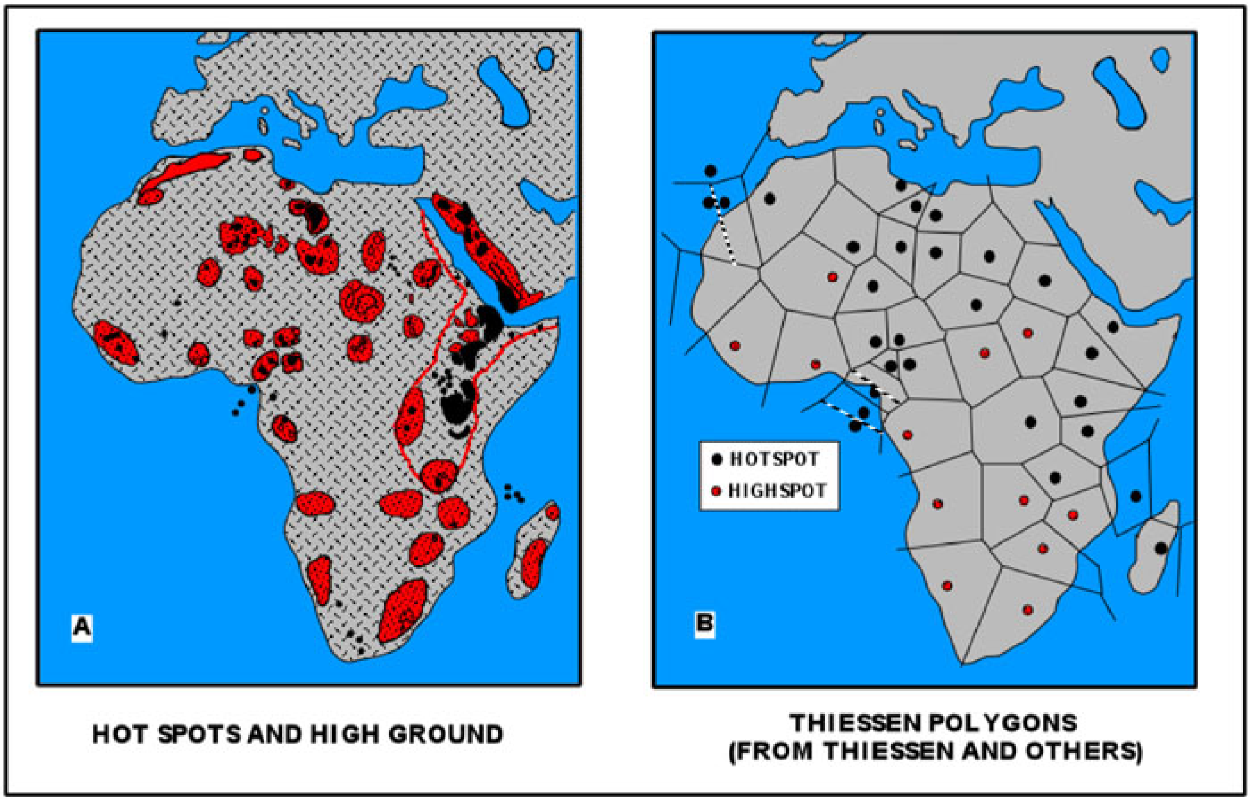


Figure 23 R. Thiessen *et al.* (1979) constructed polygons in the manner of A. Thiessen (1911) in an attempt to look for regularity in the separation of hot spots and high spots (crests of swells without volcanic rocks) on the African continent.

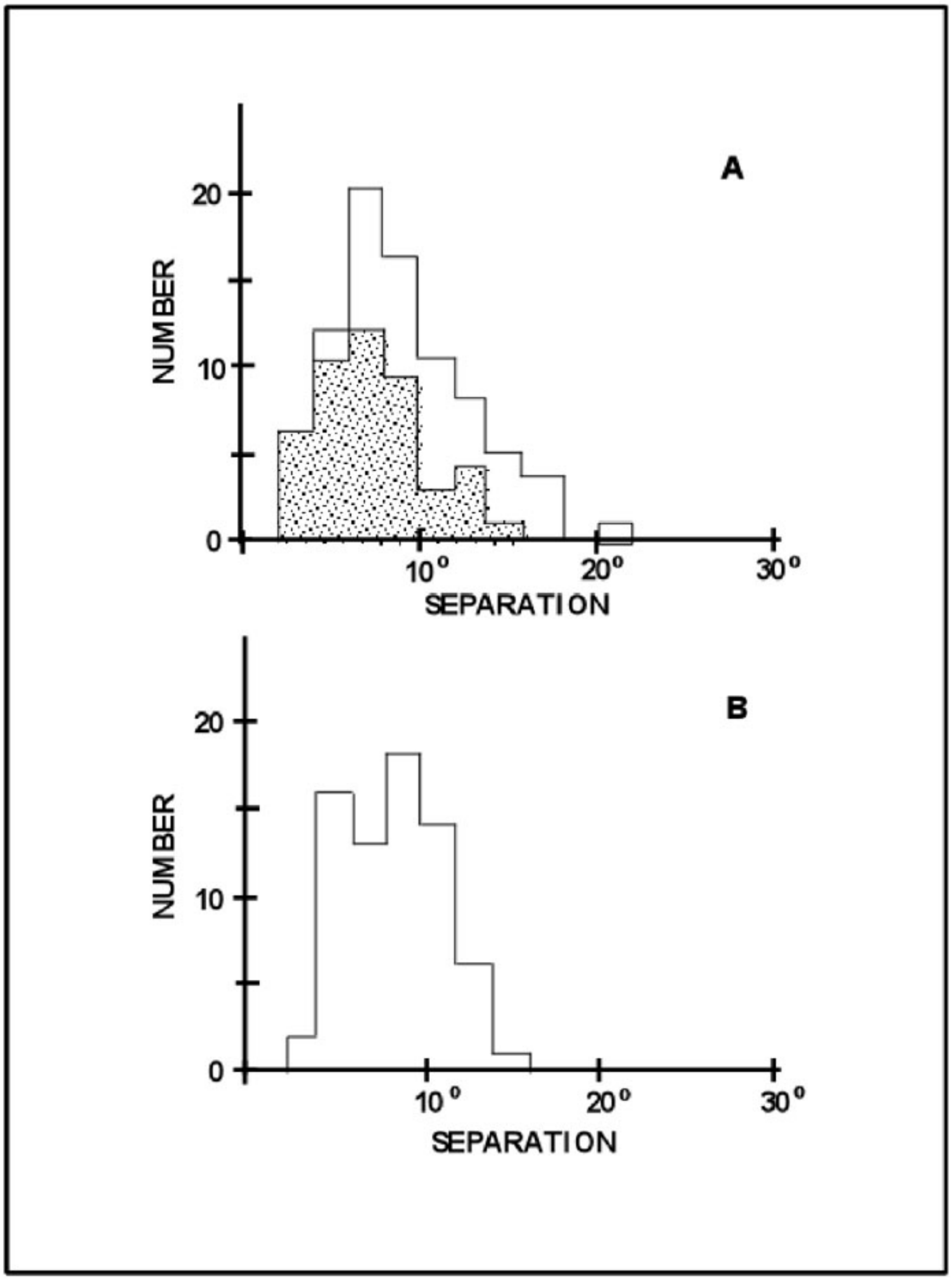


Figure 24 **(A)** Histograms generated by R. Thiessen *et al.* (1979) showing the separation of hot spots (open columns) and high spots (stippled columns) on the African continent. Separation is shown in units of degrees of arc or ~111 km. **(B)** For comparison, the separation of rising plumes in the mechanical model of Richter & Parsons (1975) is shown scaled to a 500 km thick convecting layer. Assuming an average of ~150 km for the thickness of the lithosphere, this value corresponds roughly to the depth from the surface to the base of the transition zone.

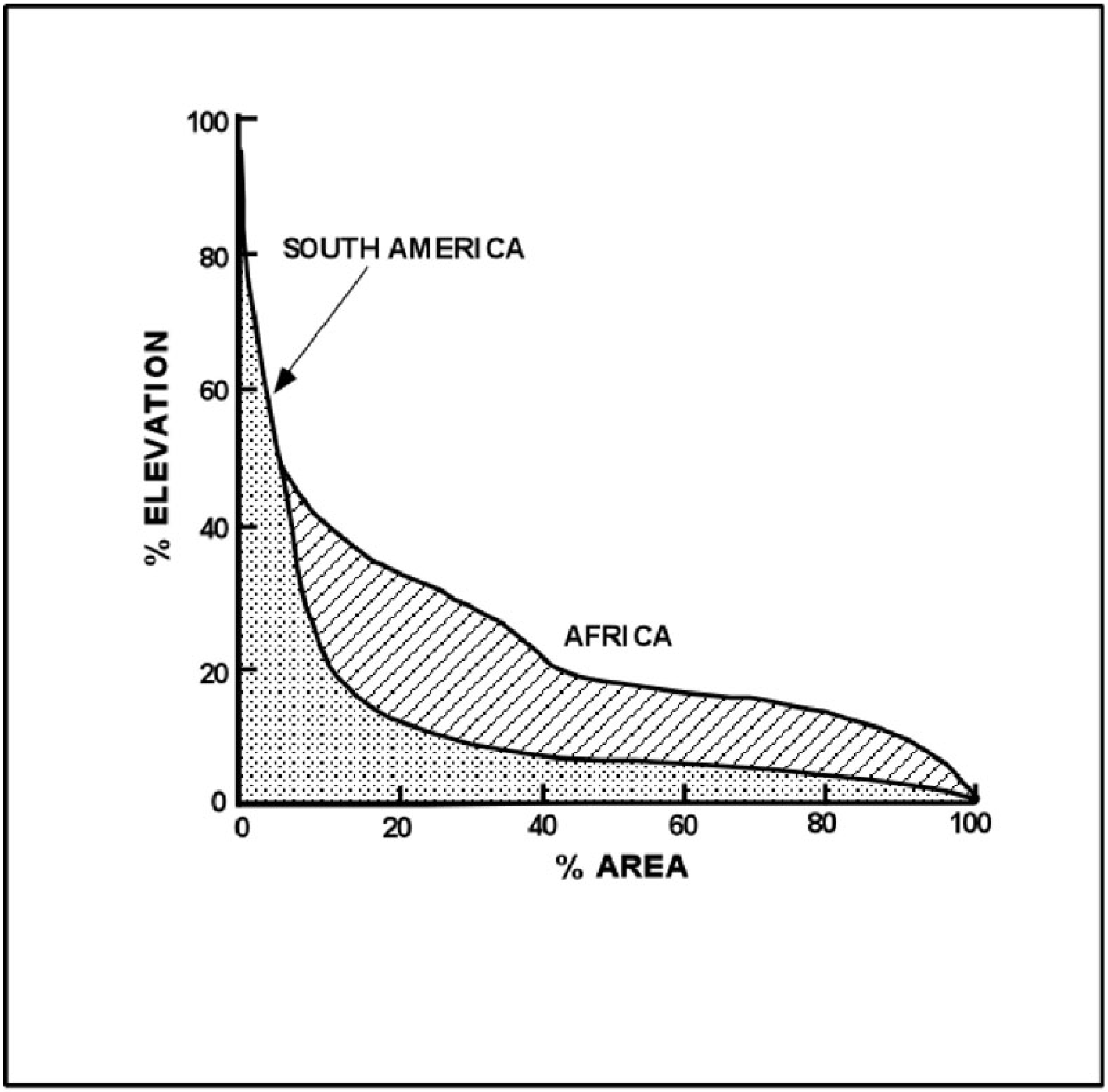


Figure 25 Comparison between hypsographic curves for Africa and South America serves to emphasise that although Africa has no mountain ranges as high as the Andes, an unusually large area of the continent is moderately elevated (figure redrafted from Cogley, 1987). The very small area of Africa close to sea level contrasts strongly with the extensive low-lying area of South America representing the Orinoco, Amazon and Plate Basins. The anomalous elevation of Africa is attributable to interaction between the stationary African Plate and underlying mantle plumes during the past 30 My.

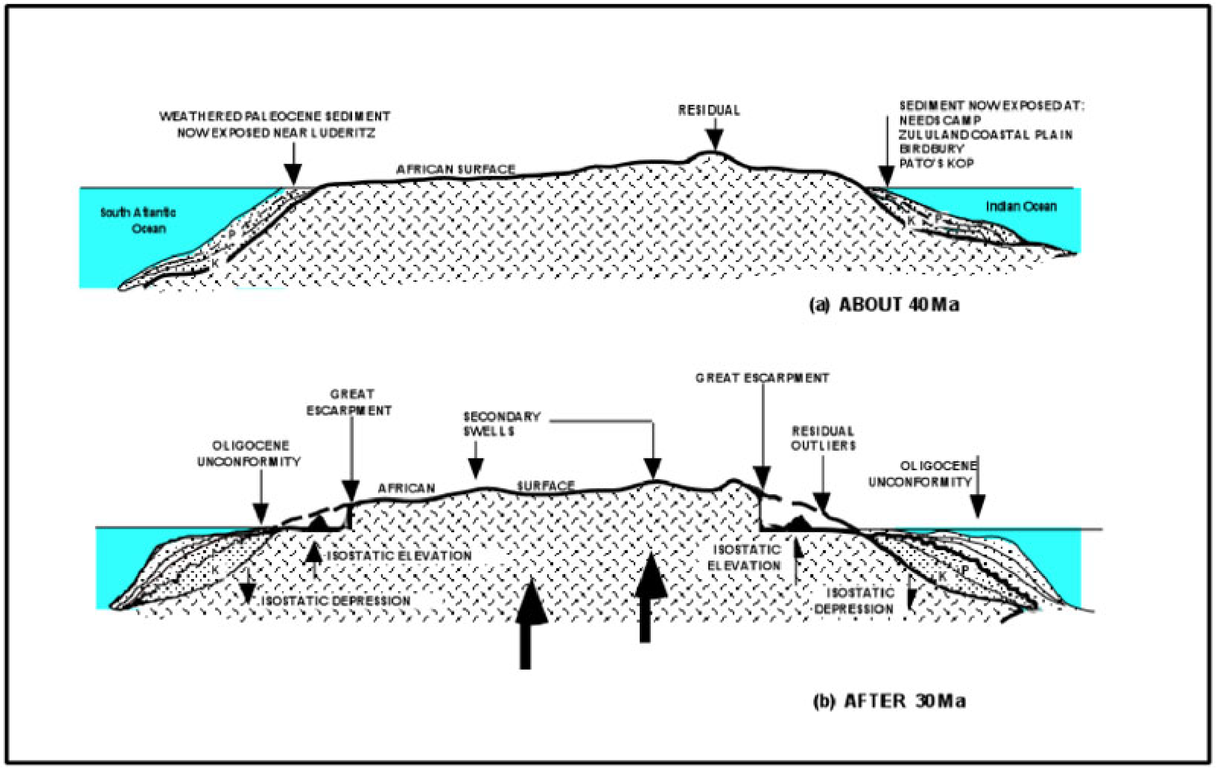


Figure 26 Sketch cross-sections illustrating the topography of southern Africa (a) at about 40 Ma, before the African Plate came to rest over the mantle circulation, and (b) after ~30 Ma, the time at which the plate stopped moving with respect to at least part of the mantle circulation. In the upper figure, the African surface is shown as lying close to sea level without many or prominent residual elevations. Cretaceous and Paleogene sediments (K and P) are indicated at the rifted margins of the continent. In the lower figure, the African surface is shown as elevated on the Great Swell of southern Africa whose uplift is represented by two large black vertical arrows. Two secondary or subswells are indicated. The Great Escarpment is beginning to cut back into the continent and sediments are beginning to be deposited offshore above the mid-Oligocene unconformity. Offshore, the newly accumulated sediments cause isostatic depression and onshore, removal of load by erosion causes isostatic elevation. Contemporary effects of world-wide sea level change and flexural effects (Mc Ginnis *et al.*, 1993), which amplify the results of uplift of the Great Swell of southern Africa near the coast, are not indicated separately in this sketch.

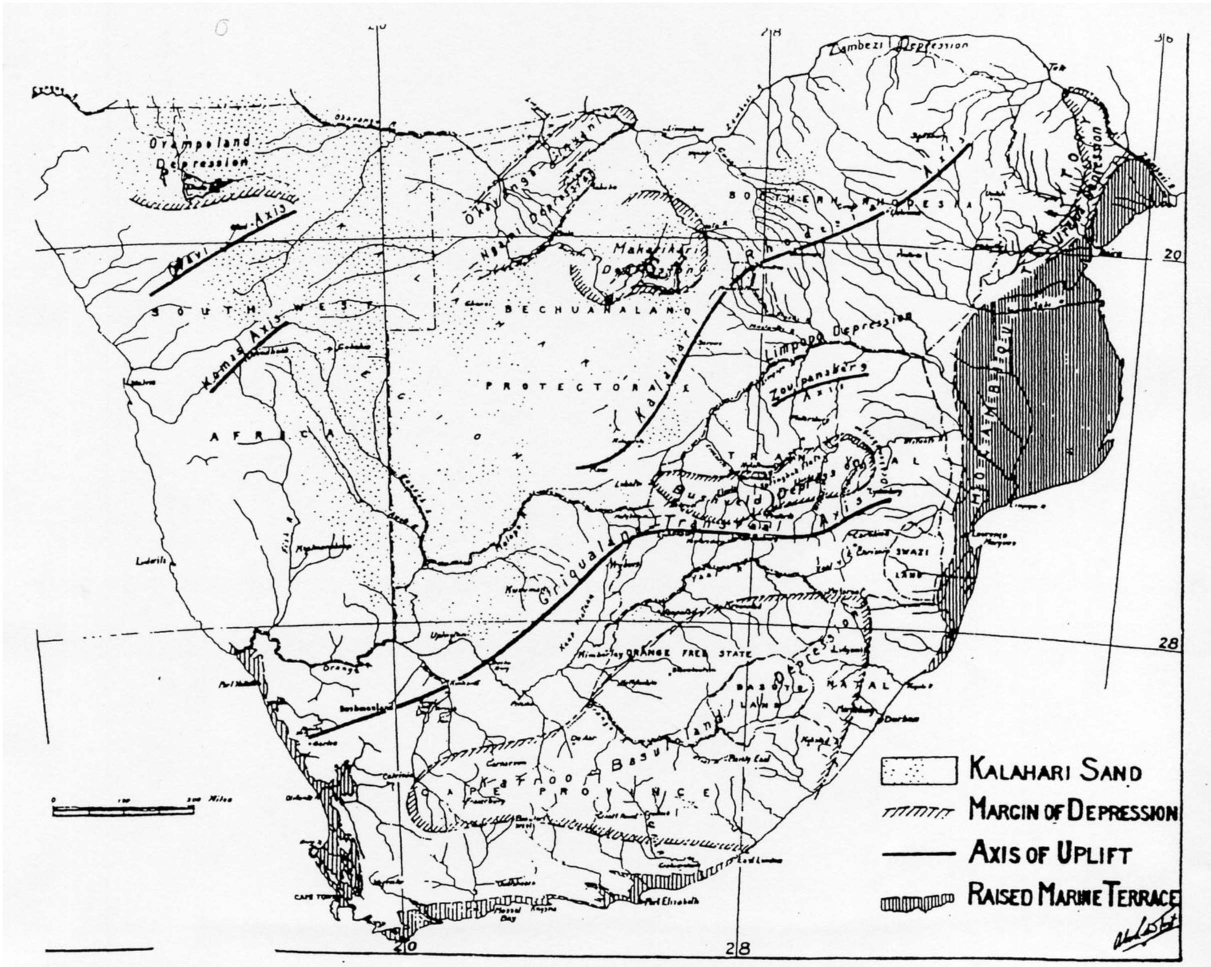


Figure 27 Figure reproduced from Alex L. du Toit's Presidential Address to the Geographical Society of South Africa (du Toit, 1933). The figure shows axes of uplift which correspond to the subswell crests of this paper. Du Toit was impressed by evidence of Quaternary tectonism.

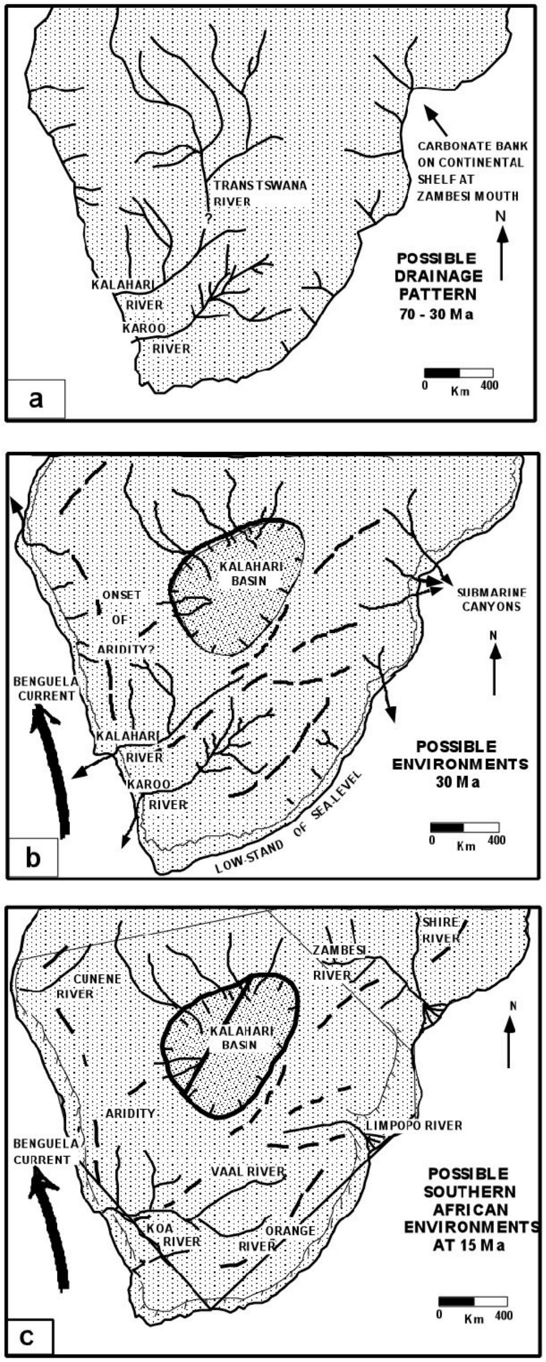


Figure 28. Sketch maps illustrating possible drainage systems and related environments for southern Africa at three times: before 30 Ma, at ~30 Ma, and at ~15 Ma. These figures are largely based on information in Dardis *et al.* (1988) and de Wit (1993), interpreted here using the ideas (1) that the uplift of the Great Swell of southern Africa began at about 30 Ma when the African Plate stopped moving over the mantle circulation, and (2) that the Great Escarpment has developed since that time. In Figure 28a, the drainage of the southern continent is shown as dominated by Trans Tswana and Karoo river systems. The coastline shown is that of today except at the mouth of the Zambesi where post 30 Ma rapid deltaic progradation (Droz & Mougenot, 1987, figure 13) has drastically changed the shape of the coastline. The existence of a carbonate bank at the Zambesi mouth in Paleogene time was suggested by De Buyl & Flores (1986, figure 19 and p. 417). Figure 28b represents conditions at about 30 Ma when sea level was low worldwide, the African continent began to be elevated and the Benguela current had begun to flow (Siesser, 1978) leading to the onset of aridity in the southwestern part of the continent. Submarine canyons were probably abundant and the positions of a few are indicated. The Kalahari Basin of interior drainage had begun to develop. Possible axes of subswells are indicated by dashed lines. Figure 28c shows conditions intermediate

between those of 30 Ma and the present as the Great Escarpment was in retreat and drainage development approached that of today.

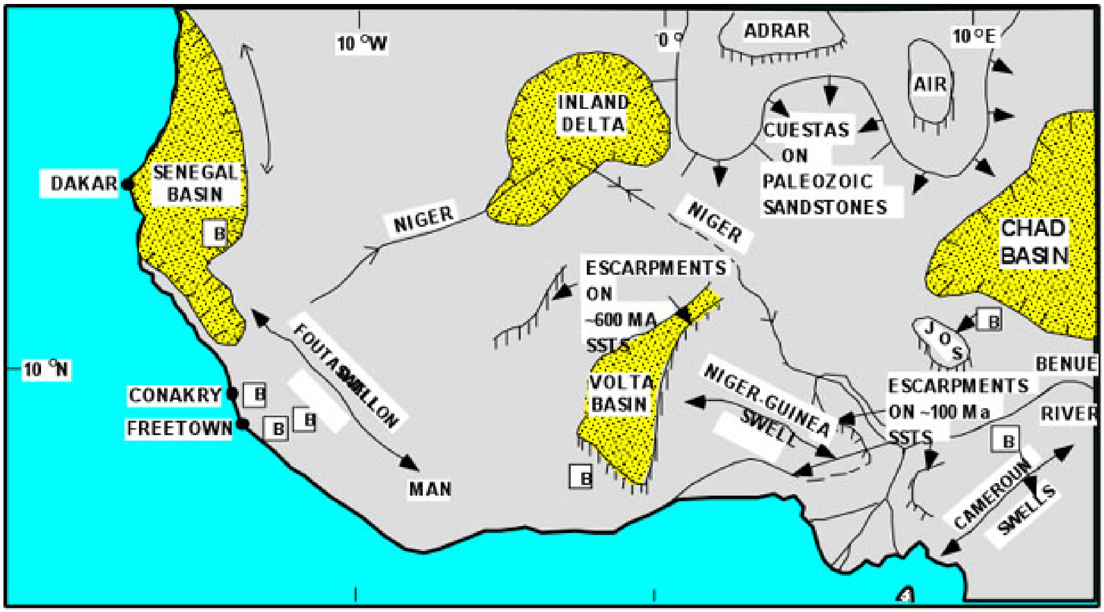


Figure 29 Basins and swells of West Africa. The interior basins of the Inland delta region of the Niger (Reclus, 1888) and the Chad Basin (Figure 20), as well as the coastal Senegal Basin, are set among 30 Ma and younger swells. Swells include the Fouta Djallon, the Niger-Guinea watershed, the Cameroun chain of several discrete swells, as well as the interior Jos (Figure 15), Air and Adrar swells. Because the uplifts of the swells have not been as intense as in much of eastern and southern Africa, escarpments and cuestas are developed around gently dipping sedimentary rock piles of various ages. The Benue and Niger rivers dominate drainage. The latter sometimes flows no farther than the inland delta and its middle reaches may flow in either direction. Localities marked with the letter B are bauxite localities, mainly from Egbogah (1975).

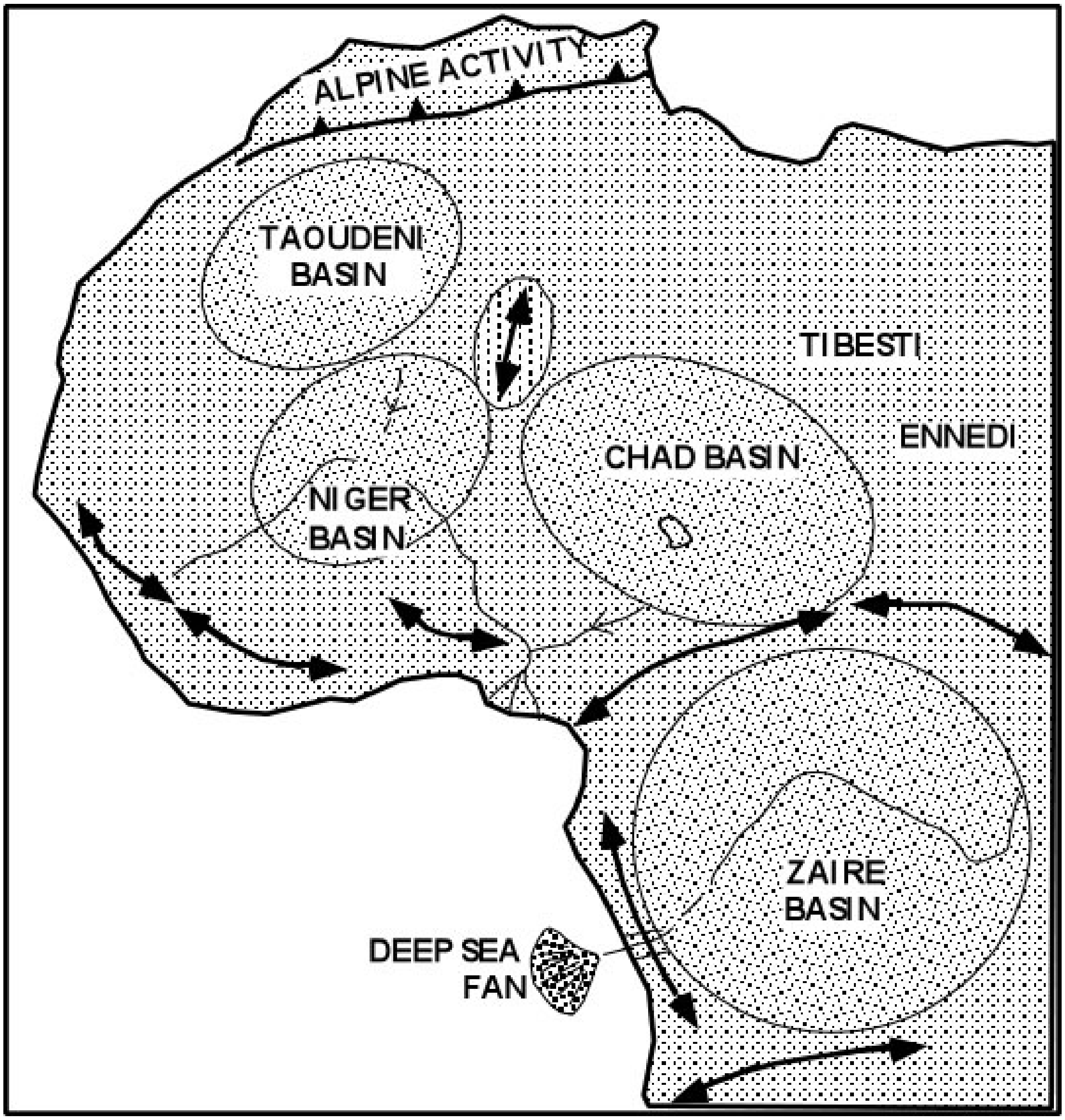


Figure 30 Interior basins and swells of northern and western Africa. The plate boundary zone of the Atlas Mountains on the northern border of the continent is not discussed in this paper. Summerfield (1996) has emphasised the importance of uplifts (indicated by lines with arrows at both ends) close to the coast which are particularly prominent in the region sketched on this map. The Zaire Basin is similar to the interior basins, but river capture has linked it to the ocean where a spectacular submarine canyon and deep sea fan have developed.

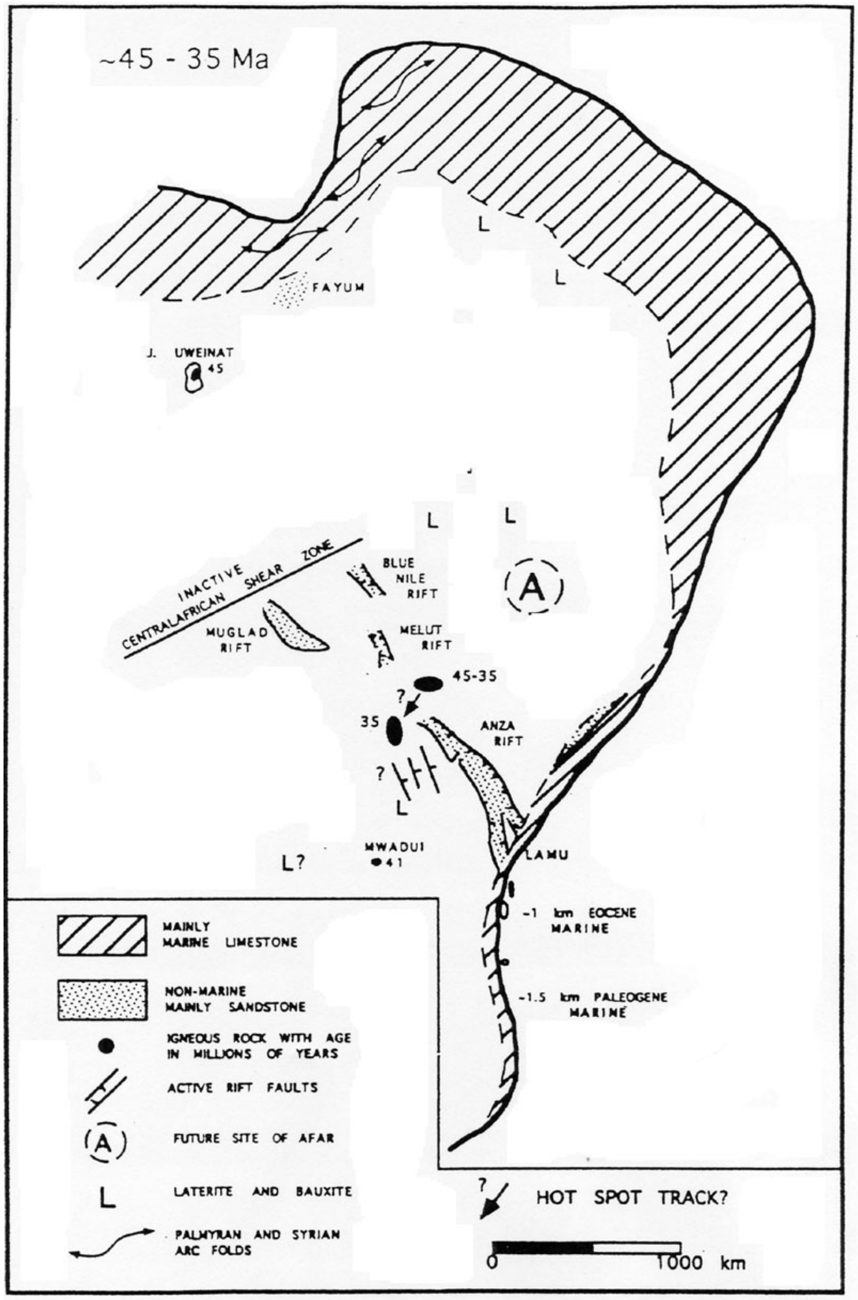


Figure 31 Sketch map showing the area over which the East African Rift System would develop starting at ~30 Ma. Arabia at this time formed a promontory of the African continent. This map illustrates conditions during later Eocene times between ~45 and ~35 Ma. Four Mesozoic rifts continued to receive sediments and some rift faults were active. Most of the interior of the continent was low-lying, and laterites (L), which formed at this time or somewhat earlier, are locally preserved. Kimberlite was erupted at Mwadui and granite was emplaced at Jebel Uweinat. Plume generated flood basalts were erupted in an area in S. Ethiopia over most of the 45 Ma to 35 Ma interval. By 35 Ma, a plume—possibly the same plume—was beginning to erupt basalt in an area in Lokitipi where eruptions have continued ever since. Palmyran and Syrian arc folds had just ceased to be active in Eygpt and the Levant. Shallow-water marine sandstones and limestones are widely but sporadically distributed over the northeast corner of the continent and a well-developed shoreline is exposed at Fayum. On the east coast of the continent thick Paleogene sections off Tanzania may indicate the existence of a river system draining the interior of East Africa. The future site of impact from below the Afar plume is indicated.

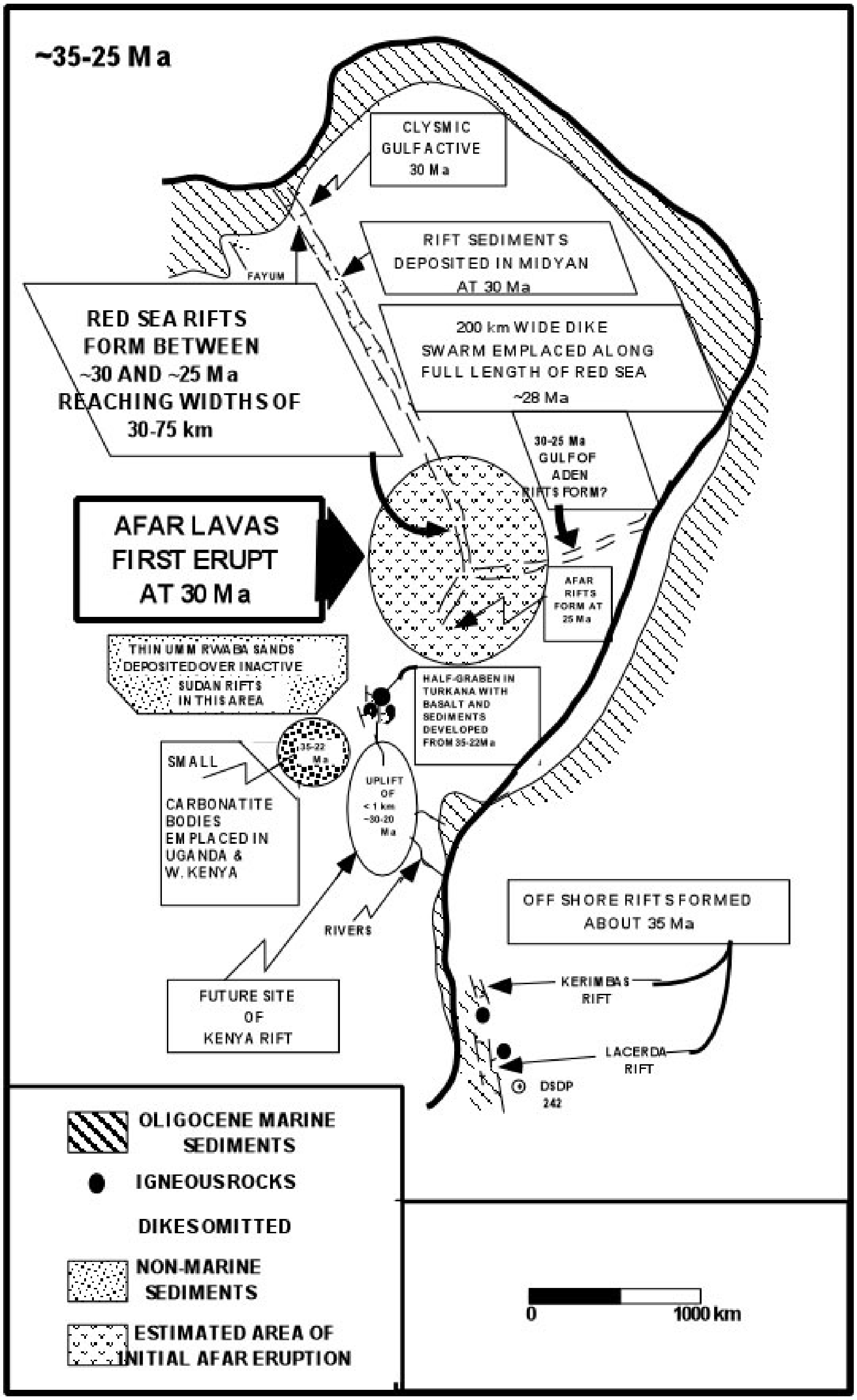


Figure 32 Sketch map showing the inception of the East African Rift System during the 10 My interval 35 to 25 Ma. The critical event may have been the eruption of the Afar plume at ~30 Ma to form the Ethiopian traps. The Afar plume has been persistently active at the same site ever since. The Red Sea Rift, including Suez, formed at ~30 Ma and both rift sediment deposition and the emplacement of a dike swarm accompanied early rift development. Rifts in the Afar had formed by 25 Ma and the Gulf of Aden Rift may have formed simultaneously. In Turkana half-graben with basalts and sediments developed over the site of the Lokitipi eruptions and also over the Anza Rift. Other Mesozoic rifts in Sudan became inactive and were overlain by thin sands of the Umm Rwaba formation. A Kenya dome of moderate elevation formed and products of erosion from that dome may be represented in the Turkana grabens and in the coastal rift basins of Tanzania. Offshore the Kerimbas and Lacerda Rifts formed in water depths of 1 km and the igneous rocks of Paisley and St. Lazare seamounts in those rifts were erupted.

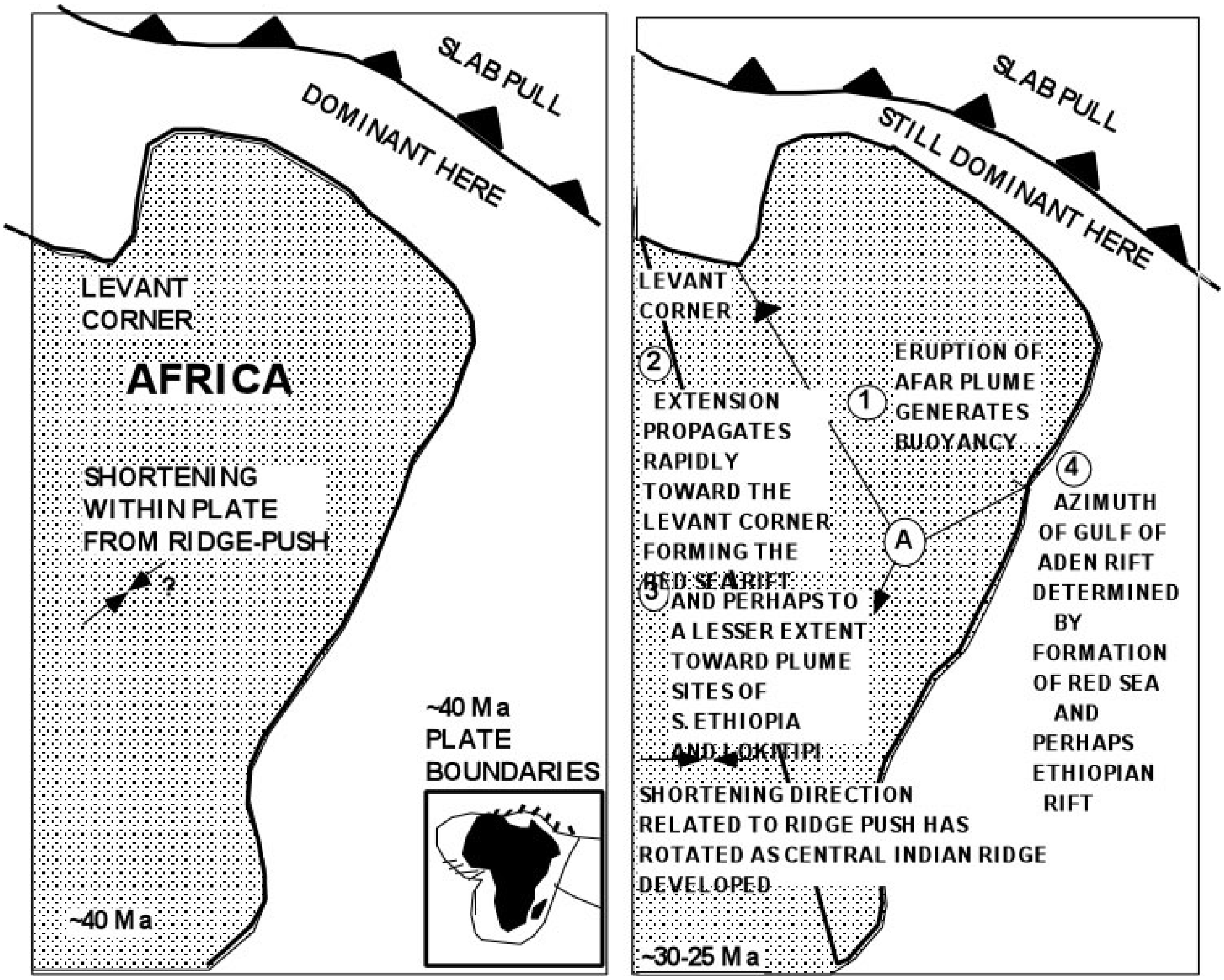


Figure 33 Sketch illustrating how the geometry of the Levant Corner may have played a critical role in the evolution of the Red Sea Rift. The stress field normally associated with rifted continental margins was amplified, because of the right angle, in the region of the Levant Corner. Eruption of the Afar plume generated buoyancy over an area perhaps 1000 km in diameter modifying the within-plate stress field. Propagation of the Afar extensional stress field, preferentially toward the Levant corner, led to formation of the exceptionally long and straight Red Sea Rift segment. Propagation of stress from the Afar in the direction of the Lokitipi plume may not have been throughgoing at this time but the azimuth of the Ethiopian rift may have been determined by that process. The azimuth of the Gulf of Aden Rift may simply record a direction dependent on the initiation of rifting in the Red Sea and the Afar.

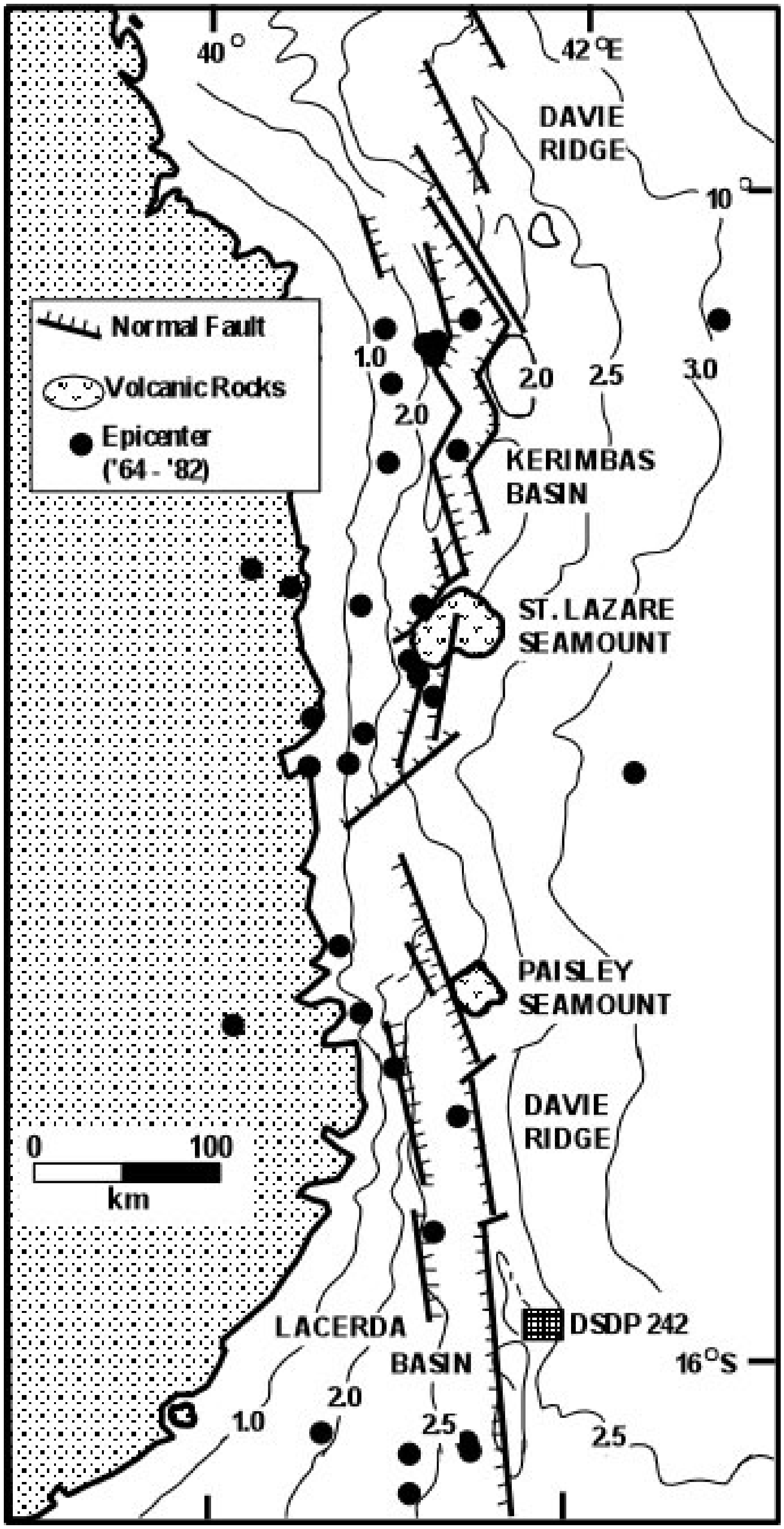


Figure 34 The Kerimbas and Lacerda Rifts offshore of Mozambique in about 1 km water depth. Figure redrawn from Mougenot *et al.* (1986). The rifts appear to have originated at ~35 Ma from the correlation of reflectors on seismic lines with DSDP 242. Volcanic rocks occupy the two seamounts and earthquakes indicate that the rifts are active today. Submarine contours at 0.5 km intervals.

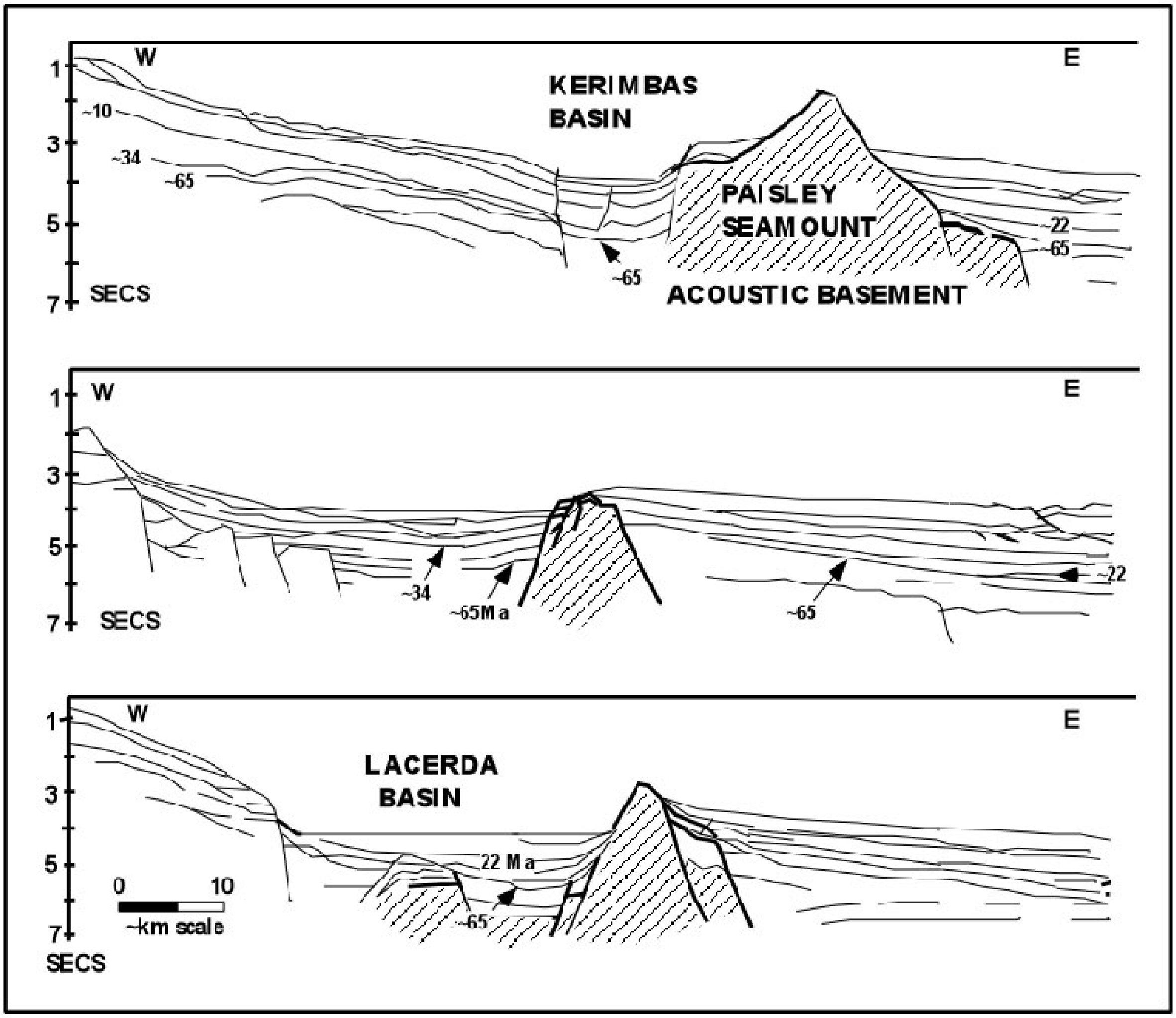


Figure 35 Cross-sections showing line drawings of seismic reflection lines through the Kerimbas and Lacerda basins to illustrate the style of active rifting. Figure redrawn from some of the lines illustrated in Mougenot *et al.* (1986, figure 2). Numbers assigned to reflectors are ages in My.

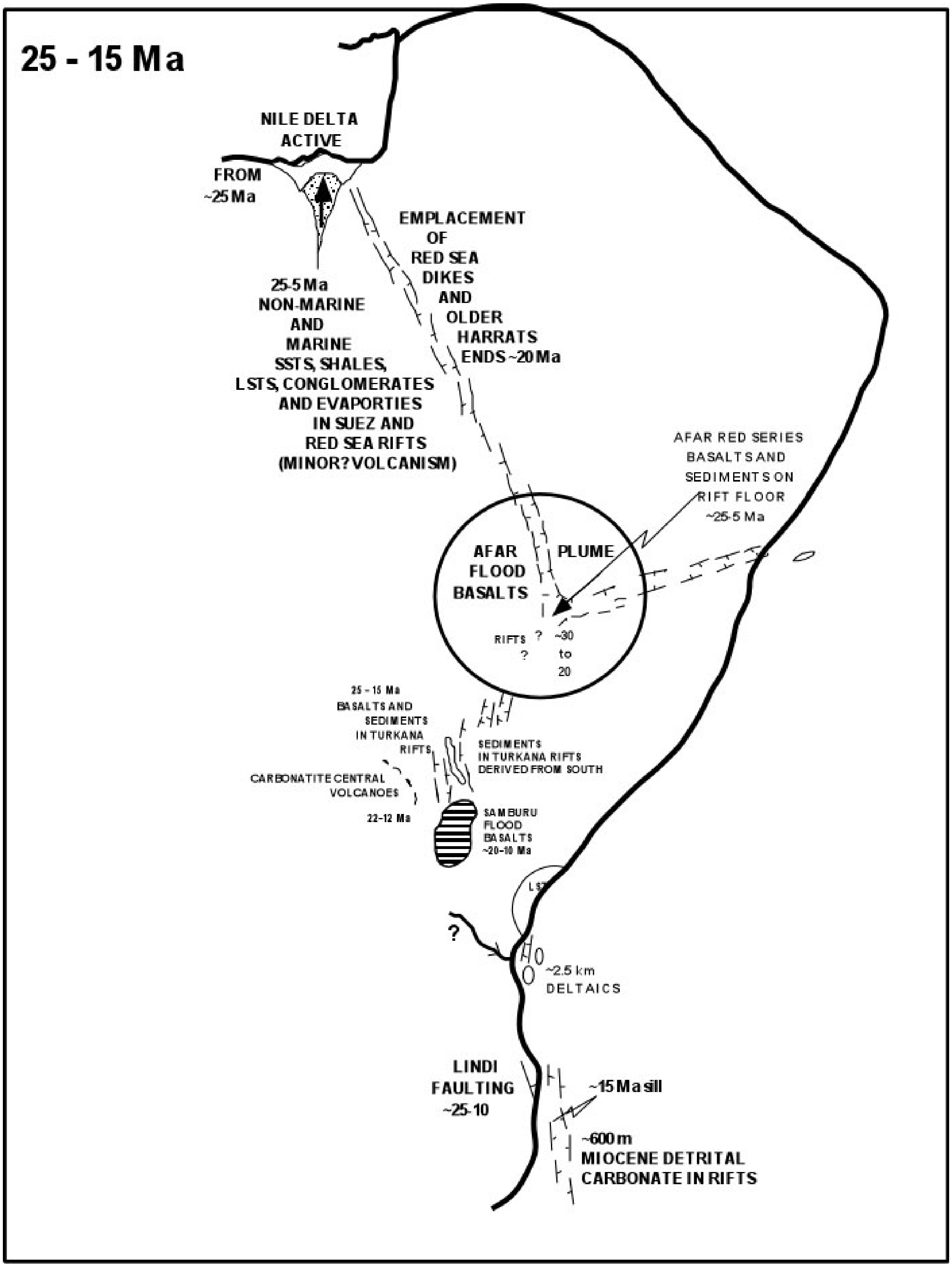


Figure 36 During the interval from 25 Ma to 15 Ma, the East African Rift System evolved without great change. The newly formed Nile river constructed a delta on the Mediterranean coast about 150 km south of its present location. Activity in Turkana persisted in its established style and the Samburu plume erupted about 1 km of flood basalt on the future site of the Gregory Rift. The coastal rift system of Tanzania continued to develop with a thick accumulation of deltaic sediments. Farther south the Kerimbas and Lacerda Rifts and their onshore neighbor, the Lindi fault system, continued to be active.

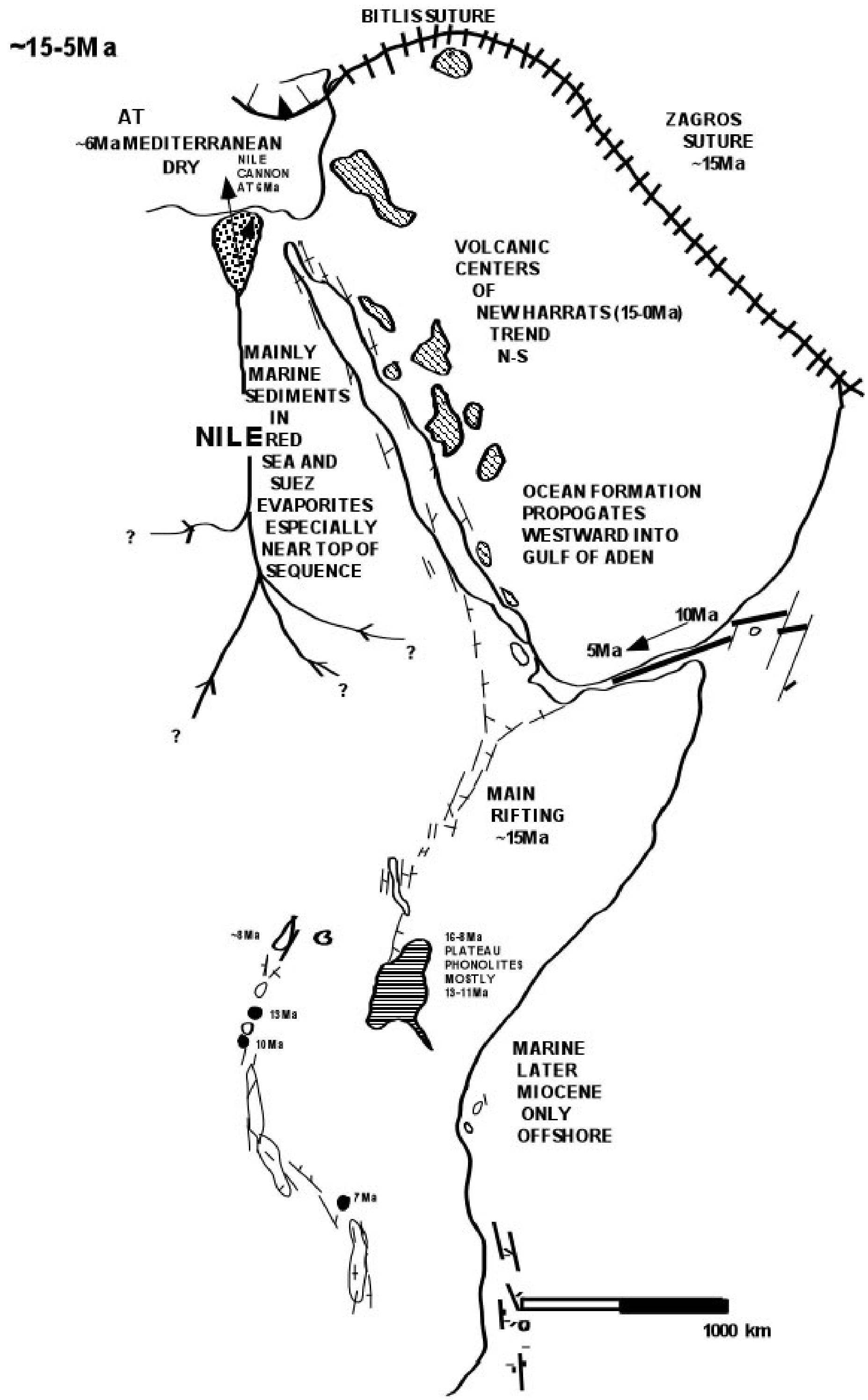


Figure 37 At 15 Ma the East African Rift System developed many features which persist today. The collision of Arabia with Asia was immediately followed by the first eruptions of the new harrats. North-south trending lines of volcanic cones indicate a new stress distribution in Arabia and the propagation of ocean floor formation into the Aden Rift for the first time at 10 Ma may be a related phenomenon. Within-rift sedimentation and volcanism in the Red Sea persisted without great change and the Nile delta continued to prograde into the Mediterranean. Main rifting in the Ethiopian Rift took place at 15 Ma but there was no great change in Turkana. The Samburu plume erupted a kilometer of plateau phonolites but the Gregory Rift was not yet a coherent structure, although establishment of the Elgeyo fault by 10 Ma provided a link between Turkana and the Gregory Rift area. The time of origin of the Western Rift is not well established. 15 Ma is chosen here as the time of origin because it is slightly older than the oldest dated volcanic rocks of the rift in the Virunga.

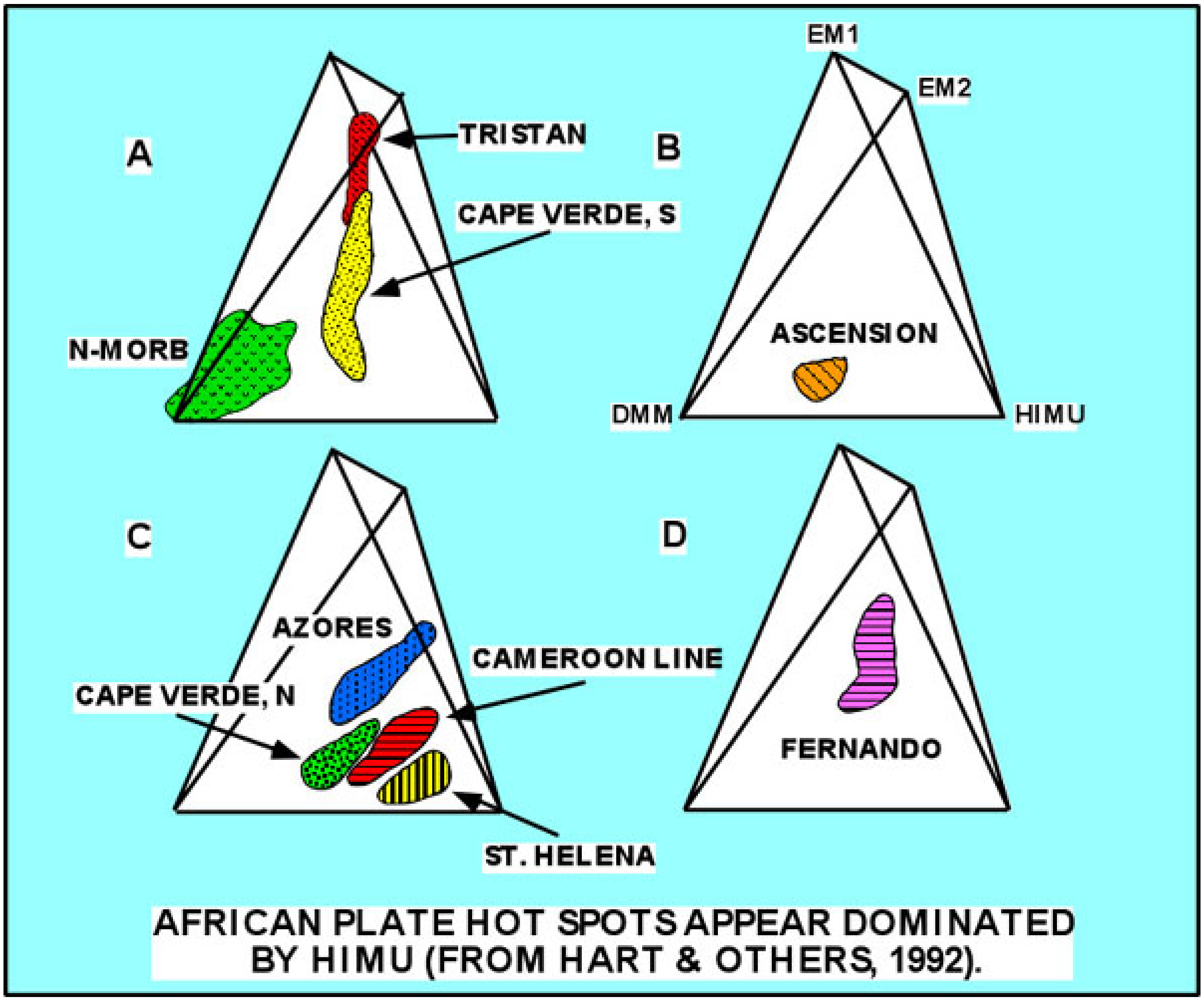


Figure 38 Isotopic compositions of hot spots plotted in terms of the EM1, EM2, HIMU and DMM sources. The figure is modified from Hart *et al.* (1992). Analyses of rocks from St. Helena, Cameroun line volcanoes and the northern Cape Verde islands plot close to the HIMU end member. Ascension, which is very near to the South Atlantic spreading center on ocean floor much less than 30 Ma, plots as a mixture between DMM (which is the MORB source) and HIMU. Tristan, which is not part of the young population of African hot spots, plots close to the join between EM1 and EM2. Only the southern Cape Verde islands among 30 Ma and younger African Plate hot spot volcanoes yield some analyses indicating involvement of EM1 and EM2 as well as the HIMU source. I interpret the results plotted in this figure as indicating that the HIMU source dominates in 30 Ma and younger African Plate hot-spot volcanic rocks.

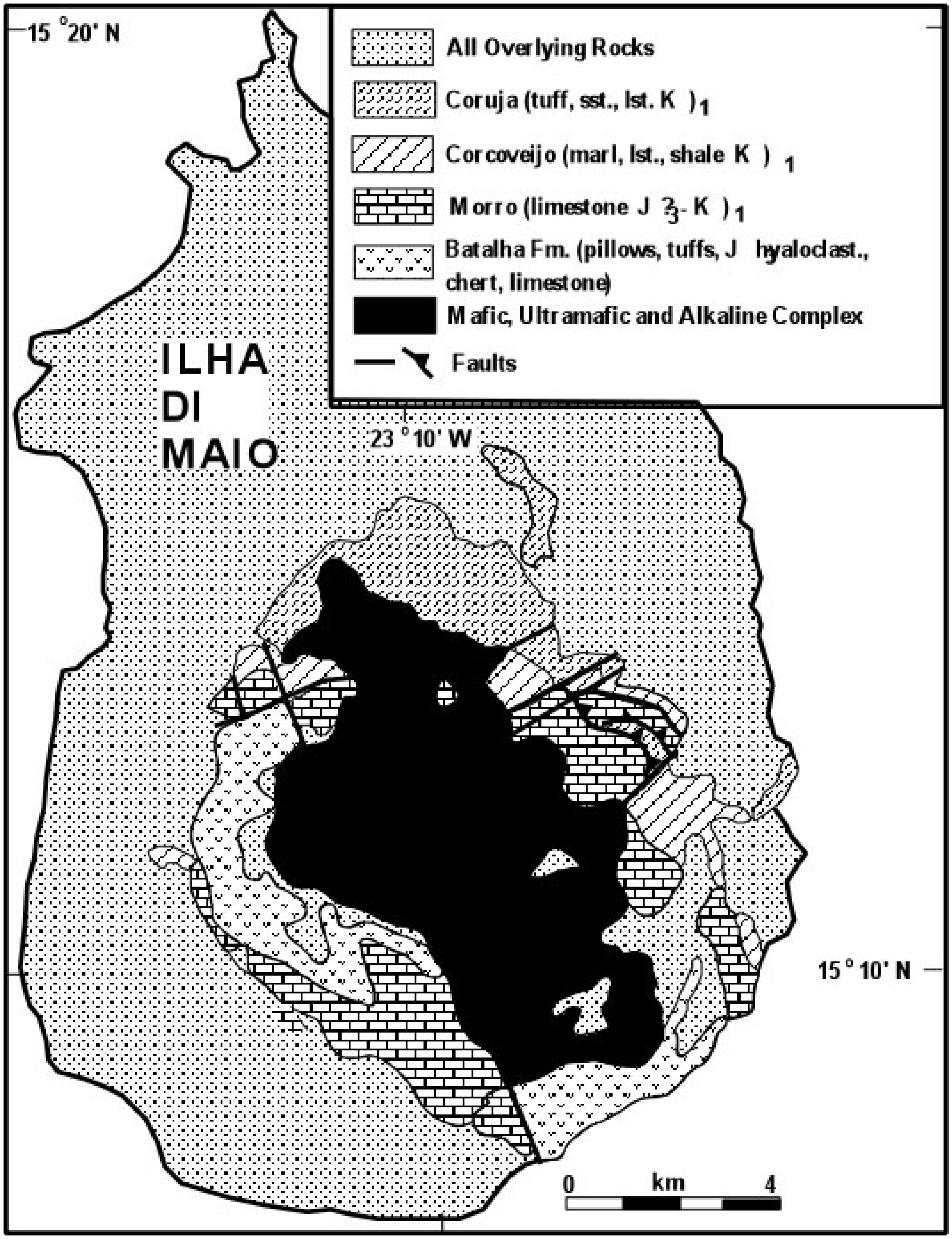


Figure 39 Sketch map simplified from Akhmet'yev *et al.* (1985) showing the uplifted Jurassic ocean floor and the overlying Jurassic and Cretaceous pelagic sediments which outcrop in the center of the island of Maio in the Cape Verde archipelago. The area occupied by the Miocene volcanic rocks erupted from the Cape Verde plume, which overlie the basement, is grey in this sketch. The outcrop pattern indicates the important role of basement uplift and small eruption volume in an African Plate oceanic hot spot. This relationship is familiar on the continent of Africa and may be common, but hard to discern on oceanic islands.

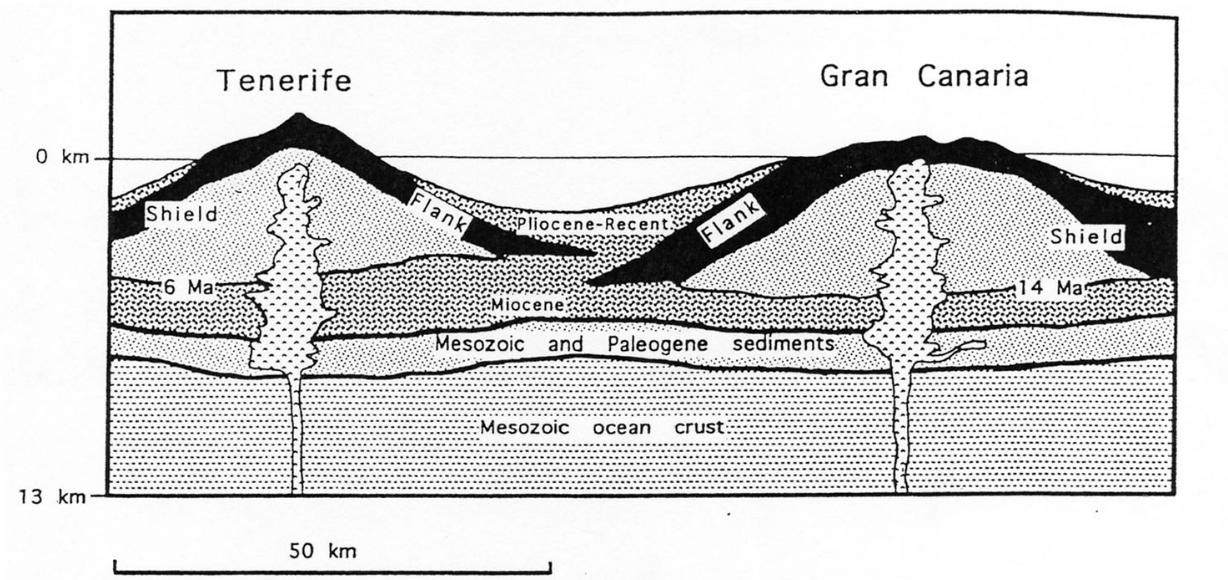


Figure 40 Cross-section redrawn from figure V-6 in Zoback & Emmerman (1994) illustrating that some authors interpret the volcanic piles of the hot-spot volcanoes of the oceanic part of the African Plate as neither a great load weighing down the underlying lithosphere like Hawaii (Figure 17) nor as sites of basement uplift like Fuenteventura and Maio (Figure 39). This figure illustrates a compromise in which the ocean floor is considered to remain roughly horizontal. Perhaps the two processes of uplift and depression balance each other. Unfortunately there are few deep seismic reflection lines across the oceanic hot-spot islands of the African Plate. What information there is (Figure 16) is suggestive of basement uplift under the oceanic hot spots just as there is basement uplift under the hot spots on the continent (Figure 14).

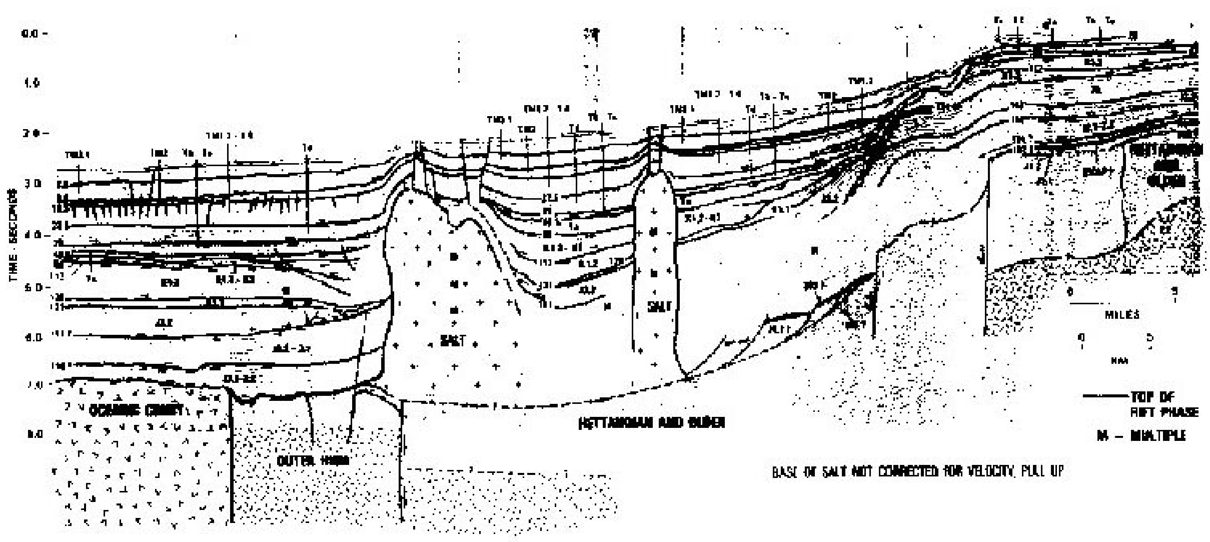


Figure 41 Seismic reflection profile probably from off the coast of Morocco reproduced from Schuepbach & Vail (1980). Although diapirs of Jurassic salt rising from the base of the section dominate the profile, the mid-Oligocene unconformity is also prominent. The mid-Oligocene unconformity is labelled on the left of the figure just below the 4.0 second mark by the number 29 which is its suggested age in millions of years. On the right side of the figure the unconformity is represented by a submarine canyon, just above the number 131, which has cut deeply into the pre-Oligocene section. In the middle of the figure the mid-Oligocene unconformity has cut out sediments with ages between 29 and 49.5 My. On the left of the figure more than a second of sediment can be seen to have accumulated in deep water above the mid-Oligocene unconformity during the past 29 My. This profile is typical of many which show the remarkable mid-Oligocene or ~30 Ma unconformity around the shores of Africa.

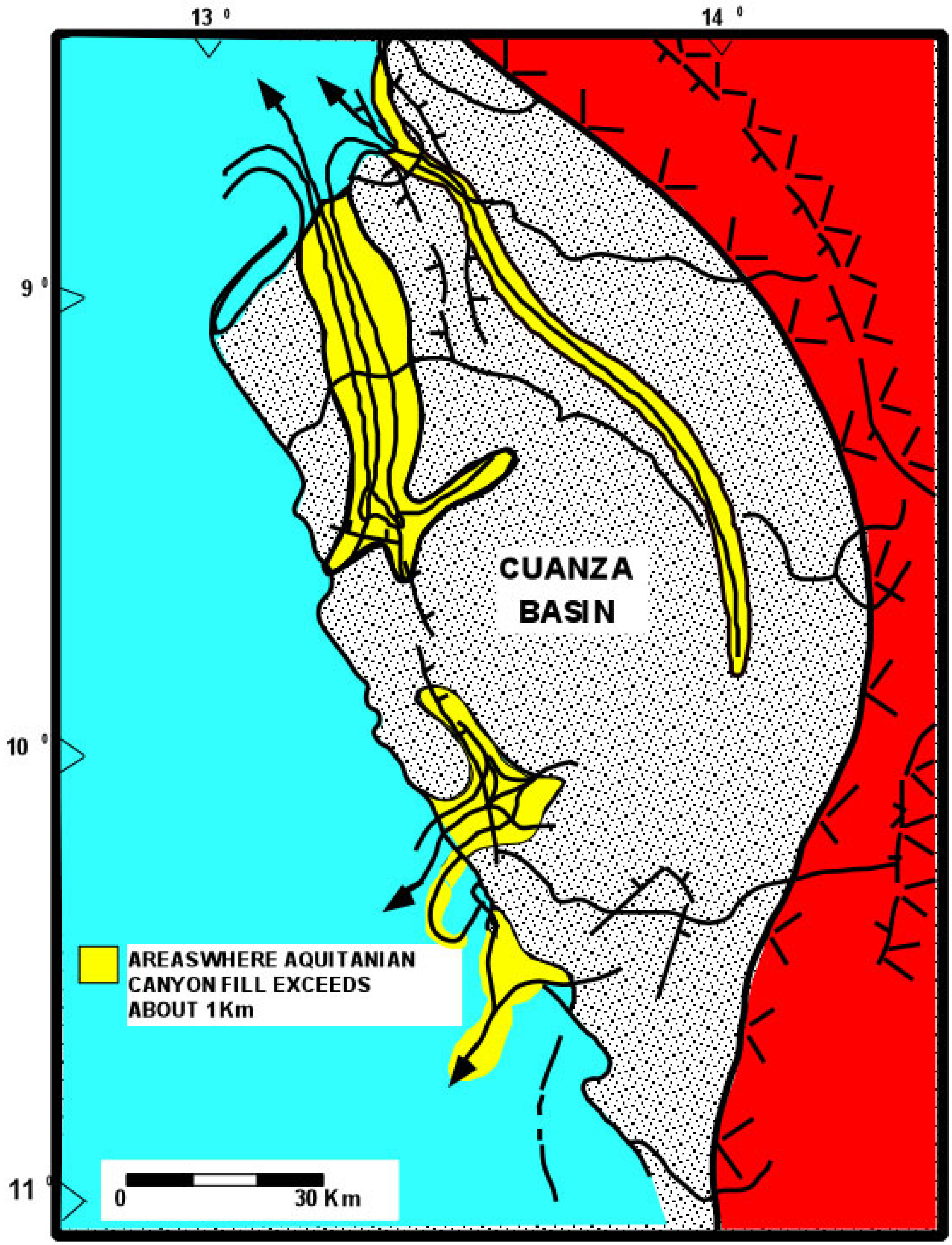


Figure 42 Map showing the distribution and depth of submarine canyons cut at the time of the 30 Ma unconformity in the Cuanza Basin. I prepared this map using data of Brognon & Verrier (1966) who did not interpret their information as indicating the presence of large submarine canyons. Salt tectonics at ~30 Ma, locally called *raft tectonics*, was a response to salt movement into the newly cut submarine canyons. This process has helped to widen the canyons (Duval *et al.*, 1992; Lundin, 1992).

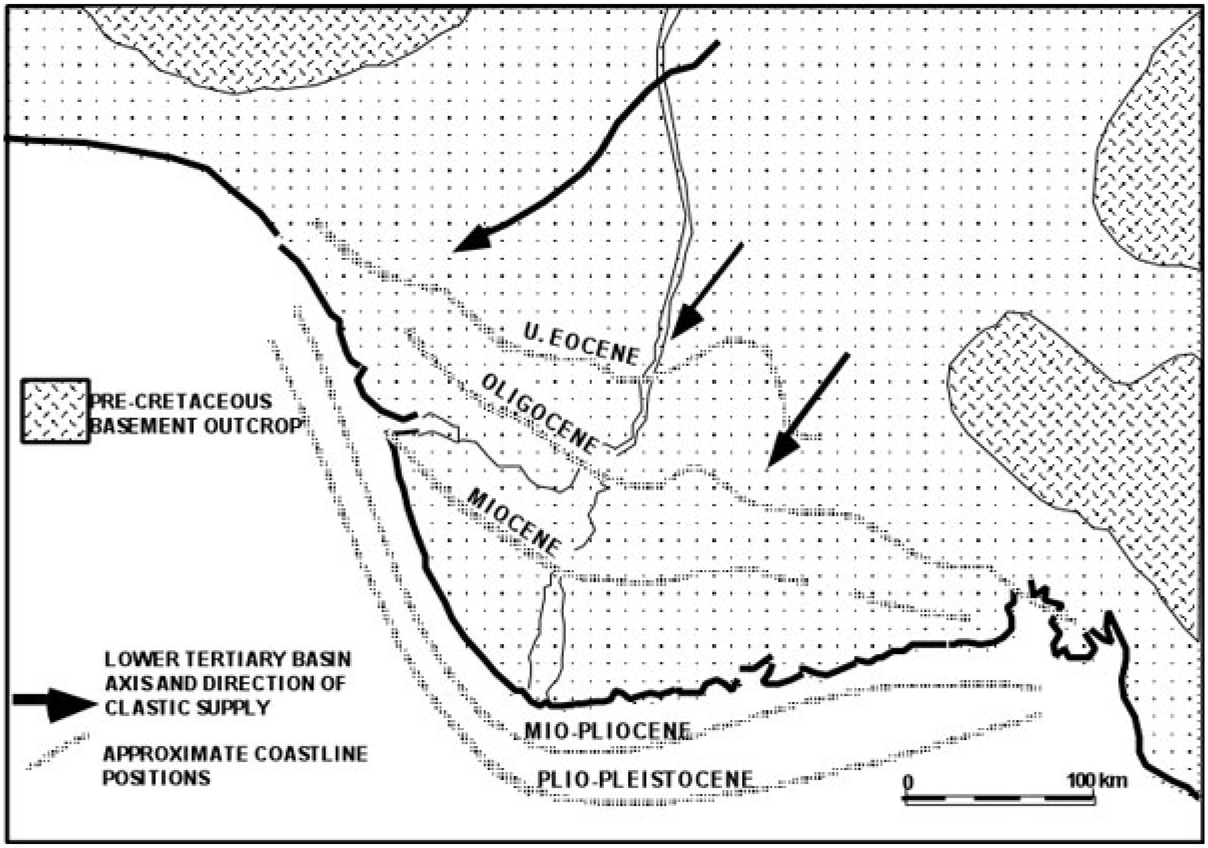


Figure 43. Map showing how the coastline of the Niger delta has prograded since 35 Ma. The delta has advanced seaward about 200 km and has broadened from a width of less than 300 km to a width of about 500 km. The great advance is attributable to an access of sediment produced by erosion of newly elevated areas of the continent. Figure simplified from Whiteman's (1982) figure 229 which embodies published work of Shell Nigeria geologists.

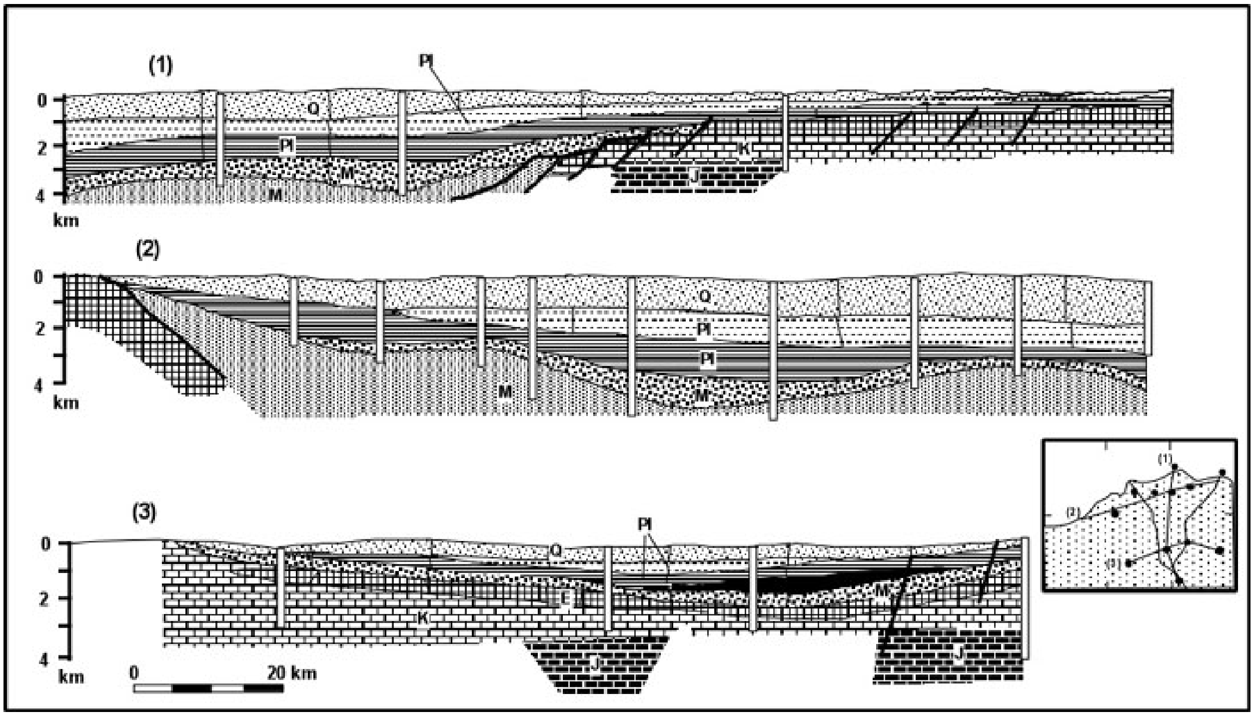


Figure 44 Three cross-sections through oil wells drilled in the Nile delta. The sections show that the Nile delta contains rocks as old as Miocene (M) in age and that the delta has been cut down into underlying carbonate sediments of Eocene (E), Cretaceous (K) and Jurassic (J) age. The delta consists of Miocene (M), Pliocene (Pl) and Quaternary (Q) sediments which together (see section 2) reach a total thickness of more than 4 km. The black layer shown in section 3 is basalt that was erupted about 25 Ma. The record indicates that the Nile or its ancestor rivers first began to flow by Miocene times. Figure based on figure 1.15 of Said (1993).

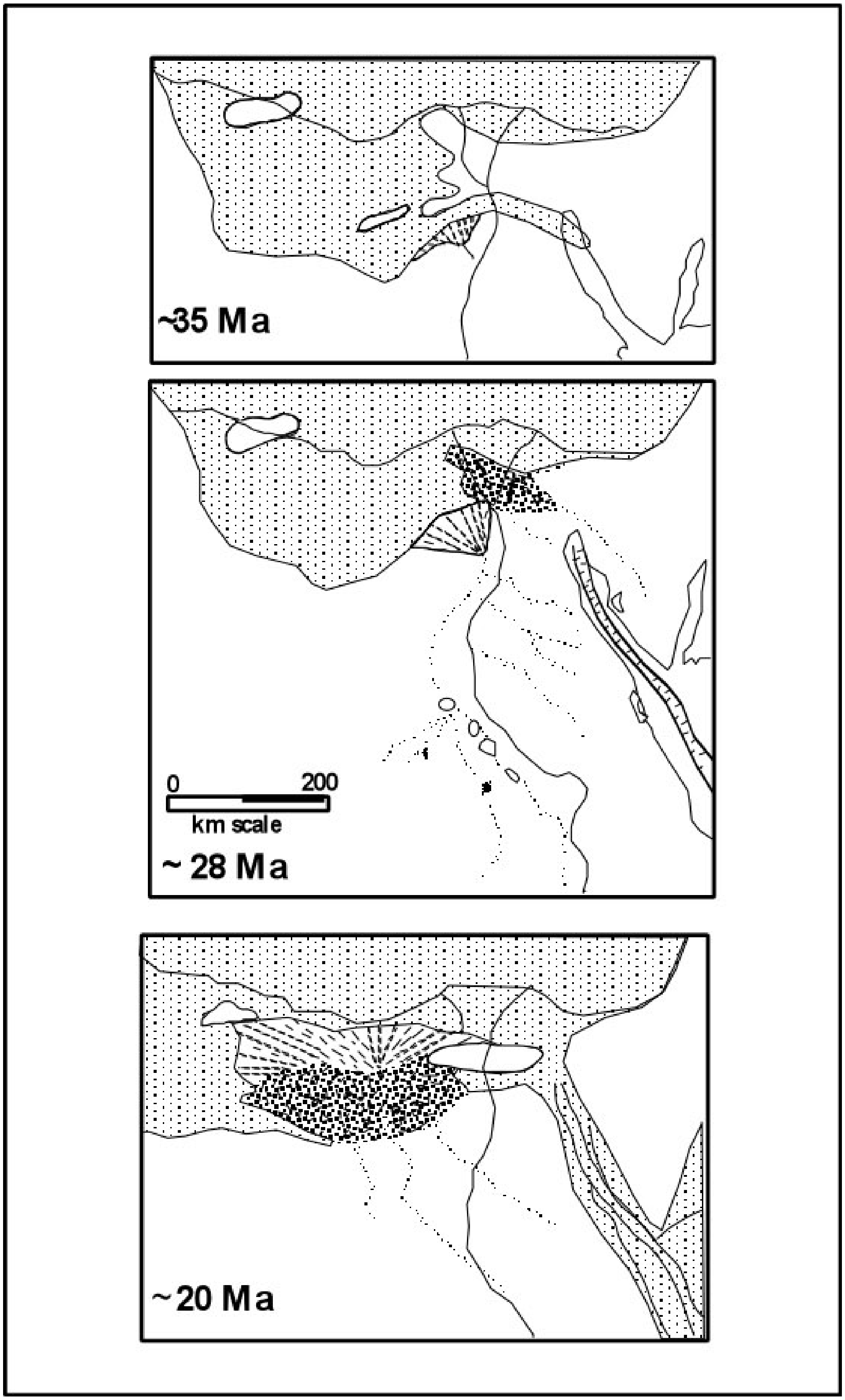


Figure 45 Sketch maps simplified from Said (1993, figure 1.16) showing that there was hardly a Nile delta at ~35 Ma (top). A delta with associated fluvial deposits, indicated by a pebble ornament, which was nearly 100 km wide had formed by ~28 Ma (center). By ~20 Ma the Nile delta was fully formed and was centered perhaps 150 km southwest of the position to which it has now prograded (bottom).

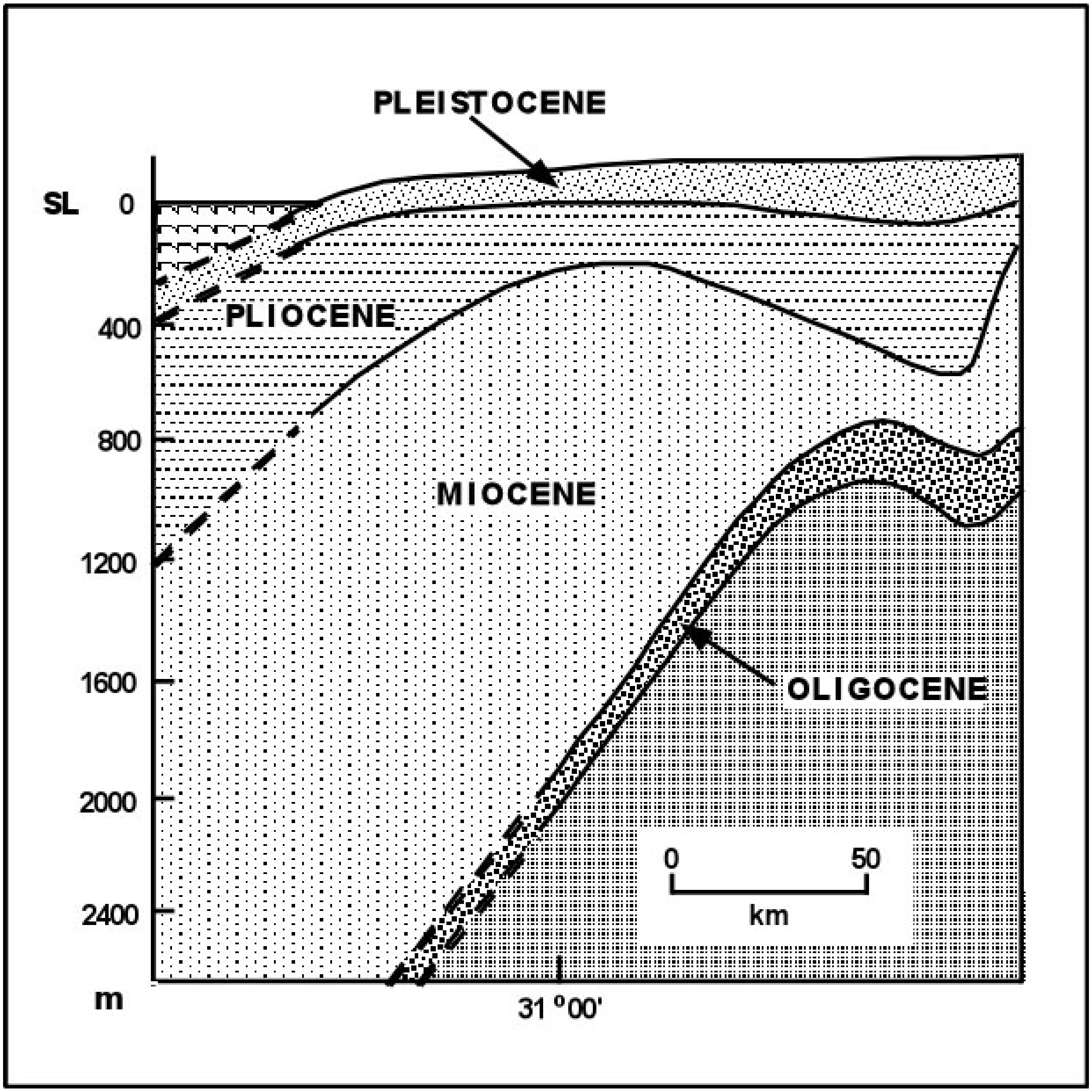


Figure 46 A cross-section based on well data showing how the Nile delta has prograded nearly 200 km since the beginning of the Miocene at about 22 Ma. Figure based on Elzarka & Radwan (1986).

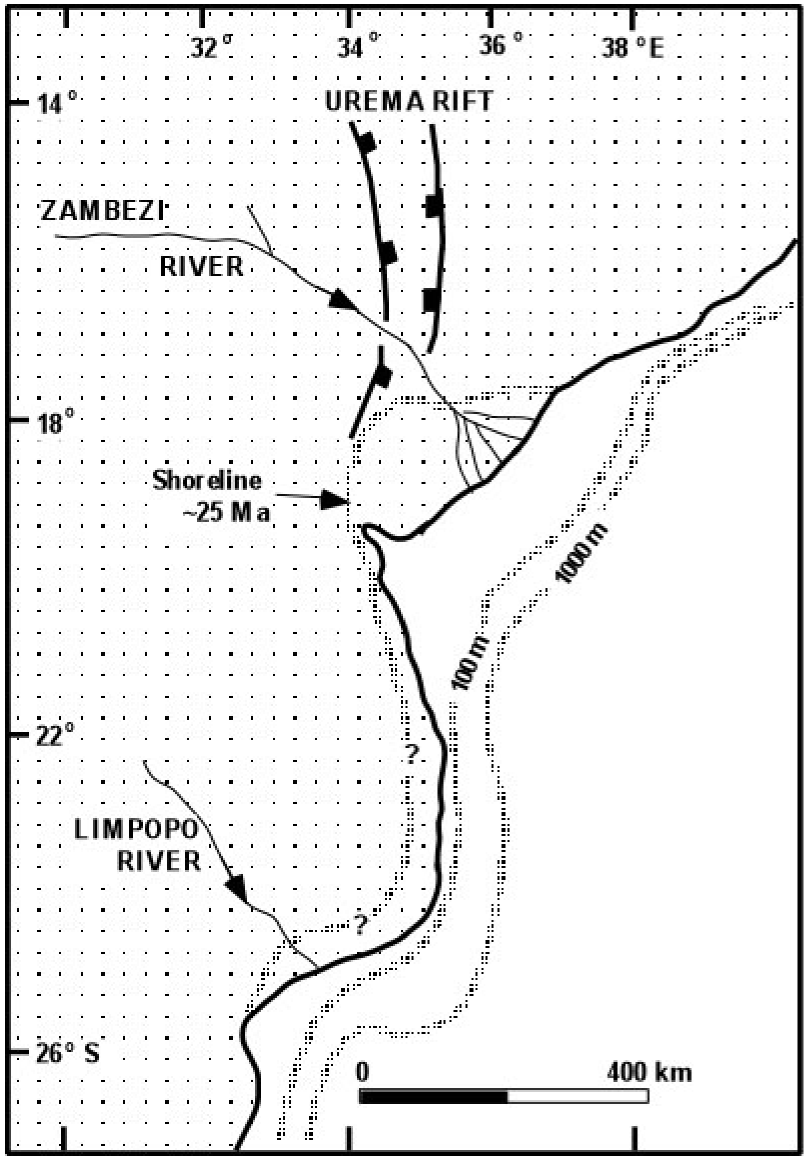


Figure 47 Sketch map based on figures in De Buyl & Flores (1986) and Droz & Mougenot (1987) showing how the Zambesi delta has prograded more than 200 km since 25 Ma. The progradation of the Limpopo delta has been much smaller. The position of the Urema graben, which extends south from the Shire valley in Malawi, is indicated. This structure was mainly active during the Cretaceous, but there are indications of activity on some of its faults within the past 30 My

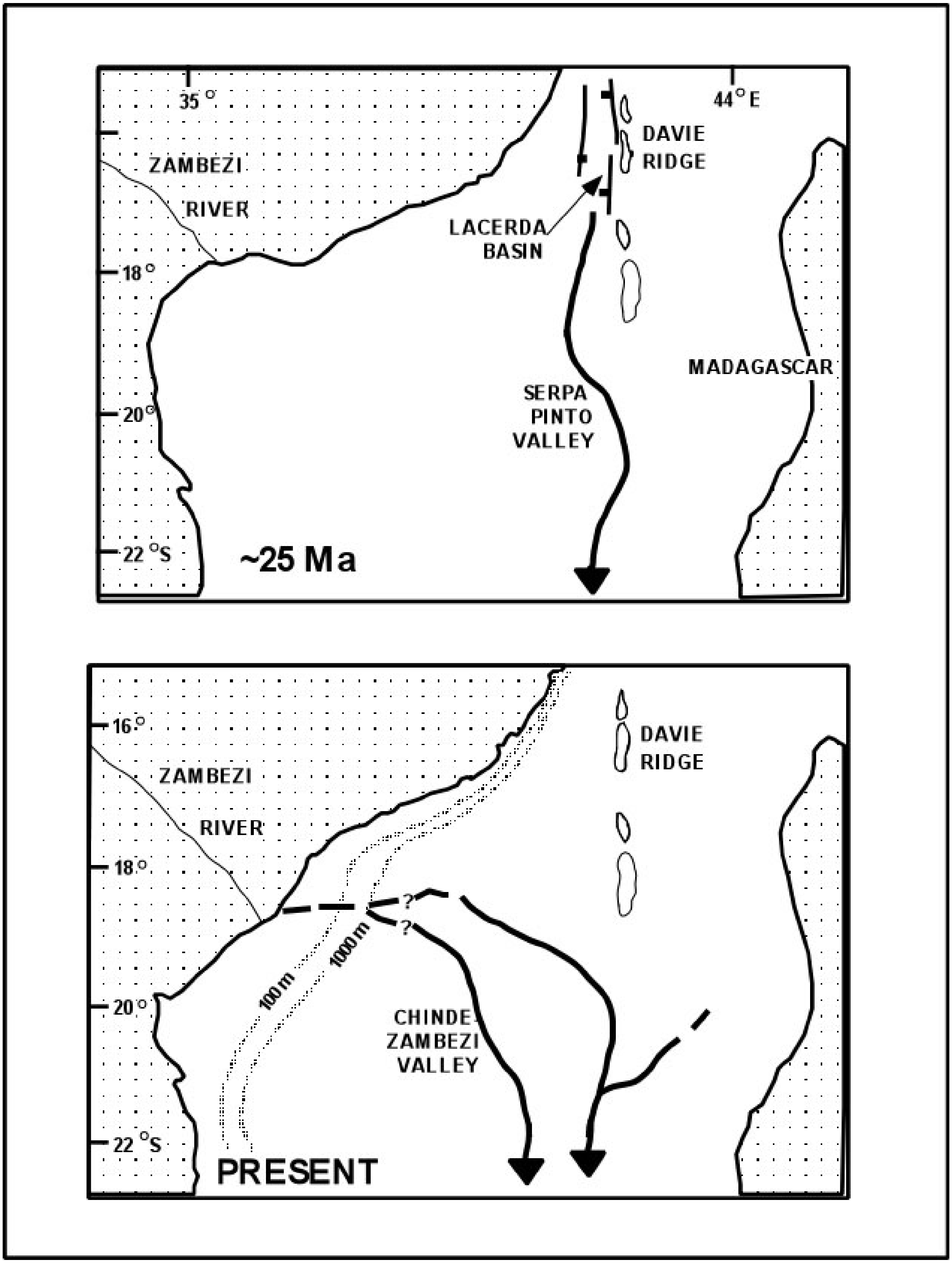


Figure 48 Deep sea channels on the upper part of the Mozambique deep sea fan were fed from the Lacerda Basin Rift ~25 Ma (top figure), but in Quaternary times, channels from the Zambesi delta had become dominant as that great delta prograded. Based on figure 13 in Droz & Mougenot (1987).

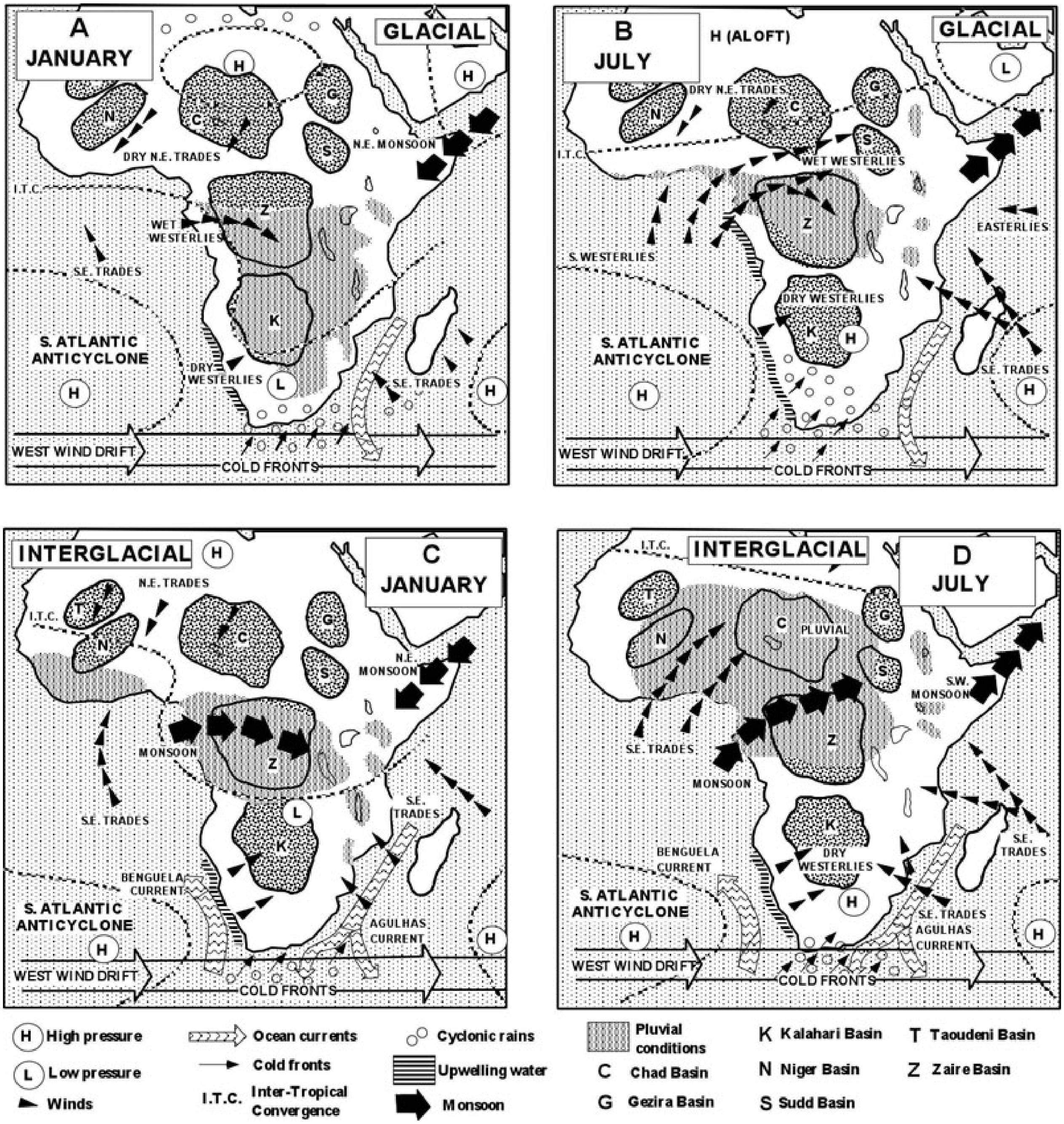


Figure 49 Figures indicating climatic variations in Africa related to changes in northern hemisphere glacial conditions (from Kroepelin, 1994, based on a figure by Shaw, 1985) incorporating the ideas of Van Zinderen Bakker (1976). Seven interior basins are indicated with checkerboard ornaments. Climatic variations have different effects in the various basins. For example, most of the Zaire Basin is subject to fluvial conditions, however the climate varies. Before northern hemisphere glaciation began at about 3 Ma, conditions similar to the interglacial conditions depicted in the lower two figures are likely to have obtained over the entire continent from about 35 Ma when the East Antarctic ice-sheet had formed and the Drake passage had opened.

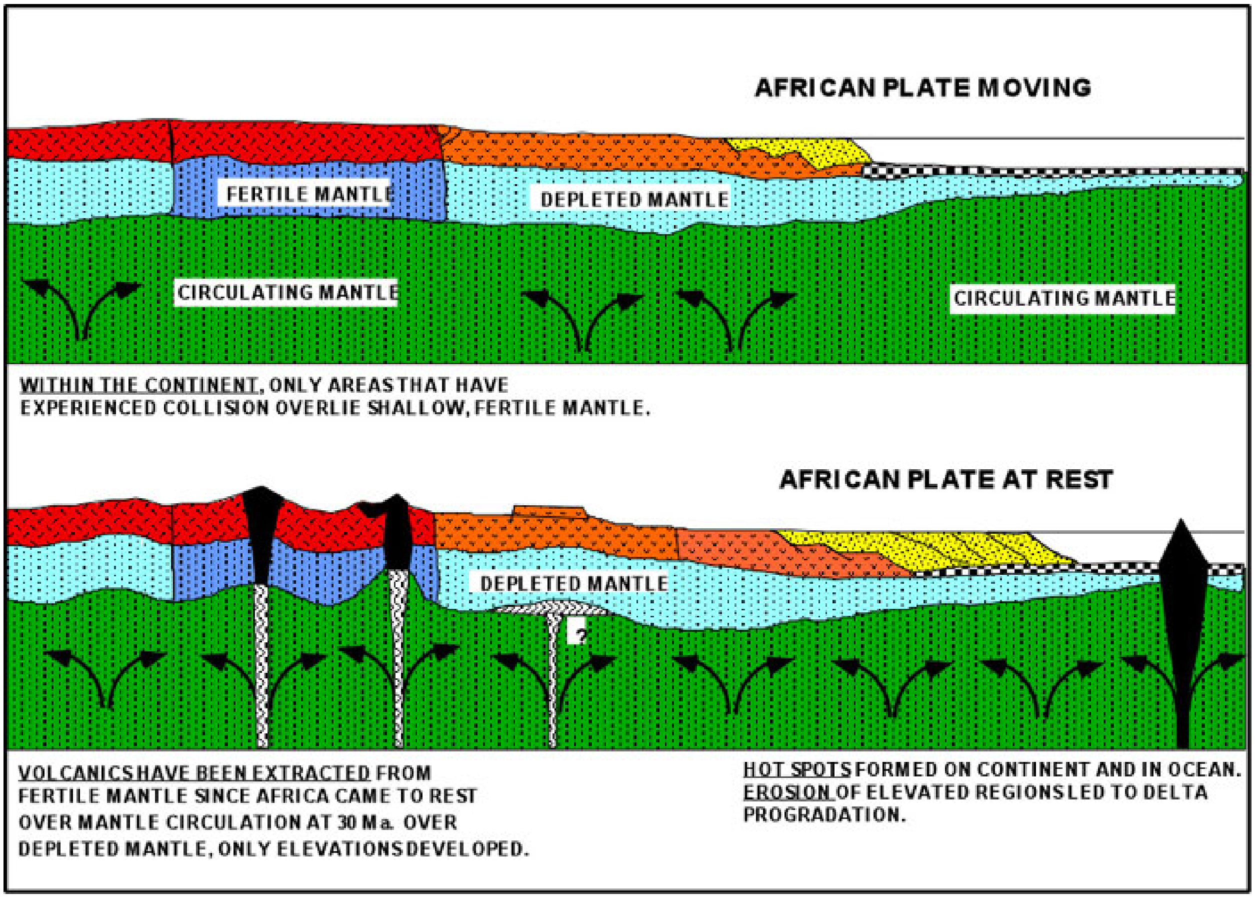


Figure 50 Sketches illustrating the difference between the state of the African Plate when it was moving over the mantle circulation and the state now that the plate is at rest with respect to the mantle circulation or at least that part of the mantle circulation represented by the plume population. The change has been from a low-lying continent with little volcanism, deep weathering and a humid climate to a continent of basins and swells with much active volcanism. That volcanism has persisted in the same places since the plate came to rest. Swells over cratonic areas underlain by depleted mantle lithosphere are not capped by volcanic rocks. Escarpments, including the Great Escarpment, have cut back into the swells and erosion from the swells has led to deposition of sediments in rapidly prograding deltas.

Table 1 Magmatic activation in the eastern part of the Central Atlantic (from Mazarovich, 1990)

Age	Millions of Years					
	32.3	25	16.8	11.8	5.4	3.4
	Pg^{1+2}	Pg^{3}	N_1^{2}	N_1^{2}	N_1^{3}	N^{1}
Fuenteventura	----	----	----	----	----	----
Lanzarote			----	----	----	----
Tenerife				----	----	----
Gran Canaria				----	----	----
Gomera				----	----	----
Selvagen Grande			----	----	----	----
Madeira				----	----	----
Porto Santo				----	----	
Maio			----	----	----	----
Josephine seamount				----	----	
Ormonde Bank			----	----	----	
Great Meteor seamount			----	----		
Er reef				----	----	
Dakar			----	----	----	----
Mazagan plateau		----	----	----	----	
Cape Bojador		----	----	----		