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Elevation as indicator of mantle-plume activity

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ABSTRACT

My contention—not universally agreed upon—in this paper is that hotspots are related to plumes, which are mantle phenomena and are known by their manifestations at Earth's surface. The most unequivocal of these manifestations is uplift. There is no other process on this planet that creates domes of lithospheric flexure (i.e., falcogenic [large-wavelength] domes) of ~1000 km diameter and 1 to 2 km amplitude within several million years. Such domes rise rapidly following arrival of a plume under the lithosphere. Uplift probably occurs by detaching pieces of the lithosphere and sinking them into the plume head. Domes do not generate enough extension to form rifts, but they do generate sufficient gravitational potential to lead to rifting. Giant radiating-dike swarms probably form as a result of spatial and temporal proximity of a plume with a pole of extensional rotation. The most active region of plume activity today is in Africa. In the Afar region, a dome of almost 1000 km radius began rising after the early Eocene, and it had probably reached an elevation of more than 1 km by early Oligocene time. Both basalt extrusion and rifting began afterward. A similar sequence of events occurred at the Kenya dome. Rifting and volcanicity at the Kenya dome progressed from north to south, giving the impression of a progressive tearing of Africa from the Afar southward. The uplift argument shows that the widespread late Devonian rifting and volcanism in eastern Europe were not plume-related phenomena, whereas the disruption of Gondwanaland in the Mesozoic was.

Dedicated to the memory of Ziyad R. Beydoun, without whose life work this paper could not have been written.

"The study of the earth's distortion is important for its own sake; it is important for its bearing on other major mysteries about the globe."
Reginald Aldworth Daly, 1926

PROLOGUE

Plumes are mantle phenomena and are known by their manifestations at Earth's surface. As stated in the abstract, my contention in this paper is that hotspots are related to plumes. Kevin Burke (2000, written commun.) objected to this statement and asserted that a hotspot is simply a locus of mid-plate magmatism that may or may not be plume related. My contention in this paper is that there is no source of significant mid-plate magmatism unrelated to plate-boundary phenomena and possibly to glacio- and hydroisostasy other than plumes (see also the discussion in Şengör and Natal'in, this volume).

INTRODUCTION

The purpose of this paper is to suggest *one* criterion that may be uniquely reliable for the identification of former mantle plumes. The procedure for developing such a criterion is to examine what are considered modern plumes, defined on the basis of diverse, but mainly topographic (or bathymetric, or stratigraphic in places where erosion rapidly destroys elevations) and

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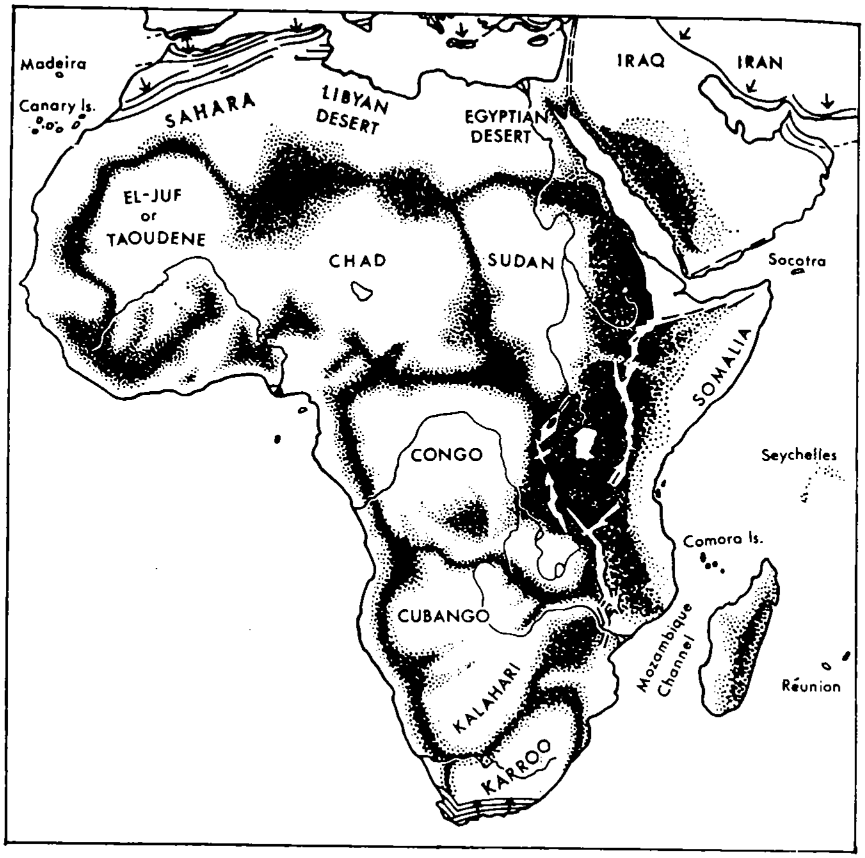

Figure 1. The plateaus and basins of Africa as depicted in Holmes's famous figure (Holmes, 1944, Fig. 223; reproduced in Holmes, 1965, Fig. 763). Though depicting a structural pattern well known since the last quarter of the nineteenth century, this figure inspired many, including Kevin Burke, to think about the origin of it within the context of the new ideas developed together with plate tectonics. Note the round and wide basins and narrower uplifts.

magmatic criteria, plus an array of developing geochemical and geophysical signatures (e.g., Wilson, 1963; Morgan, 1971, 1972a, 1972b; Burke and Wilson, 1972; Burke and Whiteman, 1973; Burke and Dewey, 1973; Courtney and White, 1986; Detrick et al., 1989; White and McKenzie, 1989; Sleep, 1990; Liu and Chase, 1991; Loper, 1991; Watson and McKenzie, 1991; Wessel, 1991; O'Connor and le Roex, 1992; Marzocchi and Mulargia, 1993; Wessel, 1993; Lee et al., 1994; Granet et al., 1995; Kingsley and Schilling, 1995; McKenzie and O'Nions, 1995; Burke, 1996; Ito et al., 1996; LeMasurier and Landis, 1996; Stewart and Rogers, 1996; Widom and Shirey, 1996; Hanan and Schilling, 1997; Dickinson, 1998; Ji and Nataf, 1998; Wessel and Kroenke, 1998; Orovetskii, 1999; Cserepes and Yuen, 2000), notwithstanding the ongoing controversy over the very existence of mantle plumes (e.g., Anderson, 1994, 1995, 1996, 1998; W.B. Hamilton, 1998, personal communication; Sheth, 1999; Smith and Lewis, 1999). Out of these criteria, one must isolate a set of easily recognizable characteristics, traces of which may be preserved in the geologic record, that would thus aid in identifying the location of past plumes. The goal of this paper is to highlight the *preservable* fingerprints of what the majority of earth scientists today think of as mantle plumes. The selection and application to the past of my *one* criterion, among a number of possible others, constitute my principal task. The selection was easy: I resolved to find a criterion, the components of which would be *directly observable.* This requirement was to avoid a *double reconstruction* of the following sort: first, the reconstruction of an event not directly observable today (e.g., magma generation that takes place away from geologists' gaze in the depths of the Earth) and then the reconstruction of its supposed analogue in the past. Instead, I wanted a process I could observe now without having to reconstruct it to understand it. I could then attempt to find its analogues in the past by appropriate single-step reconstruction. The bulk of this paper is devoted to the illustration of how my criterion works.

It was Jason Morgan (1971, 1972a, 1972b) mainly in the oceans and Kevin Burke (Burke and Wilson, 1972; Burke and Whiteman, 1973; Burke and Dewey, 1973, Burke et al., 1973) mainly on the continents who linked with Wilson's theory of mantle plumes (Wilson, 1963, 1965, 1973) the large, roughly elliptical elevations characteristic of Hawaii or the Neogene development of Africa (Fig. 1). Whereas Morgan was led to the postulate of oval uplifts over plumes of 1000 km diameter based on the "Hess gravity theorem" i.e., that one does not need to have a gravimeter to measure gravity, one needs only to look at the topography: Morgan, 1972a, p. 19),[1] Burke developed

[1] But Morgan was suspicious of continental topography. "Whether the unusually high Tibetan Plateau or southern Africa should be considered symptomatic of a subcontinental hot spot is open to question; the more uniform oceans are more amenable to this type of analysis" (Morgan, 1972a, p. 19). This statement

this view during the work he did with co-workers on the evolution of the Benue Trough and the Gulf of Guinea, (in the framework of the Benue Valley Project of the Geology Department of the University of Ibadan, Ibadan, Nigeria, that he had initiated in 1966). In a discussion remark made in December 1970 during the Conference on African Geology in Ibadan, Kevin Burke first mentioned that uplifts are associated with plumes and form one stage in a sequence of events eventually leading to continental disruption and ocean opening: "Rift valleys are not something unique and unrelated to ocean opening. Some rifts like the Cameroon and the East African are at too early a stage to show any evidence of opening and the only structures they reveal are related to uplift. Others ceased to be active before their development had gone very far, for example, the Mesozoic rift of southern east Africa. The Cretaceous Benue rift was one of the latter. It developed farther than its east African contemporary but still closed at an early stage" (published in Burke et al., 1972, p. 202).

It was on the basis of the famous study of Hans Cloos (1939) that Burke and Whiteman (1973) brought into genetic relationship big (~100–2000 km diameter and ~1 km uplift), oval *falcogenic*[2] structures and the generation of commonly three-armed rift stars. This model was extremely attractive, as the structures could be made to develop into a rift-rift-rift triple junction of the kind analyzed by McKenzie and Morgan (1969) or, alternatively, into a single spreading ridge and an abandoned rift (an aulacogen) on one of the separating continents (cf. Burke and Dewey, 1973; Şengör, 1987, 1995). The first published map of plume-related African swells and basins enclosed by them appeared in Burke and Wilson (1972) and a revised version in Burke and Whiteman (1973; see Fig. 2, A and B).

expresses a fundamental truth concerning the differences between the oceans and the continents, yet it betrays also a fairly naive view of continental topography. Kevin Burke further remarked (2000, written communication) that "There is a more serious problem with submarine elevations. Erosion does not reduce elevation below sea level so it is hard to distinguish dynamically maintained from ancient elevations, especially where carbonate banks overlie volcanic material."

[2] In a forthcoming work, I have defined as convenient short terms *falcogenic structures* and *copeogenic structures.* The term *falcogenic* was taken from the classical Greek φάλκης (= the bent rib of a ship) to indicate the large-wavelength (hundreds to thousands of kilometers), small-amplitude (a couple of kilometers at most) structures that form almost exclusively by *bending* the lithosphere and not by major faulting. I contrast falcogenic structures with copeogenic structures (from the Greek κοπή = to cut up, slaughter) that have short wavelengths (millimeter to 10 km scale) and amplitudes commonly larger than their wavelengths. Falcogenic structures include epeirogenic structures in both Stille's (1919, 1924) sense and that of koilogenic (i.e. broad and commonly faultless basin-making) and cymatogenic (i.e., arching and doming) structures, and the copeogenic structures include orogenic, taphrogenic, and keirogenic (long and broad belt of strike-slip) structures (see Şengör, 1995). A need for such a term was felt as early as 1911 (cf. Stille, 1911, footnote 2 on p. 279–280), yet geologists preferred to use an old term and change its meaning at will, instead of creating a new term. This has caused endless trouble. I know that introduction of a new term is generally unwelcome, but I think the terms *falcogenic* and *copeogenic* are justified by the need to be able to talk about large-wavelength, commonly flexural structures versus short-wavelength structures with diverse types of folding and faulting without actually specifying their particular deformations.

Though the Cloos-Burke[3] model looks attractive, the effect is spoiled by the fact that Cloos's experiments led to development of rift-like features in clay cakes only when unrealistic amounts of uplift—10 times that observed in nature—were assumed. (Cloos mistakenly assumed that a significant fraction of the postrifting shoulder uplift due to isostatic rebound was part of the prerifting dome. This assumption grossly amplified his uplift estimates). Those experiments generated only small fissures when the proportions of the experimental domes resembled those of natural lithospheric domes. Thus the basic premise of the Burke model—that "rifting could result from the formation of a dome-shaped uplift" (Burke and Whiteman, 1973, p. 738)—*appears* to be left without observational support, as has long been widely appreciated, especially among geophysicists. For a textbook statement, see Bott (1982b, p. 68). In the next section, I show in some detail that the observed doming at selected sites in Africa and Australia today (i.e., at continental sites away from active plate boundaries) is insufficient to generate enough stretching to lead to the observed styles of rifting and that it is unlikely that any doming under terrestrial conditions can accomplish this result. In the subsequent section, I examine the history of the East African rift system around the Afar and Kenyan domes and show that doming indeed preceded rifting in both as in the original Cloos-Burke model, begging the question of mechanism. In the final section, I show that even the most superficial gravitational stresses may be big enough to cause rifting but not quite big enough to lead to peripheral thrusting around the uplift. I conclude that plumes lead to rifting *always* by increasing the potential energy of a part of the lithosphere that consequently disintegrates along one or more rifts. Therefore, the presence of a prerifting domal uplift (regardless of its identification by topography or by structure and stratigraphy), commonly crowned by volcanic rocks, is a fairly safe indicator of the presence of a plume underneath.

UPLIFT

Figure 3 shows the locations of the selected hotspots used in this study. They were chosen on the basis of (1) widespread agreement that they are plume-related features and (2) their independence from plate-boundary-related phenomena. Some are far away from plate boundaries (such as Hawaii), others are closer (such as Crozet), but none coincides with a plate boundary.

Oceanic hotspot plateaus are broad regions of uniform elevation above the surrounding abyssal plains. Regardless of which plate the plateaus stand on, the age of oceanic basement

[3] Burke now has a new model for the Afar region (Burke, 1996, p. 377–378, and Fig. 33; and written communication, 2000) to which the model of Manighetti et al. (1997) is not dissimilar. It is unclear to me how it can substitute for the Cloos model as a *general* mechanism because its only difference seems to be the addition of a compressional-stress component within the lithosphere and that it is dependent on the geographic peculiarities of northeast Africa in the Neogene. Geologists still need a way to make the extension significant enough to generate rifts. Burke's new model, like that of Cloos, fails to provide it.

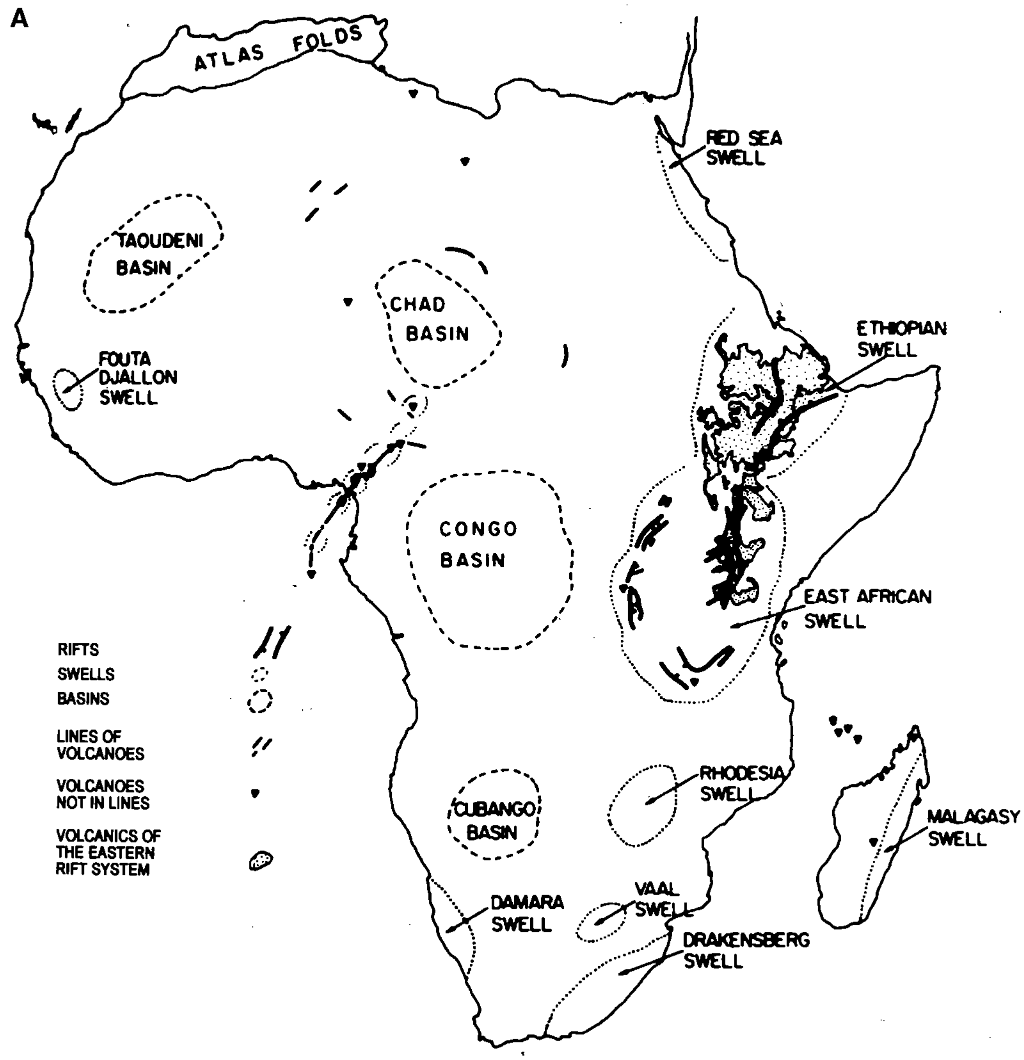

Figure 2. A, Uplifts and basins of Africa today as conceived by Burke and Wilson (1972, Fig. 2). Note the wide basins (outlined in dashes) and the narrow uplifts (Holmes's influence?). B, Uplifts (black) around the Chad basin according to Burke and Whiteman (1973, Fig. 1). Note that uplifts have been depicted to be nowhere wider than 500 km.

underlying them, or their age of initiation, they tend to stand at a uniform depth of about −4000 m (Fig. 4). Only two out of the population considered here stand at −4500 m. The anomalously shallow Kerguelen Plateau stands at −2500 m, but that elevation is also due to its thick crust (15 to 23 km: Duncan and Storey, 1992, p. 96, and the references therein), not just to thermal doming (cf. Coffin, 1992). Five of the 15 plumes shown in Figure 4 and defined in Table 1 may have ceased activity by the end of the Pleistocene. However, the Holocene constitutes insufficient time to make a significant topographic difference (at most 100 m or so by thermal augmentation and resultant subsidence of the lithosphere: Parsons and Richter, 1981, Fig. 4 therein; at most 130 m by erosion [glacial erosion excepted]: Ring et al., 1999, especially Fig. 6 therein). By contrast, the maximum elevations of the oceanic hotspots themselves scatter between −48 m and + 4169 m, showing no systematic relationship to any other parameter, except that young hotspots tend to have lower maximum elevation values (Fig. 4).

Table 2 lists the continental hotspots considered in this paper, and Figure 5 shows both their plateau heights and maximum elevations. The average continental plateau heights cluster at ~1000 m and show a greater scatter than oceanic hotspots. Maximum elevations range from 800 m to 4070 m. Figure 6 is a plot of the maximum elevations attained by *both* the oceanic *and* the continental hotspots, and the similarity of their scatter is remarkable. Though plateau height is controlled by isostasy, maximum height (i.e., the elevations the individual volcanoes crowning the plateau reach) is controlled by magma-chamber size and dynamics, as well as volume and viscosity of volcanic production (cf. Macdonald, 1972, especially p. 255–287).

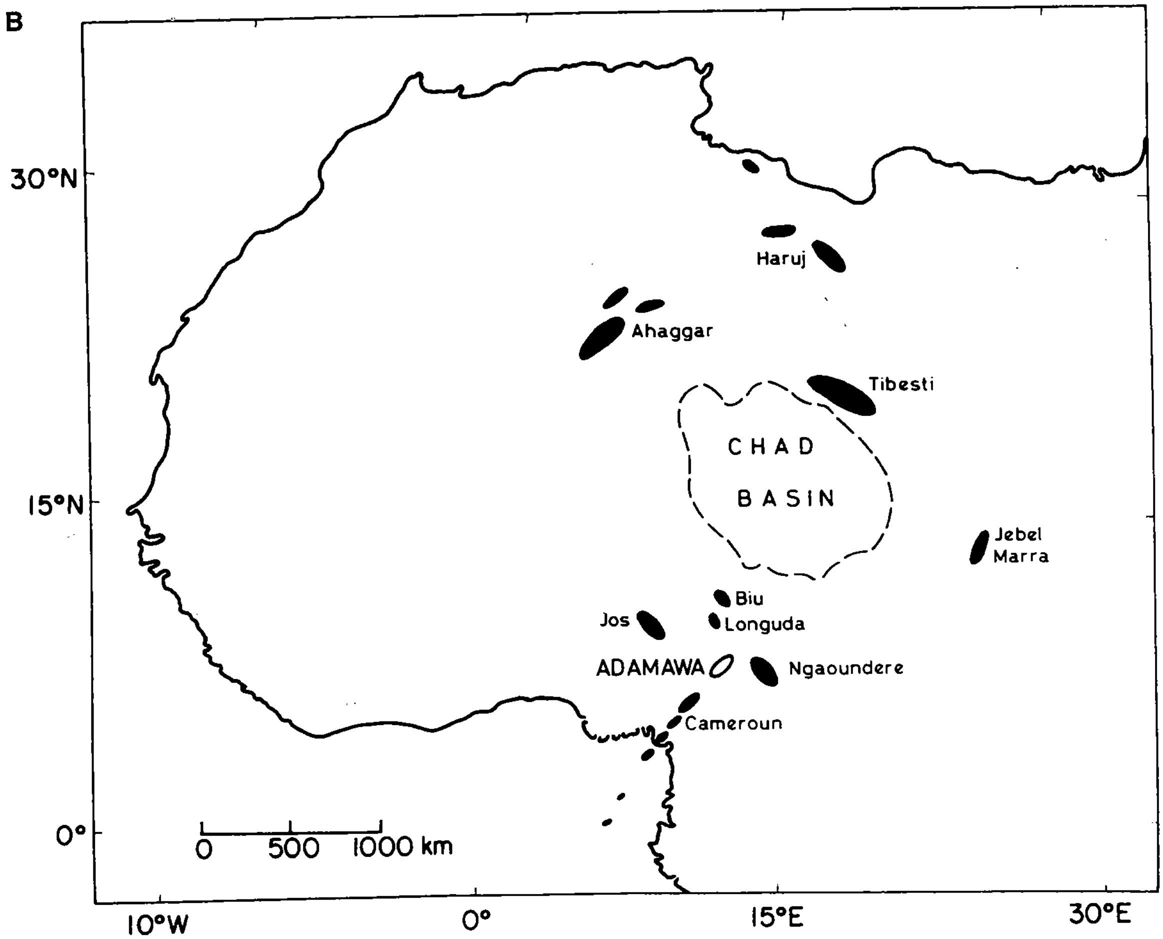

Figure 2. (continued).

Plateau heights of hotspots have been explained by two different mechanisms. Courtney and White (1986), White and McKenzie (1989), and Watson and McKenzie (1991) ascribed hotspot plateaus to dynamic uplift propelled by plume-related upwelling in the mantle, whereas Detrick and Crough (1978), Crough (1979, 1983a, 1983b), Yuen and Fleitout (1985), Liu and Chase (1989), and Dalloubeix and Fleitout (1989) held that thermal erosion of the underlying mantle lithosphere was responsible for the rise of the hotspot plateaus (Mohr, 1981, but see Von Herzen et al., 1989).

Figure 7 shows, on the basis of Lachenbruch and Morgan (1990, Fig. 3), the effects of removal of the mantle lithosphere from beneath continental crust with a constant thickness of 40 km and average density of 2.85 g/cm^3 (Christensen and Mooney, 1995; M.L. Zoback, 2000, personal commun.) and from beneath oceanic crust with a constant thickness of 8 km and average density of 2.9 g/cm^3 (Lachenbruch and Morgan, 1990). Under these conditions, the continental average of ~1000 m plateau elevation requires a mantle lithosphere—normally 130 km thick with average density of 3.25 g/cm^3—to be thinned to 65 km. If a plateau with a 40-km-thick continental crust is to rise to 1300 m, the region must reduce its mantle lithosphere to 50 km, as also suggested by gravity observations (e.g., Fairhead, 1976, 1986).[4] If a continental crustal column is 40 km thick with no mantle lithospheric root attached, it will stand 2000 m above the continental plains that are near sea level. For instance, the North Baykal Uplands in eastern Siberia

[4] I must emphasize that Nyblade and Pollack's (1990) objection to the Fairhead gravity model on account of a supposed Pan-African remanent crustal root is void, because lithospheric delamination (see Ashwal and Burke, 1989, especially Fig. 2) and reasonable rates of denudation (Ring et al., 1999), combined with the eustatic history of the Phanerozoic (Haq et al., 1988; Harrison, 1990; Hallam, 1992; Dewey and Pitman, 1998), make it exceedingly unlikely that any Pan-African (here taken as 600 Ma or older, at the youngest!) crustal root and its correlative topography could have survived the entire Phanerozoic history (i.e., since 540 Ma), as shown by Pitman and Golovchenko (1991). However, a deep lithospheric root has been inferred under the Limpopo belt by Hildebrand and Gurney (1995) on the basis of fast shear-wave velocities and the composition of kimberlitic minerals. What geologists know of the Cenozoic paleogeography of Africa corroborates this inference, despite the generalized mountains shown in Smith et al. (1994, maps 5–9; compare these with Summerfield's (1991) basin distribution, Fig. 16.25; see also Burke's (1996) statement on the flooding of Africa during the Cretaceous, p. 365). For igneous support of Fairhead's (1976, 1986) model, see Woolley (1991, especially Fig. 19) and Mohr (1992, especially Fig. 2B).

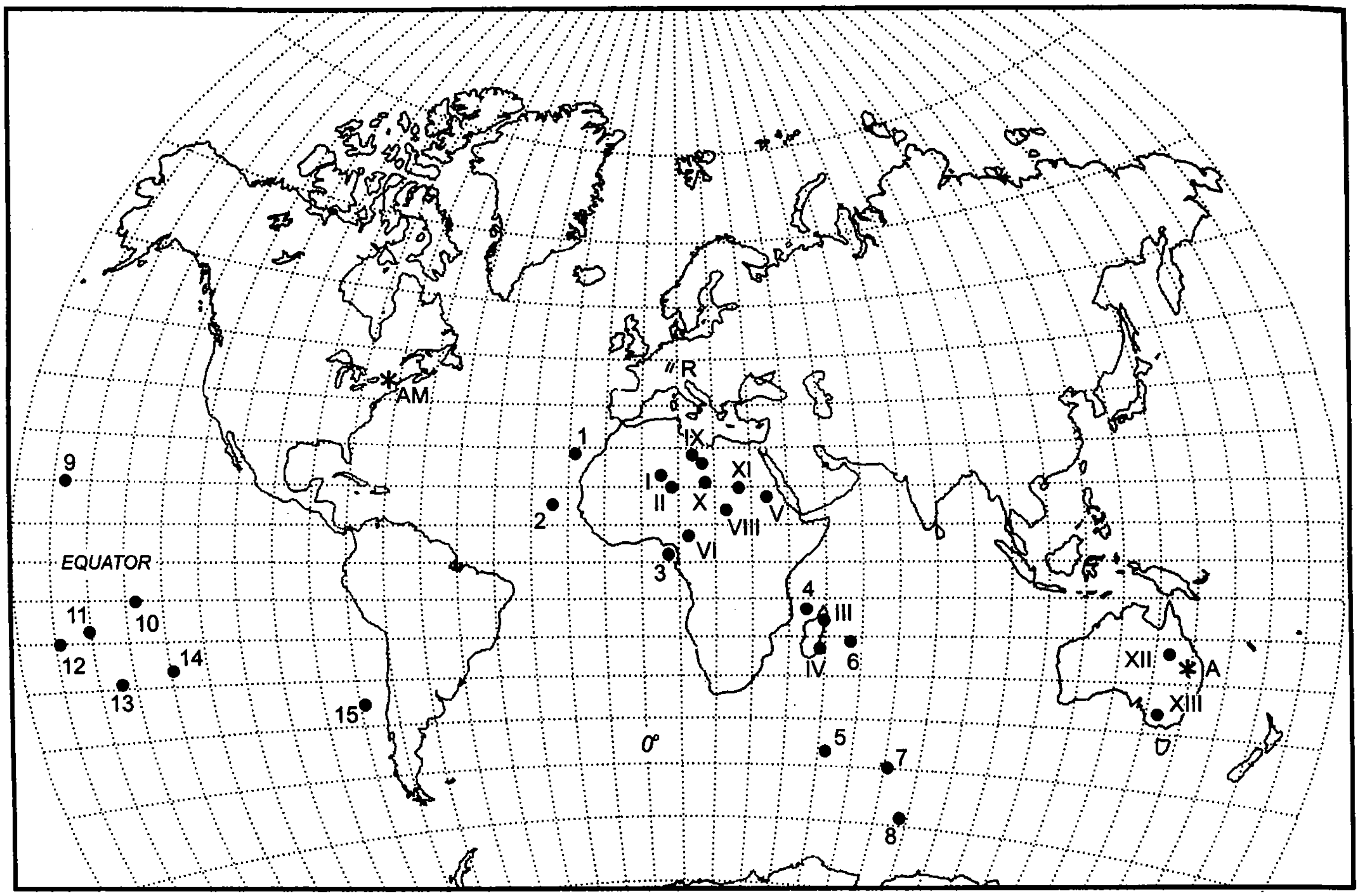

Figure 3. Hotspots used in this study. Base map modified from Wagner (1971). Arabic numerals refer to oceanic hotspots (listed in Table 1) and roman numerals to continental ones (listed in Table 2) A—Alice Tableland in Australia, AM—Adirondack Mountains in North America, R—Upper Rhine rift in Europe.

(*Severo Baykalskoie Nagorye*) stand at an average elevation of more than 2000 m. Here, Zorin et al. (1995) have established from seismic velocities that the crust is close to 50 km thick (thinning considerably in rift valleys that intervene between mountain ranges and plateaus) and that there is no mantle lithosphere. Where the mantle lithosphere thickens to ~160 km in the north, elevations drop to ~200–500 m (Zorin et al., 1995, Fig. 2 therein; also see Zorin et al., 1989; for the thickening of the continental lithosphere in time, see Kono and Amano, 1978).

In the oceans, an 8-km-thick oceanic crust (average density ~2.9 g/cm^3) requires a mantle lithosphere ~35 km thick to stand 1 km higher than adjacent abyssal plains at -5 km depth (Fig. 7 cf. Yoshii et al., 1976). Observations by Von Herzen et al. (1989) show, however, that the 20 mW·m^{-2} heat-flow anomaly expected of the presumably reduced lithospheric thickness is not observed in Hawaii, the Bermuda Rise, or the Cape Verde Islands (see also Courtney and White, 1986); therefore, Von Herzen et al. concluded that the simple lithospheric thinning model must be wrong (see also Stein and Stein, 1992). I disagree for the following four reasons:

(1) When plume-related rift chains lead to ocean opening (cf. Burke and Dewey, 1973; Burke, 1976; Şengör, 1995), the resulting continental margins generally subside at rates that follow the $t^{1/2}$ relationship (where t is time of onset of sea-floor spreading in the adjacent ocean; McKenzie, 1978; Pitman, 1978; Van Hinte, 1978), suggesting thermal augmentation of previously thin lithosphere below. Margins that grow out of strike-slip separation segments do not show this $t^{1/2}$ relationship (Scrutton, 1979).

(2) The fact that full cratons seem too strong for disruption by plate-boundary forces (Zoback and Zoback, 1997; Şengör, 1999) yet seem to yield to plumes suggests that plumes must thermally weaken a craton's lithospheric armor and provide a source of extensional stresses. The claim by King and Anderson (1995, p. 269) that "high-temperature magmas (picrites, komatiites) are found on the margins of cratonic lithosphere" only shows that in such places the cratons are commonly disrupted completely to allow these magmas to rise. The claim also ignores such spectacular examples of incomplete disruption as the late Archean Great Dyke (2.5. Ga) cutting the Archean craton in Zimbabwe in two (Wilson and Wilson, 1981), especially

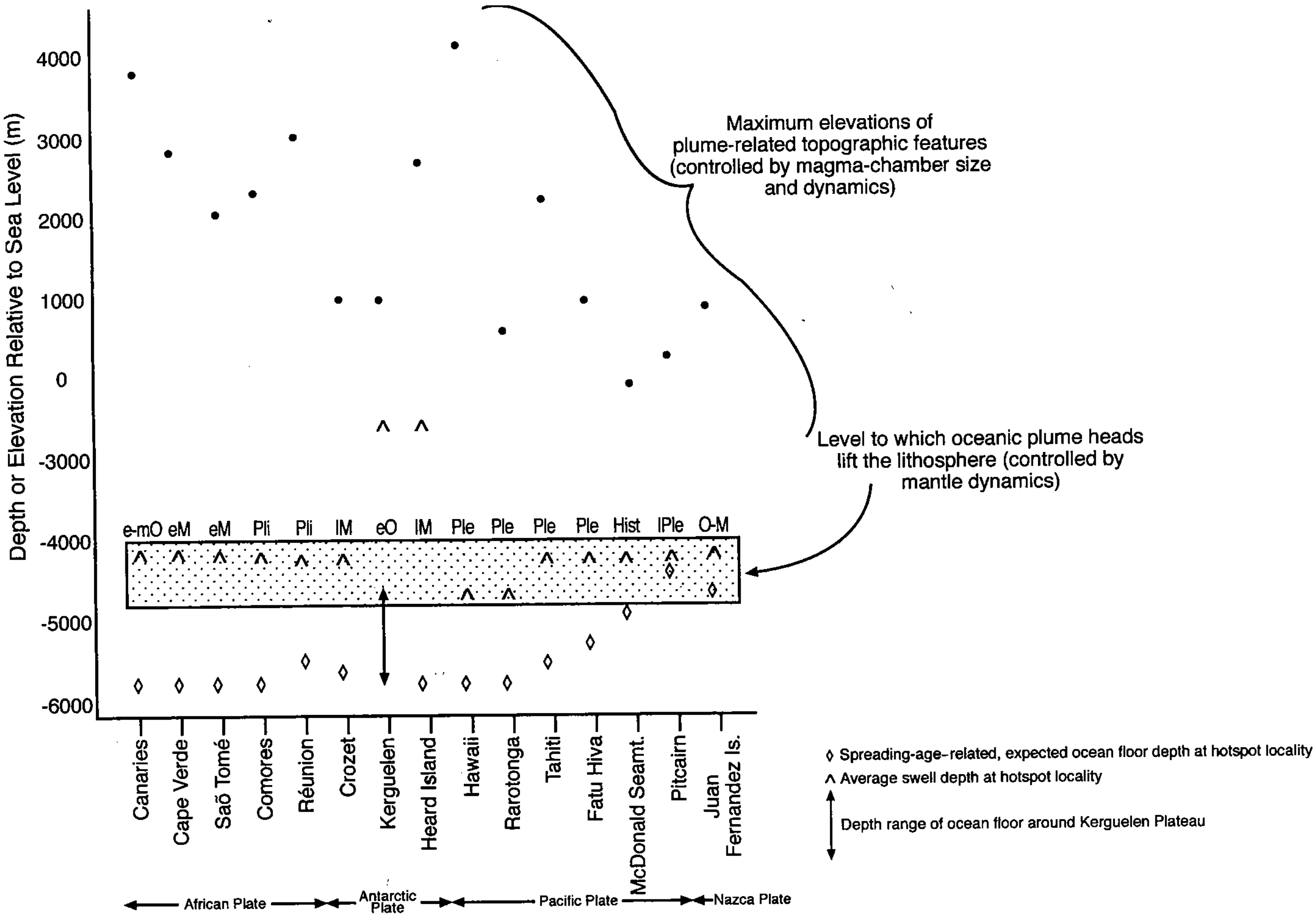

Figure 4. Plot showing both the plateau heights and the maximum elevations of the oceanic hotspots used in this study. Depths were read off the 5th edition of the GEBCO charts (1980), and the plateau outlines were determined visually on the basis of conspicuous shallowing with respect to their surroundings. Beneath the hotspots are the age-related expected depths of the oceanic basement on which they sit. The plateau heights are isostatically and mechanically controlled and are functions of the crustal and lithospheric thicknesses and plume dynamics. By contrast, maximum heights of oceanic hotspots represent volcanic constructions, the building of which are dependent on magma-chamber size and dynamics, which affect magma-supply volume and rate, and also on the local rainfall that controls erosion. Above the hotspot plateau heights are ages of origin of the hotspots (O—Oligocene, M—Miocene, Pli—Pliocene, Ple—Pleistocene, Hist—Historical, e—early, m—middle, l—late). For a summary display and the sources of data, see Table 1.

through the craton's *oldest* part (3.5 Ga: Blenkinsop et al., 1997),[5] or the Early Cretaceous alnoites erupting through the Archean lithosphere of the Prince Charles Mountains in East Antarctica (Andronikov and Egorov, 1993; Mikhalsky et al., 1993). Furthermore, the Deccan Traps and even the associated picrites (Narain, 1984) appear to have erupted through an Archean craton (see map in Kumar, 1985, p. 47; the most recent—but sparse—age information from the Bhandara craton is consistent with this interpretation [Rogers and Giral, 1997]). Raja Rao (1984) and Mahoney (1988) have summarized the observations showing how little control the existing Gondwana rifts exercised on the eruption of the Deccan Traps, the eruptive centers of which align along the north-trending Archean fabric of the Dharwar craton (see also Chandrasekharam and Parth, as in Parthasarathy, 1984). There are all sorts of magmas, including carbonatites, erupted through Archean cratons (see Kampunzu and Popoff, 1991, Fig. 2, and Burke, 1996, Figs. 6 and 23; compare with Ashwal and Burke, 1989, Fig. 1, for African examples), and these owe their origin to some localized heat source rising from below the lithosphere (e.g., Nixon and Davies, 1987, Figs. 372 and 373; Tainton and McKenzie, 1994), though it is not easy, even for plumes, to destroy a well-developed tectonospheric armor (Morgan et al., 1995).

(3) The observed heat-flow anomaly in Hawaii may be as little as 4 $mW \cdot m^{-2}$ (Stein and Stein, 1992), i.e., almost within observational error. If hydrothermal circulation within the vol-

[5] King and Anderson's (1995) argument loses power also when one considers that the present-day Archean shield surface is hardly 2% of the total exposed land surface. Though common, it is dangerous to base an empirical statement pertaining to generalities on that amount of information.

TABLE 1. OCEANIC HOTSPOTS USED IN THIS PAPER AND SOME OF THEIR PROPERTIES

Name*	Magnetic anomalies[1]	Expected depth[2] (m)	Observed plateau depth (m)	Depth anomaly (m)	Time of origin	Last eruption[3]
1. Canaries	Between M25 and African continental margin	5630	4000	1630	Early to middle Oligocene (ca. 30 Ma)[4]	Historic
2. Cape Verde	M0-M 16	5630	4000	1630	Early Miocene (19 Ma ?)[5]	Historic
3. Saõ Tome	?[6]	5630	4000?	1630	Early Miocene (15.7 Ma)[7,8]	100 000 yr ago[7]
4. Comores	M17	5630	4000	1630	Pliocene (<4 Ma)	Historic
5. Crozet	32	5490	4000	1490	Late Miocene (>8.1 ± 0.6 Ma)[9]	Holocene
6. Réunion	28	5322	4000	1322	Pliocene (2.1 Ma[10], possibly <5 Ma)	Historic
7. Kerguelen	SW:M0-M8[11] NE: north to south 11, 13, 18	SW: 5630 NE: 4480-4768	2500	3130-1960 (mostly owing to thick crust)	Early Oligocene (35 Ma),[12,13]	Fumarolic
8. Heard Island	W: M0-M8, E: 18[8]	SW: 5630 NE: 18	2500	3130-1960 (mostly owing to thick crust)	Late Miocene[14]	Historic
9. Hawaii	Beyond 34	5630	4500	1130	Pleistocene (>0.5 Ma)[15,16]	Historic
10. Marquesas (Fatu Hiva)	N: 24, S: 22	N: 5119 S: 5023	4000	N: 1119 S: 1023	Pleistocene[17] (~1.5 Ma)	Pleistocene (~100 000 yr ago)
11. Tahiti	30	5365	4000	1365	Pleistocene (<1 Ma)	No historic eruption[18]
12. Rarotonga	Beyond 30	5630	4500	1130	Pleistocene (ca. 1 Ma)	No historic eruption
13. McDonald Seamount	18?	4768	4000	768	Possibly historic	Historic (Hydrophonic)
14. Pitcairn	7	4285	4000	285	Late Pleistocene (0.93 – 0.45 Ma)[19]	No historic eruption
15. Juan Fernandez	12	4510	3900	610	?Oligocene-Miocene[20]	Historic[21]

* Numbers refer to Figure 3.
[1] From Cande et al. (1989) unless otherwise indicated.
[2] From the empirical relationship depth = 2500 + 350 $t^{1/2}$, where t is the age of ocean floor.
[3] From Simkin et al. (1981) unless otherwise indicated.
[4] Araña and Ortiz (1991).
[5] Emery and Uchupi (1984, p. 475).
[6] Cretaceous qartzose sandstones with metamorphic minerals is the lowest observed unit in the island: see Déruelle et al. (1991, p. 307–308, and the references therein).
[7] Grunau et al. (1975).
[8] Déruelle et al. (1991).
[9] Chevallier et al. (1983).
[10] Stieltjes (1986).
[11] Nogi et al. (1991).
[12] Duncan and Storey (1992).
[13] Bitschene et al. (1992).
[14] Clarke et al. (1983).
[15] Langenheim and Clague (1987).
[16] Peterson and Moore (1987).
[17] Dickinson (1998).
[18] But Moua Pihaa within 150 km may have had two eruptions in 1969 and 1970 (see Simkin et al., 1981, p. 89).
[19] Duncan et al. (1974).
[20] I am not aware of a study establishing directly the age of the Juan Fernandez hotspot. Biological evidence indicates Eocene-Oligocene (Kuschel, 1963), but is not terribly reliable (McBirney and Willims, 1969, p. 105). Lacroix's original opinion that the volcanic islands are of late Tertiary, Pleistocene, in some cases Holocene age is shared by later workers (e.g., Quensel, 1952, p. 41 and 81). Kay et al. (1988) noted that the shallowing of the slab to the east of the Juan Fernandez Ridge probably began at 18 Ma (early Miocene), but accelerated at 11 Ma (middle Miocene). If this is because of the introduction of a buoyant object into the subduction zone such as an aseismic ridge with a thermally eroded lithosphere beneath, then the age of that object is at least as old as the onset of shallowing. I would feel comfortable with a latest Oligocene-early Eocene age for the Juan Fernandez Ridge.
[21] Quensel (1952).

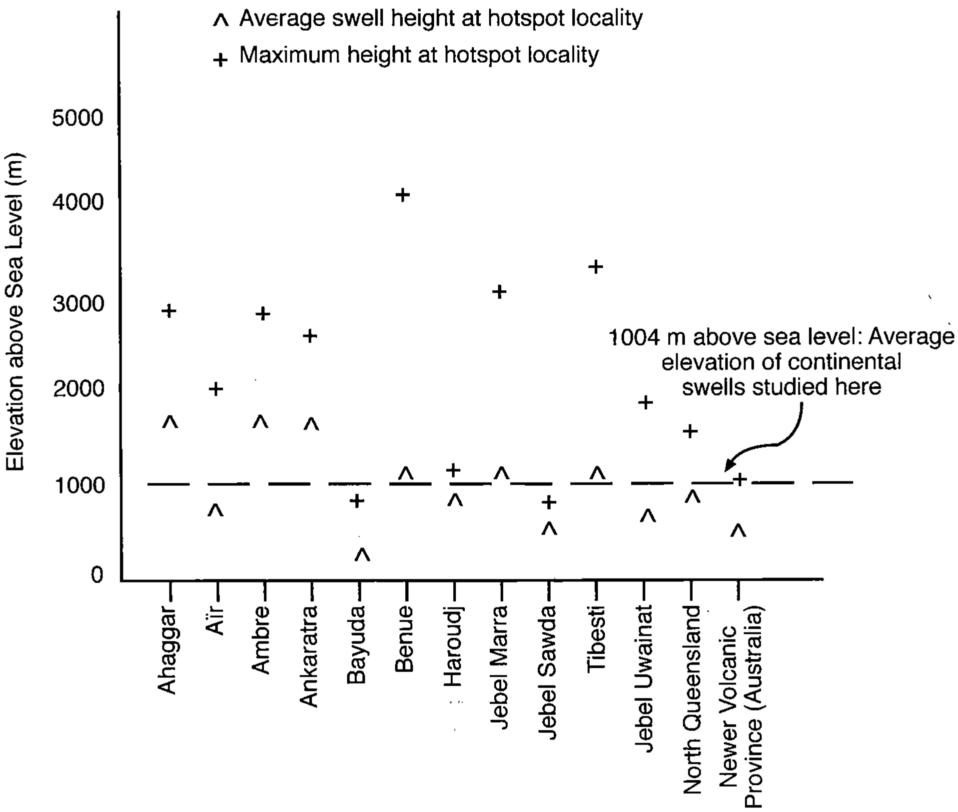

Figure 5. Plot showing both the plateau heights and the maximum heights of the continental hotspots used in this study. Elevation data are from the 9th edition (1992) of *The Times Atlas of the World*. For data display, see Table 2. Hotspot plateaus were outlined visually on the basis of conspicuous rise in elevation with respect to surrounding plains and checked against geology.

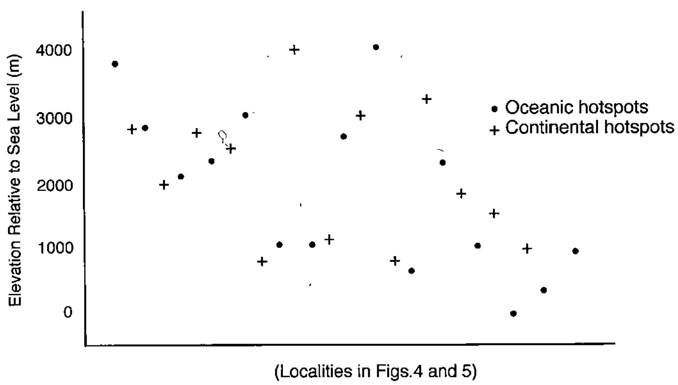

Figure 6. Plot showing a comparison between the maximum elevations attained by oceanic and continental hotspots considered in this study. The ordinate of the graph contains the ordinates of both Figure 4 and 5, but localities are not shown on the abscissa to prevent cluttering (they are the same as those in Figs. 4 and 5). Note the great similarity of the ranges of maximum elevations of both hotspot populations that have very different plateau heights.

canic edifice is considerable (e.g., Fournier, 1987), if water extends farther down into the 13-km-deep volcanic edifice (Sparks, 1981) than commonly thought—in view of the very deep, fluid-filled fractures encountered in the Kola superdeep well—and if hot rock comes closer to water at the sides of the volcano than usually thought (Fournier, 1987), then convective heat removal may take out significant quantities of heat.

(4) There is evidence for the thinning of the lithosphere (and the elastic layer) over the South Pacific superswell (Calmant and Cazenave, 1987; McKenzie, 1994). I must also emphasize that the lithospheric-thinning and the dynamic-support models cannot be exclusive of each other (see the discussion in Detrick et al., 1989, p. 182–187; Kevin Burke, 2000, written communication).

Figure 7 points out a significant consequence of these observations. The lower boundary of the lithosphere beneath both oceanic and continental hotspots in general lies within a depth range of 35–65 km below sea level. If continental crustal thickness is taken as 35 km, then to attain an elevation of 1 km, reduction of the thickness of the lithospheric-mantle root to only 40 km would be necessary (Fig. 7). It looks as if plume heads reach a hydrostatic and/or thermal level somewhere at the -35 to -40 km depth. Those plume heads under the continents commonly do not make it that high below unrifted domes, presumably because, compared to their counterparts rising under oceanic lithosphere, they have more (and more refractory) lithosphere to devour. When they do devour it, they raise the surface higher, by which time rifts will have disrupted the rising dome and uplift becomes more complicated. Such a mechanism more readily explains the fairly uniform average heights, with a good measure of scatter about the mean, of hotspot-related highlands than if they were a consequence of draping of a uniform-thickness lithosphere over rising plume

TABLE 2. CONTINENTAL HOTSPOTS USED IN THIS PAPER AND SOME OF THEIR PROPERTIES

Hotspot (on continents)*	Location of center	Maximum elevation (masl)	Plateau height (masl)	Average structural relief (m)	Uplift radius (km)	Average topographic slope of uplift (°)	Total area created by doming-related extension (km²)	Cumulative dyke width at apex of uplift: rectangular plan/triangular plan (m)
AFRICA								
I. Ahaggar	23°18′N, 5°33′E	2918	1500–2000	5000	700	0.2	39.3	56/102
II. Aïr[1]	20°N, 8°38′ E	2000	500–1000	2500	500	0.1	19.7	39/78
III. Ambre Bobaomby	12°48′S, 49°10′E	2876	1500–2000	5000 (in part due to rifting)	Uncertain		Uncertain	Uncertain
IV. Ankaratra	19°40′S, 47°20′E	2643	1500–2000	9000 (in part due to rifting)	400[2]	0.3	127	317/634[3]
V. Bayuda	18°10′N, 32°45′E	800	300–500	300	Uncertain		0.2	Uncertain
VI. Benue	6°N, 10°E	4070	1000–1500	2500	800	0.1	19.7	25/50
VII. Haroudj[4]	27°25′N, 17°50′E	1200	750–1000	4000	180	0.3	25	138/276
VIII. Jebel Marra	14°N, 24°15′E	3071	1000–1500	1500	450	0.2	3.5	8/16
IX. Jebel Sawda[4]	28°45′N, 15°E	840	500–750	4500	250	0.2	31.8	127/254
X. Tibesti	22°N, 17°30′E	3325	1000–1500	4500	500	0.2	31.8	64/128
XI. Jebel Uweinat	21°56′N, 25°04′E	1893	500–1000	3000	350	0.2	14.13	40/80
AUSTRALIA								
XII. N. Queensland	20°02′S, 145°13′E	1622	750–1250	2250[5]	250	0.3	7.9	32/64
XIII. Newer volcanic province (western Victoria)	37°45′S, 143°15′E	1011	300–800	1100[6]	250[7]	0.3	1.9	8/16

Note: For comparison, please note that volcanic conduits range in diameter from several meters to nearly 1 km or so. Most typically they range from ~50 m for cinder cones of a similar height to ~500 m for volcanoes thousands of meters high. Most are surrounded by radial dike systems, such as the case in Shiprock (Williams, 1936), where the conduit width is ~400 m, and total extension represented by dikes near the conduit amounts to ~18 m. Conduits in general narrow downward; dikes enlarge downward. Had all these features been products of extension, instead of forceful injection, only some of the uplifts in the table would have been capable of accommodating one to a few volcanoes of the size of the Shiprock volcano. Some could not generate even one. It is clear that domal uplift of the lithosphere of the magnitudes shown here can at best generate a fractured lithosphere and probably lead to the formation of a conduit to support a volcano, but it cannot generate rifts. The sequence of events atop a plume-related uplift must thus be uplift → fracturing → volcanism → rifting. Sources: All elevations are from *The Times Atlas of the World,* 9th comprehensive edition, 1992. Structural relief for all African and Madagascar localities was computed from sedimentary thickness data in the *Tectonic Map of Africa,* scale 1:15 000 000 (Choubert, 1968) combined with the topographic data. For the Australian localities the following were used: Newer Province: Veevers (1984), Joyce (1988), Nicholls and Joyce (1989); Northern Queensland: Stephenson (1989).

* Numbers refer to Figure 3.
[1] Possible leakage from the Ahaggar plume, along reactivated north-south Pan-African structures.
[2] Eastern continental-margin fault delimits the uplift, where uplift radius diminishes abruptly to ~200 km.
[3] Calculated by ignoring the presence of eastern continental-margin fault and is therefore excessive.
[4] Very likely leakage from the Tibesti plume, according to Late Cretaceous Sirte rift structures (that may have in turn exploited old Pan-African north-south fabric).
[5] As measured from the pre-Tertiary basement of the Hillsborough and the Duaring basins, which are both ~1 km deep (Palfreyman, 1984; Veevers, 1984, p. 207, Fig. 130).
[6] Including the 500 m thickness of the Millewa Group in the Murray Basin (see Lawrence, 1988). Thicknesses in excess of 10 km have been reported in the Otway Basin to the south (e.g., Palfreyman, 1984, p. 83), but this is a rifted margin with much faulting and stretching.
[7] From Veevers (1984, Figs. 56A and 59A).

fountains of mantle material. It would be hard to imagine why such immense fountains would occasion such diversity of elevations tightly clustered close to a readily discerned mean.

A serious problem with the thermal-thinning hypothesis for the lithosphere is, as Detrick and Crough (1978) noted, the great difficulty of heating the mantle lithosphere adequately by conduction. The time scale involved to eliminate half of the average lithospheric thickness is nearly as long as the whole of the Cenozoic. There are good arguments, however, for believing that in most cases, a substantial chunk of the lithosphere is taken out in just a few millions of years upon encounter with a mantle plume. Detrick and Crough (1978) argued that Hawaiian lithosphere had been eliminated within 3 m.y., and Burke (1996) showed that African hotspots appeared as soon as the continent stopped or slowed sufficiently with respect to the mantle circulation to allow it to affect the overlying continent. These findings imply rates of thinning nearly twice as high as the most extreme rates allowed by Wendlandt and Morgan's (1982) and

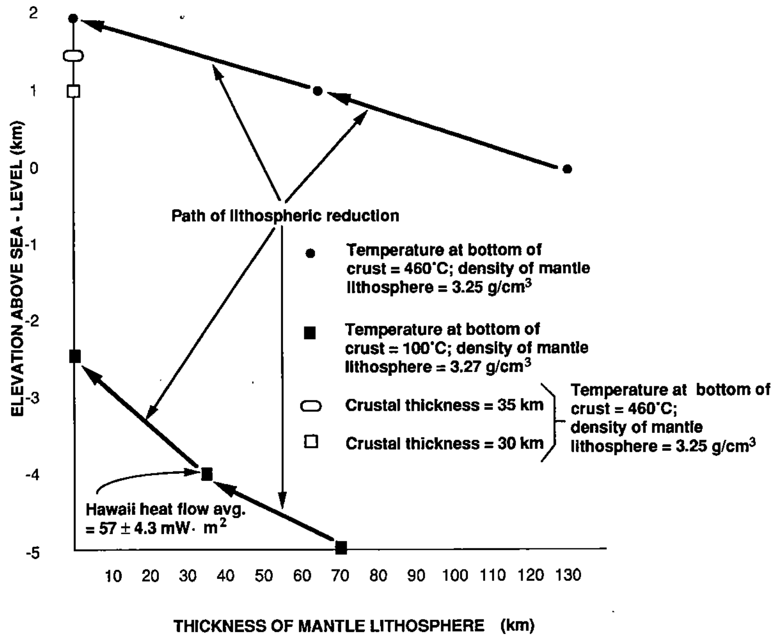

Figure 7. The effects of the removal of the mantle lithosphere from a constant continental crustal thickness of 40 km with average density of 2.85 g/cm^3(solid circle; Christensen and Mooney, 1995; Mary Lou Zoback, personal communication, 2000), and a constant oceanic crustal thickness of 8 km (solid square), with average density of 2.9 g/cm^3 (using) Lachenbruch and Morgan (1990, fig. 3).

Morgan's (1983) fast, conductive, thermal-thinning models, which also include tectonic thinning! Yuen and Fleitout (1985) and Dalloubeix and Fleitout (1989) have attempted to circumvent this difficulty by showing that secondary convection within the plume head may devour the superjacent lithosphere in some 5 Ma, but even that may not be fast enough. It seems clear that if the lithosphere is to be thinned so rapidly, it has to be done mechanically (see the various mechanisms suggested in Morgan, 1983, p. 283), *yet in a way to allow the underlying plume a controlling role.* I propose the following model to provide one possible scenario.

Because a plume head has a density lower than that of the surrounding asthenosphere (otherwise the plume cannot rise through), it uplifts, stretches, and fractures the overlying lithosphere (Fig. 8A). These fractures are likely to be filled with magma (for conditions for intrusive fracturing, see Fig. 8B). If melting occurs in a manner proposed by Morgan et al. (1995), then this effect is naturally enhanced. I do not think that the plume head can heat the overlying lithosphere in time scales of a few millions of years so as to uplift it thermally, unless there is virtual replacement of mantle lithosphere with intrusive material. Most lower crust does not give the impression that such replacement has occurred (e.g., see the papers in Salisbury and Fountain, 1990). If such replacement has occurred anywhere, it would have created an environment tectonically identical with what the model here proposed may bring about (see also Platt and England, 1994). According to this model, initially the entire lithosphere will rise into a barely perceptible dome owing to a mantle-density decrease that creates a network of fractures through which plume-related magmas may intrude the mantle lithosphere (Fig. 8B). Such intrusion may be further facilitated by the mechanism Yuen and Fleitout (1985) and Dalloubeix and Fleitout (1989) suggested. These fractures, enveloping an interconnected network of dikes and sills, may isolate a chunk of the mantle lithosphere and eventually begin to tear it away. This partly detached bit will weigh down the overlying carapace, thereby creating a sag basin. Eventually the weighted-down lithosphere may detach and sink into the plume head (similar to the spontaneous lower-lithosphere-detachment scenarios in Parsons and McKenzie, 1978, and Sclater et al., 1981; convective detachment, similar to the scenario developed by Houseman et al., 1981, may occur if lower lithosphere is removed in separate chunks, thereby providing horizontal density gradients around the later-removed bits and enhance the Yuen and Fleitout (1985) and Dalloubeix and Fleitout (1989) mechanism), while the thus liberated crust and any part of the mantle lithosphere still attached will rise into a major dome commonly adorned with alkalic volcanic rocks (Fig. 8C). Such a sinking into a vigorous plume head is possible despite the probable average 50 cm/yr upwelling rates owing to the much lower viscosity within the plume chimneys than the infalling lithospheric chunks (cf. Lliboutry, 2000, p. 445–446). The 50-km-deep Honuma event of 26 April 1973 beneath the Hawaiian dome (Unger and Ward, 1979) would have been hard to account for in such a hot place, if a detachment of mantle lithosphere had not taken place.

It is interesting that immediately before the onset of widespread rifting at 30 Ma, the whole interior of Africa was occupied by a shallow drainage basin (or a number of overlapping basins) reaching from the Okavango-Ngami basin to the Sahara (Summerfield, 1991, Fig. 16.25; cf. Haughton, 1963, Fig. 45; Tankard et al., 1982, p. 450). Similarly, structures that resemble sag basins existed atop the future Red Sea, Gulf of Aden, and the Kenya and Ethiopia parts of the Great Rift Valley (discussed

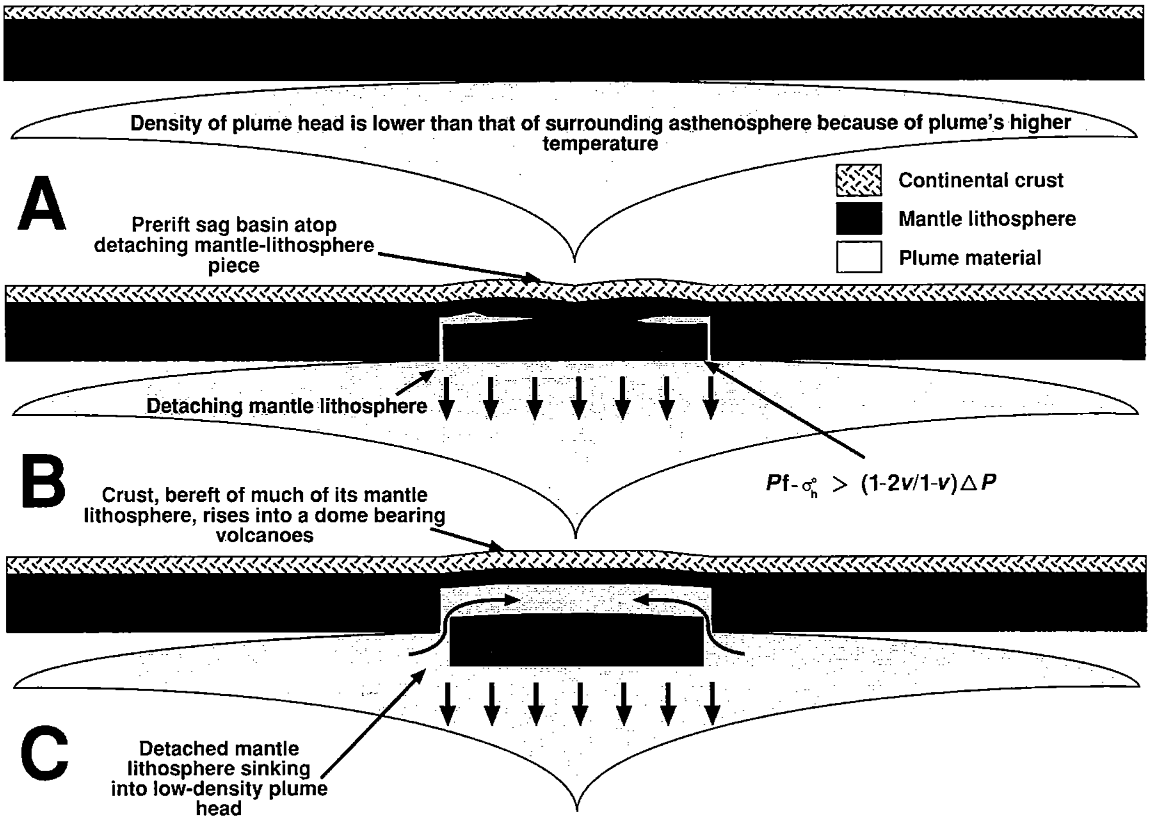

Figure 8. A possible mechanism to account for the very short time required (~1–3 m.y.) for the rise of hotspot swells after a plume arrives beneath them. A: Plume rises off the base of a lithosphere carrying a normal-thickness continental crust. Diameter of the plume head as depicted may range from hundreds of kilometers to more than 1000 km. B: The plume head must have a density lower than surrounding asthenosphere because it is ~200 to 300°C hotter than the ambient asthenosphere and actively rises through the asthenosphere. This process would raise, stretch, and fracture the overlying lithosphere. Such fractures, filled with magma from melting induced by the plume head, may isolate a piece of lithosphere atop the plume head, which will begin sinking into it. This sinking is made easier by the lower density of the plume head compared to its surroundings. The sinking piece may—but does not have to—initially entrain a part of the overlying crust and then pull it downward to form a broad and shallow basin above. The formula indicates the condition of formation of vertical dikes into the bottom of the lithosphere (from Mandl, 1988, p. 274): P_f = pressure of fracture fluid; σ°_{h} = smallest principal horizontal stress; n = Poisson's ratio; ΔP = pore-liquid (in this case, magma) pressure. C: When the mantle lithosphere detaches completely, the overlying crust will rapidly rise and form a major dome. Surrounding asthenosphere will penetrate into the space vacated by the sinking mantle-lithosphere piece, undergo adiabatic melting, and feed volcanism above. The detaching piece of lithosphere need not separate along the crust-mantle boundary, but may use any near-horizontal plane where injection pressure of magma equals the overburden stress (Mandl, 1988, p. 274). I think that such variations may account for the scatter of plateau heights about the mean of 1000 m (see Fig. 5).

subsequently). Broad and shallow basins also formed on the future site of the Gulf of Aden (Bosellini, 1989). One critically important predictive aspect of the lithosphere-detachment model for hotspot-related highlands is thus borne out by data from a number of well-known cases.

UPLIFT AND EXTENSION

As there are domes today atop what most people think to be mantle plumes, it might be useful to know what sort of structural consequences dome generation may have. To investigate this, one must first calculate the amounts of extension that doming generates. In this section, I calculate the areal increase of domes by assuming three different dome configurations to emphasize the resulting differences and to show whether any one of them can be used as a satisfactory approximation for the others.

The question of structural consequences of doming was first addressed by Élie de Beaumont in 1834 within the framework of the ill-fated craters of elevation theory (Dufrénoy and Élie de Beaumont, 1834). Élie de Beaumont followed Leopold von Buch in thinking that large shield volcanoes were, in reality, domes of uplift and the radial rills (*barrancas*) around the sum-

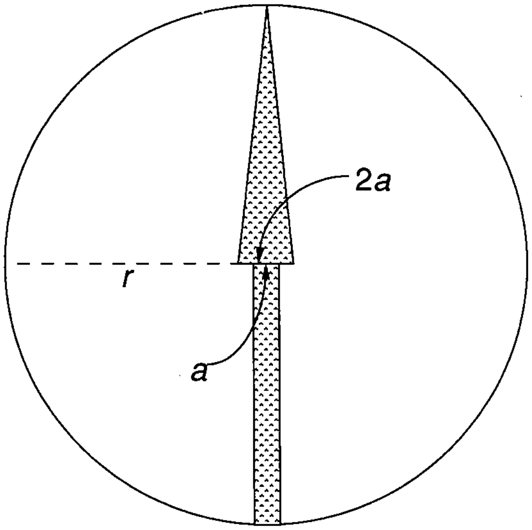

Figure 9. A circular uplift of radius r with two dikes of identical map area disrupting it. One dike is rectangular in map view, the other an isosceles triangle. The triangular dike will have twice the width ($2a$) at the base (i.e. at apex of dome) as the rectangular dike (a) See Table 2.

mit craters were fissures formed by doming-related stretching. He assumed the domes to be perfect cones and calculated the doming-related extension to be

$$\pi R\sqrt{R^2 + H^2} - \pi R^2$$

where R is the basal radius and H is the height of the cone. If I solve this expression, it opens into an infinite series. However, for cases in which $H \ll R$, which obtains for most shield volcanoes in which Élie de Beaumont was interested (and very especially for the domal uplifts of the lithosphere, which are of interest here!), the series can be very closely approximated by $\frac{1}{2}\pi H^2$. Note that the radius of the uplift has fallen out of the solution and only the magnitude of the uplift appears to matter.

Assuming all uplifts listed in Table 2 to be perfect cones (an inappropriate assumption as I show next!), I calculated the amount of extension they generate and expressed this in two ways: One, as the area of the outcrop of a *cumulative dike,*[6] which is assumed to be a rectangle, and the other in which the dike's map view is assumed to be an isosceles triangle with its apex at the circumference of the basal circle of the cone (Fig. 9). In each case, the width of dike used is that near the apex of the dome. The results are listed in Table 2. Under these conditions, none of the hotspot uplifts in Africa could generate more than 300 m of extension, and only Ankaratra on Madagascar could create 634 m. This last value is very uncertain, however, for a good part of the assumed structural relief in Ankaratra is a consequence of the subsidence of the nearby faulted margin, i.e., it is not falcogenic, but copeogenic (i.e., it has a short wavelength).

[6] The reason I use the expression "cumulative dike" here instead of "cumulative extension" is that what is calculated is areal increase representing a parallel-walled opening. Such an opening would be like a dike and not like a rift. Whether such a dike actually forms is not relevant to my argument here.

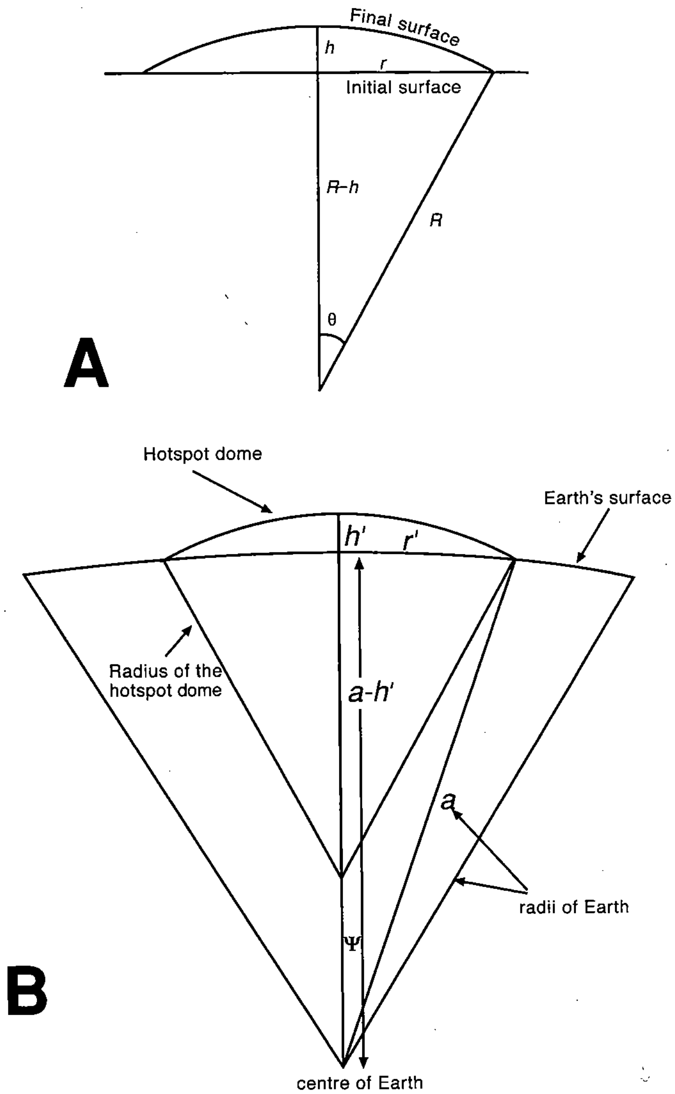

Figure 10. A, Elements necessary in the calculation of areal increase upon uplift of spherical-cap-shaped highland rising on a flat surface. B, Elements necessary in the computation of the areal increase upon uplift of a spherical-cap-shaped highland on the surface of a sphere with radius a.

The amount of extension will be more if one assumes spherical-cap—shaped uplifts rather than perfect cones (Fig. 10A). To calculate the amount of extension consequent upon doming in such a case, I take the area of the initial circle to be πr^2 (see Fig 10A). The area of the shell is $2\pi R^2(1 - \cos\theta)$. By straightforward trigonometry $R = (r^2 + h^2)/2h$ and $\cos\theta = 1 - (h/R)$; thus the change in area is πh^2. Note that the area generated by extension resulting from doming in a spherical uplift is *exactly twice* that in a conical uplift. Table 3 lists the amounts for each of the continental hotspot highlands considered herein.

TABLE 3. EXTENSION GENERATED BY A SPHERICAL UPLIFT ON A FLAT BASE

Hotspot (on continents)	Total area created by doming-related extension (km²)	Cumulative dike width (m) at apex of uplift: rectangular plan/ triangular plan
AFRICA		
Ahaggar	78.5	112/224
Aïr	19.7	39/78
Ambre Bobaomby	Uncertain	Uncertain
Ankaratra	254.34	635/1270
Bayuda	0.2	Uncertain
Benue	19.7	39/78
Haroudj	50.24	279/558
Jebel Marra	7	15/30
Jebel Sawda	63.6	254/508
Tibesti	63.6	127/254
Jebel Uweinat	28.3	80/160
AUSTRALIA		
N. Queensland	16	64/128
Newer volcanic province (western Victoria)	3.8	16/32

Note: See Figures 9 and 10.
height = *h*, radius = *r*

TABLE 4. EXTENSION GENERATED BY A CIRCULAR UPLIFT ON THE SURFACE OF A SPHERE

Hotspot (on continents)	Total area created by doming-related extension (km²)	Cumulative dike width (m) at apex of uplift: rectangular plan/ triangular plan
AFRICA		
Ahaggar	1206	1723/3446
Aïr	307.7	615/1230
Ambre Bobaomby	Uncertain	Uncertain
Ankaratra	709	1772/3544
Bayuda	Uncertain	Uncertain
Benue	787	984/1969
Haroudj	64	354/177
Jebel Marra	150	332/664
Jebel Sawda	138.5	554/1108
Tibesti	553	1108/2216
Jebel Uweinat	28.3	181/362
AUSTRALIA		
N. Queensland	69	276/552
Newer volcanic province (western Victoria)	33.8	135/270

Note: See Figures 9 and 10.
height = *h*′, radius = *r*′ on the surface of a sphere with radius *r* = *a*

It is surprising that the amount of extension increases much more still if the spherical uplift occurs on the surface of a larger sphere and not on a flat base (Fig. 10B). If the initial surface of a dome is itself a spherical cap (of angular radius r') of a sphere of radius a uplifted above that surface by a height h' (Fig. 10B), then the change in area as a result of doming can be given as $\pi h'(h' + 2b)$, where $b = a[1 - \cos(r'/a)]$. If $r'/a \ll 1$ (which, in the case of terrestrial domes, certainly is true), then the expression for change in area owing to doming can be rewritten as $\pi h'(h' + r'^2/a)$. In this last equation, the h' term is ~200 times smaller in magnitude than the r'^2/a term. Therefore, the h' term can be ignored without much harm to the precision of the calculation of areal increase upon doming. The expression for change of area consequent upon uplift then reduces to $\pi h' r'^2/a$. Table 4 lists the resulting extension for the continental hotspot domes considered in this paper. The differences between the amounts of extension obtained on spherical uplifts rising over flat bases and those rising over spherical bases are considerable, ranging from 3 times more to 40 times more. The amounts of extension expressed as cumulative dike widths range from 135 m to nearly 4 km. It is clear that none of the three approximations used for computing can be a reasonable approximation for the others.

The question now becomes, What do these increased amounts of extension do structurally? What kind and what magnitude of structures can they generate?

Table 5 is a list of selected rift valleys and the amounts of extension measured across them.[7] A number of them have been claimed to have resulted from doming-related areal increase, mainly following Cloos's (1939) seductive, but misleading, model. As can been seen from the numbers obtained through various methods, however, all have far too much extension to have resulted from a dome-shaped intumescence of the lithosphere. Therefore, one positive conclusion is that *doming-related areal increase* is not a cause for rift genesis atop mantle plumes. One might be tempted to glide from this conclusion to the claim that no rift has ever been preceded by doming. Such a claim has in fact been made frequently in the literature, yet it is even more common to think that domes do precede at least those rifts considered to be genetically related to mantle plumes. In the next section, I examine the geologic history of one of the best-known and most controversial of the major domes, which is dissected by a major rift and also by a growing arm of ocean, namely, the Afar (or Ethiopian-Yemeni) dome, with a view to establishing whether doming preceded rifting.

THE EVOLUTION OF THE AFAR DOME AND TRIPLE JUNCTION

The Afar dome and triple junction (see esp. Manighetti et al., 1997 and 1998, for the detailed structural geology of the triple junction) together form the grandest structure of the East African dome-and-rift system, which is an immense taphrogen reaching northward from south of Lake Malawi to the shores of the Red Sea and the Gulf of Aden (Fig. 11; Suess, 1891, 1909, p. 304–317; Şengör, 1995; Burke, 1996). In fact, before sea-floor spreading commenced at ca. 10 Ma in the Gulf of Aden and later in the Red Sea, the rifts that led to their opening were parts of the East African taphrogen. I first examine the paleotopography around the future Afar dome, to see whether

[7] Taphrogens such as the Basin and Range in the United States and the West Siberian basin in Russia are not included because the large extensions measured across them are clearly related to plate-boundary events and not to plumes, although plumes, such as that under Yellowstone, may contribute (see especially Atwater and Stock, 1998; Wernicke and Snow, 1998).

TABLE 5. AMOUNTS OF EXTENSION AND SHOULDER UPLIFT AT SELECTED CONTINENTAL RIFTS

Name of rift	Amount extended (km)	Uplift of shoulder area (km)	Reference
N. Kenya[1]	35–40	>1.4–1.7	Bishop and Trendall (1966: for shoulder uplift), Morley et al. (1992: for extension)
Gregory[1]	10	2	Schlüter (1997)
Malawi[1]	7.5 (from transfer fault offset)	2.8 (strike-slip affected?)	Krenkel (1925, p. 233: for shoulder uplift), Tiercelin et al. (1988: for extension)
Suez North	16.3	1.5	Patton et al. (1994)[2]
Suez Central	17.7–32.6		
Suez South	29.3 (all from fault geometry)	~500 m	
Upper Rhine[1]	17 (from crustal configuration), 5–7 (from fault geometry)	2.2	Illies (1967: for shoulder uplift), Brun et al. (1992: for extension)
Oslo	28–36 (from crustal configuration), 11–13 (from fault geometry)	1	Ro and Faleide (1992)
Viking	100–130 (from crustal configuration), 30 (from fault geometry)	3	Ziegler (1992)
Central (North Sea)	100–105 (from crustal configuration), 15 from fault geometry)	1.5	Ziegler (1992)
Benue[1]	100[3]	?	Burke et al. (1971)
Baykal[1]	15–20 (from crustal configuration), 10 (from fault geometry)	2–3	Logatchev and Zorin (1992)
Española basin[1]	5.5	1.5	Baltz (1978: for shoulder uplift), Golombek et al. (1983: for extension)
Albuquerque Basin North[1]	10 (from fault geometry)	~1	Lambert (1978, fig. S55: for shoulder uplift),
Albuquerque Basin South[1]	16 (from fault geometry)	0.9	Russell and Snelson (1994: for extension)

Note: Owing to probable onset of seafloor spreading in the Wonji fault belt, the Ethiopian segment is not included in this table.
[1] Rift for which a doming-related origin has been seriously claimed.
[2] This paper lists all previous estimates of extension on the Gulf of Suez rift. The ranges are as follows: Northern Suez 4–16.3 km; central Suez 17.7–32.6 km; southern Suez 15–40 km. For references, see Table 1 in Patton et al. (1994).
[3] Benkhelil et al. (1988) established at least 10 km of shortening during the following compression.

the dome existed before the onset of volcanism (ca. 30 Ma) and rifting (ca. 35 Ma). To that end, it is helpful to start with the early Eocene paleogeography of the northeast quadrant of Africa, including the whole of the Arabian peninsula.

Paleogeography north of Afar

Contrary to some recent claims (e.g., Bohannon et al., 1989), a *complicated* coastline limited Africa to the northeast during the early Eocene (Fig. 12). A prominent feature of this coastline was three pronounced north-directed "fingers" jutting into the southern shelf of the Neotethys (cf. Şengör and Natal'in, 1996a) as long and narrow peninsulas. In Arabia, they followed the so-called "Arabian trend" of late Pan-African age (see Şengör and Natal'in, 1996a) and took the form of the Hail-Rutbah arch (Fig. 12; see al-Laboun, 1986, Fig. 5). Here, the elevation of the early Eocene peninsula above sea level could not have been very much, for to the north and east, it is now surrounded by the Umm er Radhuma carbonates that are singularly free of any terrigenous detritus (Powers et al., 1966; Bohannon et al., 1989, p. 1686; Jones and Racey, 1994). The western boundary is determined by two formations, whose ages remain controversial. Figure 13 shows the Umm Himar succession (stratigraphic column A) as reported by Madden et al. (1980) and, on the basis of their work, by Brown et al. (1989, Fig. 44). Figures 11 and 12 show the location of Jabal Umm Himar, after which the formation was named by Madden and his collaborators. All the Paleogene around the Harrat Hadan belongs to this formation (Brown et al., 1989, p. A118). The middle mudstone and shale member contains fossils of sharks, rays, other fishes, turtles, primitive crocodiles, and a primitive lungfish. Because both the fossil assemblage and the lithologic succession greatly resemble those of the so-called infrabasaltic sedimentary rocks at Jabal Oharba (17°13′N, 35°34′E: Figs. 11, 12, and B in 13: Delaney, 1954; Vail, 1988, p. 78–79) in Sudan and because Delaney (1954) found wrasse and catfish fossils giving a range from Eocene to Recent in the latter, I am inclined to date both the infrabasaltic sedimentary rocks of Sudan and at least the upper part of the Umm Himar Formation as Eocene (to earliest Oligocene? Vail [1988, p. 79] says "perhaps late Tertiary" for which I see no good reason). Latest Oligocene or earliest Miocene latite and phonolite intrude the Umm Himar Formation (Brown et al., 1989, p. 119).

The Usfan Formation (Abou Ouf and Gheith, 1998) just northwest of Mecca (for location, see Figs. 12 and 13) is better dated. Two K-Ar glauconite ages (may be unreliable!) and a fossil assemblage containing *Venericardia sindensis, Phacoi-*

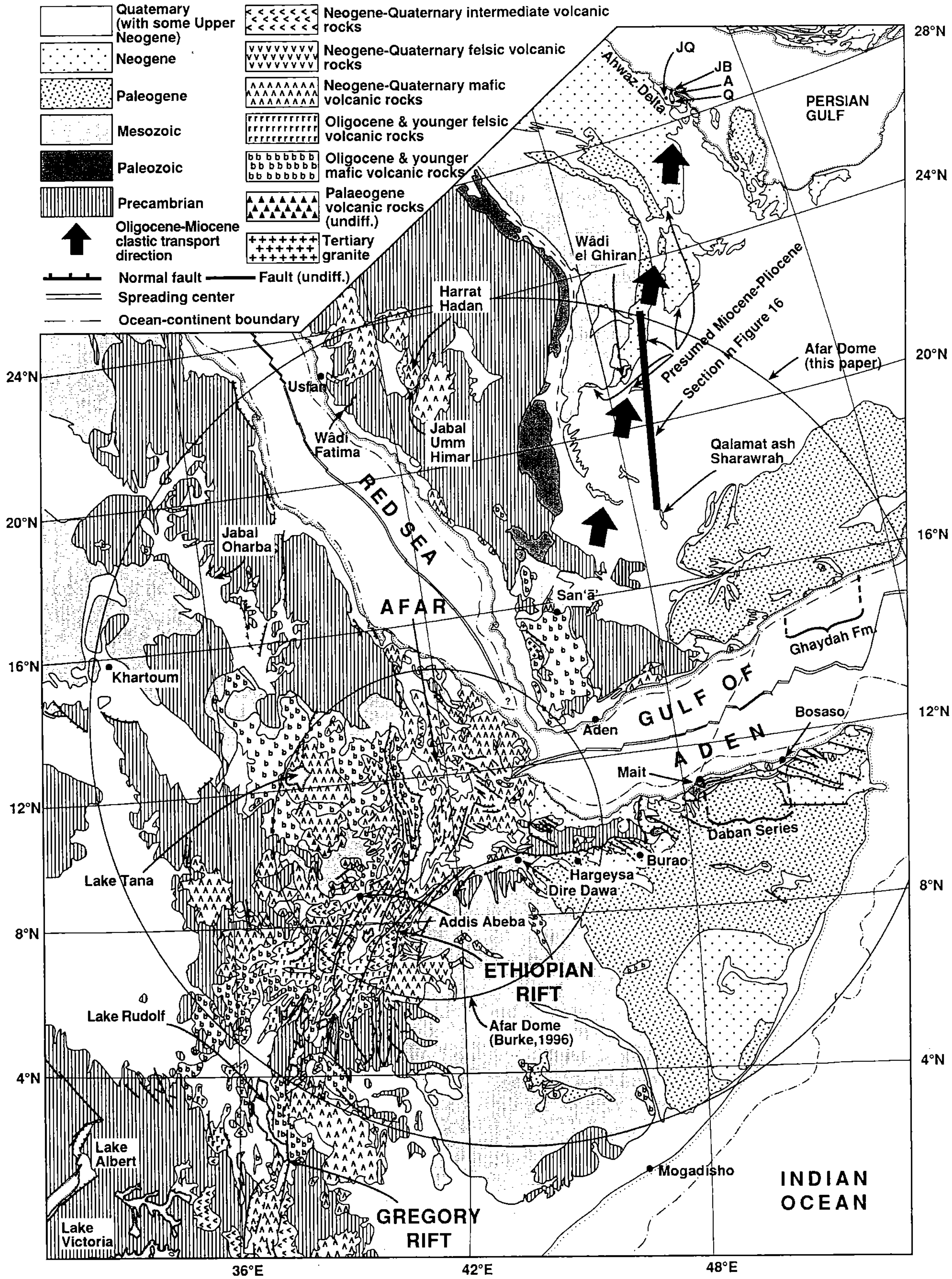

Figure 11. A simplified geologic map of the Afar dome and the surrounding regions (simplified from Choubert and Faure-Muret, 1985, sheet 3). Abbreviations in northeastern part: A—Al ᶜAlâh, JB—Al Jubayl al Barri, JQ—Jabal Qurayn and Q—Al Qatif.

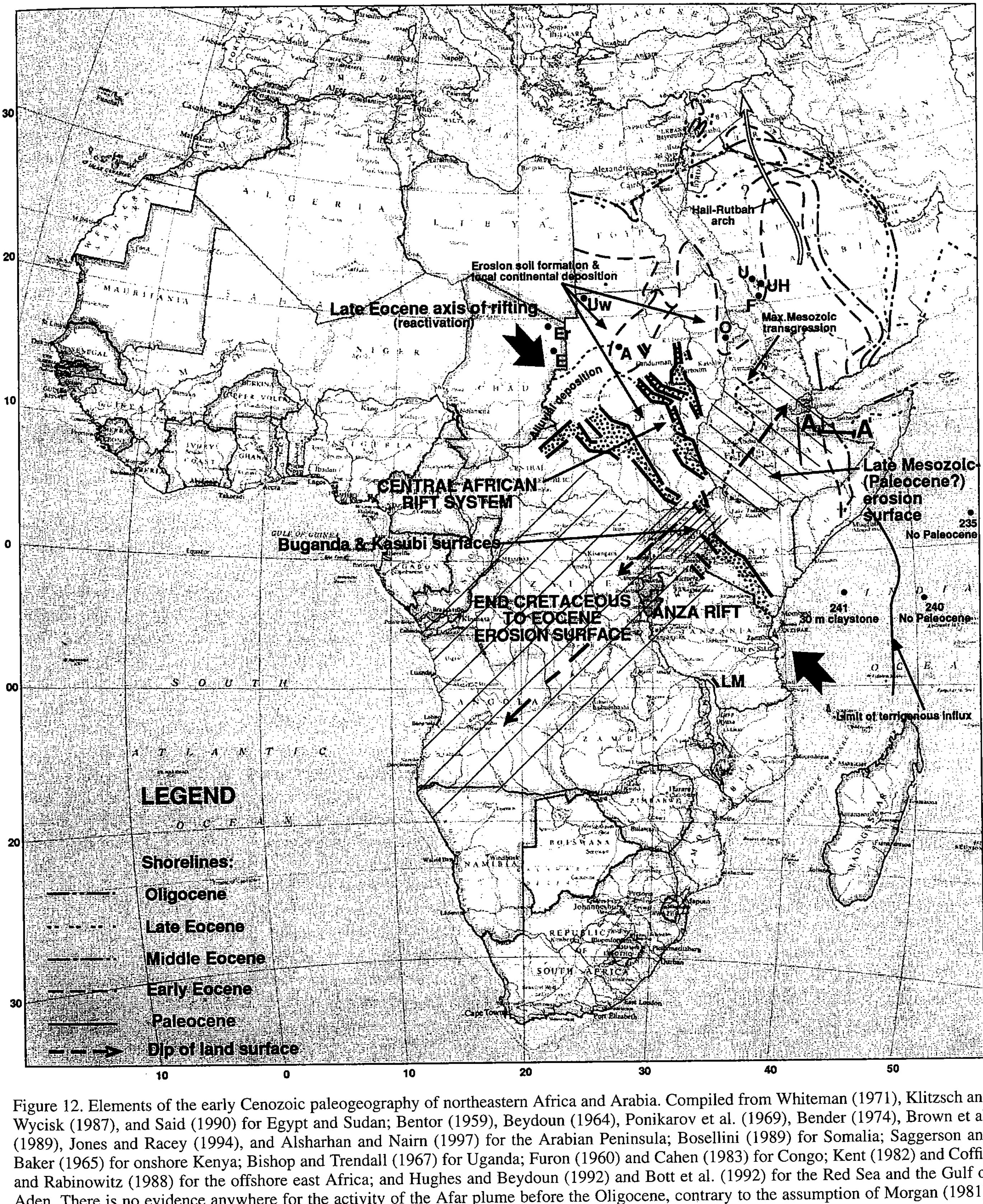

Figure 12. Elements of the early Cenozoic paleogeography of northeastern Africa and Arabia. Compiled from Whiteman (1971), Klitzsch and Wycisk (1987), and Said (1990) for Egypt and Sudan; Bentor (1959), Beydoun (1964), Ponikarov et al. (1969), Bender (1974), Brown et al. (1989), Jones and Racey (1994), and Alsharhan and Nairn (1997) for the Arabian Peninsula; Bosellini (1989) for Somalia; Saggerson and Baker (1965) for onshore Kenya; Bishop and Trendall (1967) for Uganda; Furon (1960) and Cahen (1983) for Congo; Kent (1982) and Coffin and Rabinowitz (1988) for the offshore east Africa; and Hughes and Beydoun (1992) and Bott et al. (1992) for the Red Sea and the Gulf of Aden. There is no evidence anywhere for the activity of the Afar plume before the Oligocene, contrary to the assumption of Morgan (1981), except perhaps in the gradual thickening of the Dammam Formation (lower and middle Eocene) from the shores of the Persian Gulf westward. Geographic base is from the 9th edition of the *Times Atlas of the World,* map 84. Key to abbreviations: A—Jabal Abyad, E—Ennedi massif, Er—Erdi Dji, F—Wadi Fatima, Uw—Jabal Uweinat, LM—Livingstone Mountains, O—Jabal Oharba (Khor Durundum is only 30 km to the east; see Fig. 13B), U—Usfan, UH—Jabal Umm Himar. DSDP (Deep Sea Drilling Project) Sites 235, 240, and 241 are shown.

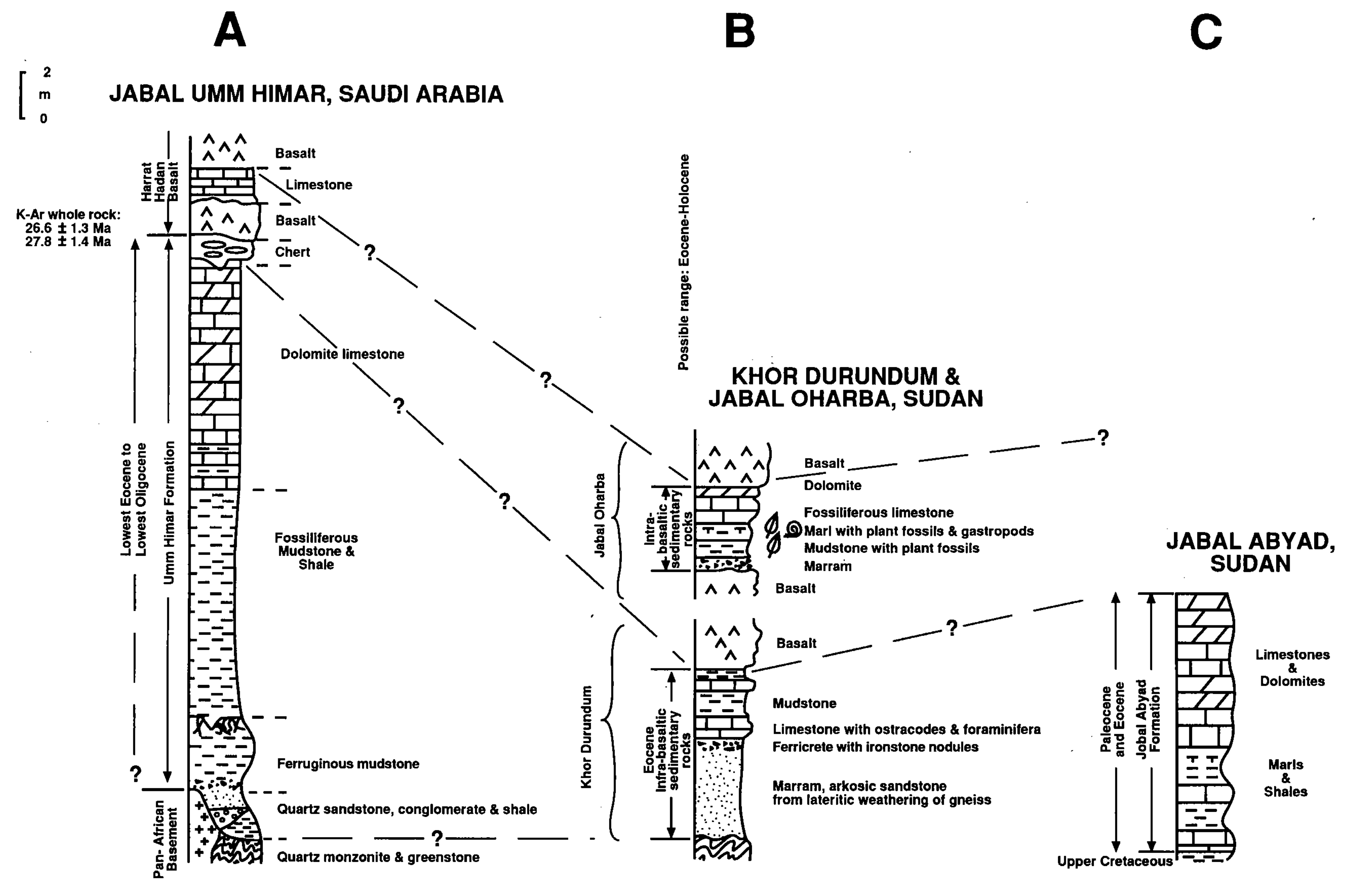

Figure 13. Stratigraphic columns. A, The Umm Himar succession in the Jabal Umm Himar region, western Saudi Arabia (after Brown et al., 1989, Fig. 44). For location, see Figures 11 and 12. Compare with column B. B, The infrabasaltic sedimentary rocks at Khor Durundum (17°12′N, 35°53′E) and Jabal Oharba (17°13′N, 35°34′E) in Sudan, drawn from the description in Whiteman (1971, p. 101–103). Of these, only the succession at Jabal Oharba is fossiliferous. Compare this with column C. C, The typical sequence of the Jabal Abyad Formation at Jabal Abyad, Sudan (~19°N, 29°E; redrawn from Klitzsch and Wycisk, 1987, Fig. 19 therein). Compare with columns A and B. I consider all three successions displayed in this figure to be time correlative. None has fossils with narrow enough ranges to permit detailed correlations. All are commonly agreed to have a substantial thickness representing some part of the Eocene. If the limestone and marl at the Jabal Oharba section (B) really do occupy a stratigraphic position similar to the lacustrine Oligocene Hudi Chert along the Nile Valley (Whiteman, 1971, p. 82 and 147), then the upper parts of these sections may extend well into the Lower Oligocene. In the Umm Himar section (A), the lowest basalt flows of the Harrat Hadan are dated at ca. 27 Ma (Brown et al., 1989, p. 160), which is early late Oligocene. Beneath them is a formation of white chert. These conditions in nearby Saudi Arabia give support to what seem tentative correlations and age assignments in Sudan.

des (Miltha?), Mesalia fasciata, and *Turritella delletrei* plus shark teeth suggest an early Eocene Tethyan (Maastrichtian to early Eocene: Hughes and Filatoff, 1995) setting (see Brown et al., 1989, p. 115–117, and Hughes and Filatoff, 1995, for detailed discussion of the age question). By middle Eocene time, the coastline was too far north to allow these rocks to have been deposited where they occur today unless the long and narrow embayment persisted (note the query in Fig. 12 at the northwestern end of the middle Eocene shoreline). I therefore envisage the Usfan as the more open-marine equivalent of the estuarine Umm Himar and the infrabasaltic sedimentary rocks. Whiteman (1971) once expressed hesitance as to the paleogeographic affiliation of the Eocene marine and estuarine sedimentary rocks of Sudan, but both the nature of the Usfan fossils and the paleogeography of the southern Red Sea–Gulf of Aden area now leave little doubt that they are Tethyan (i.e., that they represent sediments deposited in a gulf opening to the north). This interpretation is corroborated by the interfingering of the marine strata southward with the terrestrial Shumaysi Formation north of Wadi Fatima (Fig. 12; Bohannon et al., 1989). The early Eocene coastline in the middle Red Sea area shown in Figure 12 was drawn with these assumptions and correlations in mind.

The dominance of mudstone, shale, and carbonate rocks in this long and narrow gulf also indicates that the hinterland relief on the Arabian side was probably very subdued, or nonexistent! A similar relationship is encountered in the Jabal Abyad area of north Sudan that marks the terminus of another marine embayment (Figs. 12 and C in 13). At Jabal Abyad (Fig. 12), the Campanian—Maastrichtian Kababish Formation (Klitzsch and Wycisk, 1987) is overlain by the dolomites, limestones, marls, and shales of the Jabal Abyad Formation as redefined and described by Barazi (1985). It contains lamellibranches (*Cardidacea, Corbulidea, Tellindae*), gastropods, echinoderm fragments, teeth of *Stephanodus* (a relative of triggerfishes and filefishes with a range from Late Cretaceous to Eocene), ostracodes, and planktonic as well as benthic foraminifera. Klitzsch and Wycisk (1987) dated the Jabal Abyad as early Tertiary and gave it a range from Paleocene to Eocene, which I accept. The fine-grained clastic rocks occur near the base, and carbonates dominate the upper half, of the section with a total thickness of ~10 m. In overall aspect and in terms of sedimentary environment, the Jabal Abyad Formation is very similar to the Umm Himar and the infrabasaltic sedimentary rocks (Fig. 13C). Like them the Jabal Abyad Formation indicates a hinterland of extremely flat topography. Therefore, if one drew a line across the early Eocene paleogeographic map from Aden to Khartoum and a bit beyond (see Figs. 11 and 12), north of that line would be three northward-jutting peninsulas of extremely low relief that may at times have been temporarily flooded.

Paleogeography south of Afar

The question now is what the relief south of that line was. In southeastern Ethiopia and Somalia, there are reliable shoreline indicators for the early Cenozoic (Bosellini, 1989, and the literature therein; also especially Beydoun, 1964; Hughes and Beydoun, 1992). Between the early and late Eocene, the shoreline in Somalia retreated ~216 km eastward from Hargeysa to Burao (locations shown in Fig. 11; also line A-A′ in Fig. 12). During that time, worldwide sea level has been suggested to have dropped ~50 m (e.g., Haq et al., 1988; Harrison, 1990, Fig. 8.4; Hallam, 1992, p. 124–136, Fig. 6.1; ~40 m: Dewey and Pitman, 1998, Fig. 1; for one test of reliability of the curves used, see Miller and Kent, 1987). This sea-level drop gives a slope of retreat of ~23 cm/km. Willis (1936, p. 31) maintained that for east Africa in general, one could assume that, before late Cenozoic deformation, a slope of 20 cm/km from the region of the rift valleys to the shore was reasonable (which gives a maximum elevation of 120 m above present-day sea level; after eustatic correction, this elevation almost completely disappears, and a near sea-level position is obtained for the area of the East African taphrogen, which is compatible with other lines of evidence, except near the Anza rift in Kenya: see Burke, 1996). Saggerson and Baker (1965, p. 60) also agreed with Willis's estimate. If this scenario is taken as a base to build on, then the early Eocene–late Eocene retreat of the sea in southwestern Ethiopia and Somalia (Fig. 12) must have been purely eustatic. This conclusion indicates an average elevation of ~50 m (an upper bound because the estimate is along the steepest slope) in the region of Hargeysa (which is now at 1500 m above sea level) at the end of the Eocene. If I take the eustatic fluctuations as a guide, I must even conclude that during the later Paleocene, western Ethiopia and Somalia must have risen to allow the early Eocene shoreline to stand to the east of the Paleocene shoreline roughly along the Şan'ā–Dire Dawa line (see the Paleocene shoreline in Fig. 12; compare: Al-Subbary et al., 1998, Fig. C1.7; Bosellini, 1989, Fig. 51; for localities, see Fig. 11). Farther east, to the line of maximum Mesozoic flooding in Ethiopia (Fig. 12), the distance is even larger, and there the relief must have been gentler. Much of Ethiopia, therefore was an area of extreme low relief at the end of the late Eocene, notwithstanding a possible episode of uplift in the Paleocene. Indeed, it was in this very area that Merla and Minucci (1938) identified an end-Cretaceous to ?Eocene erosion surface (*erosione pre-vulcanica,* p. 347).

Farther south and west, one must follow erosion surfaces to reconstruct the morphology and paleoelevations of the land surface in east Africa. In Kenya, the end-Cretaceous erosion bevel stands ~400–500 m higher than the middle Tertiary bevel. This interval decreases to ~400 m near Moyale at the Ethiopia-Kenya frontier and to 300 m near El Wak at the northern bend on the Kenya-Somalia frontier (Saggerson and Baker, 1965). This decrease means that sometime in the early Cenozoic in central and northern Kenya, a northeast-dipping slope originated, which is consistent with the apparent fission-track age of 65 Ma, indicating onset of cooling, and, by inference, relatively rapid denudation beginning at 70 Ma near the Anza rift shoulders to the south and southwest of this surface. By 60 Ma, ~2.6 km of denudation had already taken place here and, at least in places, laterites had covered the erosion level (Foster and Gleadow, 1996).

By contrast, in Uganda, the Buganda (thought to be late Mesozoic on geomorphologic grounds) and the Kasubi (thought to be early Cenozoic on geomorphologic grounds) surfaces are tied to the Atlantic-directed drainage (Bishop and Trendall, 1967), which, across the Congo, created a vast end-Cretaceous to Eocene erosion surface. Cahen (1983) provided a maximum age for this surface in the Congo by dating the Nsele Subgroup as early to middle Cenomanian. This unit is truncated by the erosion surface. The "polymorphous sandstones" (the *Grès polymorphes*) cover the erosion surface and begin with a basal conglomerate, which contains mollusks, ostracods, and plants (*Chara*), but whose age is only presumed to be Eocene–Oligocene (Furon, 1960, p. 306–307; Cahen earlier had only stated early Tertiary as reported by Haughton, 1963, p. 318).

Therefore, it is clear that the land surface rises from both the Atlantic and the Indian Oceans toward a northwest-trending axis (Fig. 12). I do not consider that the elevation of this culmination was ever much higher than 0.5 km, as is clearly seen both in Kenya and in Uganda, where the Buganda surface is never higher than ~300–400 m above the late Cenozoic Kasubi surface (Bishop and Trendall, 1967). What was the cause of this early Cenozoic culmination? Figure 12 shows that right along the axis of the culmination lies one of the grandest taphrogens in Africa, the Central African rift system (see Wycisk et al., 1990, p. 75 and following, and the references therein) connecting with the Anza rift in Kenya (Winn et al., 1993; Bosworth and Morley, 1994; Burke, 1996, p. 373). This system originally formed in the late Mesozoic (numerous phases and loci of rifting from Kimmeridgian to Campanian) but was rejuvenated as a rift system in the late Eocene (Wycisk et al., 1990; Foster and Gleadow, 1996; Burke, 1996). This rejuvenation involved renewed clastic sedimentation in east-central Sudan with massive fanglomerates whose presence suggests that shoulder uplift was starting already in the Paleocene (Wycisk et al., 1990, especially Fig. 13; see also Foster and Gleadow, 1996). In east-central and northwest Sudan, rifting died out completely by the Miocene, and presumably the shoulders lost elevation both by thermal contraction of the underlying mantle cushion and by erosion. The topography would have sloped northeastward and southeastward down to sea level from remnant rift highs of a couple of hundred meters at most, if nothing else had intervened after the Eocene.

Doming around Afar

Did something else intervene after the Eocene to alter the topography northeast of the Central African rift system? If yes, what intervened and when? To answer that question, one must go north, to the lands surrounding Afar. There is here a discrepancy between what most of the geochemists, geophysicists, and tectonic modelers say and what some field geologists claim to be a legitimate inference from field observations. The former maintain that a major dome at least 1000 km in diameter must have preceded the eruption of the massive plateau basalts and the rifting (e.g., Cloos, 1939; Burke and Whiteman, 1973; Burke and Dewey, 1973; White and McKenzie, 1989; but Burke, 1996, has stated that the dome had a diameter of not more than 400 km—see Fig. 11), whereas the latter point to the tropical soils that immediately and abundantly underlie the basalts and argue that here topography could not have had sufficient slopes to justify such a dome: "The laterite of As Sirat and Yemen indicates that this area was near sea level prior to initiation of volcanism" (Coleman, 1993, p. 82). "Regionally, the paleosols represent a widespread and lengthy period of humid conditions in the early Tertiary and developed on a near-sea-level peneplain" (Drury et al., 1994, p. 1373, citing Dainelli, 1943, and Bohannon et al., 1989); "Laterite shows no pre-rifting doming" (Warren Hamilton, 1996, 1998, personal communications).

All authors I have just cited agree that many thick laterites represent soils generated under conditions of prolonged stability and intense weathering, accompanied by essentially no erosion or aggradation (van Houten, 1982) and therefore indicate essentially flat morphology almost at sea level. Even in 1960, however, in his book on the geology of Africa, Furon (1960, p. 88–90) reminded his readers of the complexities of laterites to warn them against too-facile interpretations of its message. It is therefore important to examine what these statements mean in terms of the uplift versus no-uplift scenarios in east Africa and around the Afar triple junction before the statements can be used to support one or the other group of scenarios (also see the discussion in Burke, 1996, p. 374–375 and 377, who pointed to the laterite-free area in Ethiopia to argue for a 400-km-diameter uplift, discussed subsequently).

Lateritization process and rates of uplift. First, in the process of lateritization, thousands of millions of tonnes of Si, Ca, Mg, K, and Na are transported away (into the ocean) as a consequence of bio-rhexistasy (Erhart, 1955, 1956; Burke and Durotoye, 1971, emphasizing the role of earthworms over the termites). Kaolinite and sesquioxides are also transported, but in a much more limited area. Organic materials in the soil must be swept away as well for lateritization to take place, because otherwise they neutralize the positive charges of the oxides (ionized goethite and limonite, for example) and prevent grain growth (Çepel, 1985, p. 204). Burke and Durotoye (1971, and the references on p. 438) emphasized that laterites do not form in tropical forests, but are typical of regions where dry and wet seasons alternate and water-table fluctuations are relatively large. Therefore, a *rigorous washing* of the soils is necessary for lateritization, which requires *runoff* and consequently *slopes*. Neither runoff nor slopes can be so large as to induce widespread erosion, however (cf. King, 1967, p. 177: "weakly-drained land surface"). How steep can the slopes be for the laterites to form and to be preserved? To answer this question, the slopes that today are supporting laterites must be examined; then the question arises of whether similar slopes existed during the formation and later preservation of the laterites around the Afar triple junction.

It was already well known in the 1830s, less than three decades later than the recognition of laterites in 1807 by Buchanan in India, that they form and remain on appreciable slopes and heights: "laterites climb onto the Ghat Mountains as far as the highest points and spread in wide areas as the cover material of the high plains" (Ritter, 1835, p. 702). In Africa (Uganda), Pallister (1954) thought that laterite formed at various times from the early Cenozoic to the Quaternary and extended over an almost continuous surface with dips as much as 10°. Significant patches of laterite have been preserved on much steeper and more rugged topography. For example, a residual laterite surface covers the Alice Tableland in Australia, located at an elevation of ~1000 m above sea level astride the Great Dividing Range in Queensland near the North Queensland hotspot (Fig. 3); the topsoil has been eroded away, and locally a pisolitic laterite and in other places the leached sub-lateritic soil cap the plateau (King, 1967, p. 178). This circumstance means that it is possible to protect large parts of the lateritic cover atop a dome, even after it has reached elevations of ~1000 m; if this dome had fairly uniform slopes of 10° or less, there might be rigorous washing, yet very limited erosion of the laterite cover. In fact, King (1967, p. 177) thought residual-laterite covers on tablelands and hills to constitute a good piece of evidence against the Davisian erosion cycle and for Penck's parallel slope-retreat hypothesis.

In north Africa, the *marram,* a ferricrete, forms a capping on suitable rocks such as shales on highlands between lat 18° and 22°N, especially near the continental divide Ennedi-Erdi-Uweinat (Fig. 12; King, 1967, p. 178). In the Livingstone Mountains (LM in Fig. 12) northeast of Lake Malawi, C. Gillman (cited in Machatschek, 1940, p. 105) had noticed already in the 1920s that an entire soil cover is preserved at ~2900 m elevation! The formation of laterites themselves typically proves an impediment to further erosion, as they commonly form duricrusts: "where laterites cover the ground, an unproductive field forms, which turns stone-hard in the summer" (Ritter, 1835, p. 715). "The largest part of this coastal landscape is covered by the *hard* laterite, the clay-ground" (Ritter, 1835, p. 758, emphasis mine). That is why it is not surprising that laterites resist erosion and denudation.

Today, average structural slopes on the African domes generally do *not* reach 1°, as Table 2 shows. Most of King's cymatogens (i.e., arches or domes) also have flank slopes between 0.5° and 1° (King, 1967, p. 205). Presumably it was with a still flatter surface that they began rising from the early Cenozoic erosion surfaces. If it is assumed that the areas began rising 30 m.y. ago (Burke, 1996), the highest of them, the Afar dome, must have reached its present peak elevation of ~3.5 km (actually 4 km, but ~500 m of it is basalt; WoldeGabriel et al., 1990) at an average rate of ~110 m/m.y. This rate is comparable with other falcogenic (or cymatogenic) rates of uplift from around the world (e.g., Summerfield, 1991, p. 378), and it is about an order of magnitude faster than falcogenic intracratonic basin subsidence as exemplified by subsidence rates of the U.S. interior basins (e.g., Leighton, 1996). One would expect that plume-related domes would rise more slowly if uplift is not fault dominated than if it is (notwithstanding the 9 m/m.y. maximum rate of the Scandinavian glacio-isostatic doming [Ekman, 1989] and the reported [Isachsen, 1975], but disputed [Brown and Reilinger, 1980], ~2.5 m/m.y. tectonic rise of the Adirondack dome in the northeastern United States; Fig. 3). If one assumes that the inverse stratigraphy from the Jurassic upward to the Hercynian basement—preserved in the clasts in the *Küstenkonglomerate* along the southern master faults of the Upper Rhine rift (Fig. 3; Doebl, 1970; Şengör et al., 1978)—indicates the stripping of the 1000-m-thick basement to the Jurassic section (as inferred from the immediately adjoining cover of the Table Jura; e.g., Dorn and Lotze, 1971, Fig. 131; also see Martin, 1976, Fig. 6) along the shoulders of the Upper Rhine rift, and if shoulder uplift kept up with erosion along the faults that formed in the Eocene, one can calculate a rate of fault-dominated uplift of ~200 m/m.y., which is not unreasonable. But if such a rate was maintained since the ca. 30 Ma onset of faulting around the Afar triple junction (Burke's, 1996, preferred time of onset for faulting; but see subsequent discussion), elevations would have now reached ~6000 m at the apex of the dome. If one looks at Quaternary vertical motions with respect to local base levels around the Afar triple junction, the rates range from 0.2 m/m.y. to 0.5 m/m.y. (Faure, 1975, 1976), which are all too fast to produce the present average 3500 m maximum elevation in the past 30 m.y., or even 25 m.y. But this problem of too fast motions is not a local problem around the Afar triple junction. Most currently measured vertical motions in the world seem too fast from the viewpoint of what they might produce in the tectonic environments in which they take place if continued for only a few m.y. and if no erosion occurred. The conclusion is commonly drawn that they thus must be either episodic or oscillatory (e.g., Brown and Reilinger, 1986). In the particular case here under discussion, the rates most likely point to a local episodic nature of faulting.

Because observations on present-day analogous processes do not define what sorts of surface slopes might have come into existence by the time faulting commenced around the Afar triple junction, I fall back on average rates of uplift to see what may have been produced between 30 Ma, when presumably the Afar plume arrived at the base of the lithosphere, and 25 Ma, when it is generally admitted that widespread normal faulting commenced. Clearly, a no-dome scenario is entirely compatible with the inferred rate of uplift along faults since 30 Ma. An average rate of topographic uplift computed since 35 Ma is 110 m/m.y. as already shown, and in 5 m.y., this uplift rate would have produced a dome with an apical elevation of 550 m. The slopes on the dome would depend on the radius of the dome. If I take 300 km as a minimum radius allowed by various authors for plume-generated uplifts, this elevation would give a surface slope of 0.12°! Such a slope seems insufficient to sweep away lateritic duricrusts. In fact, as has been shown, even if slopes reached ~10°, they would still not suffice to get rid of the lateritic cover!

This conclusion is further supported by invoking ordinary rates of denudation. Average rates of denudation were estimated to be ~30 m/m.y. in the Cenozoic on the eastern slopes of the Appalachians (Summerfield, 1991). This is the closest analogous environment and yields an average estimate of the rate of denudation along the early to middle Cenozoic east African margin. The laterite-bearing sedimentary rocks closest to the middle of the presumed dome centered on the Afar are the top parts of the Medj-zir Formation of the Tawilah Group in Yemen (so ~400 km away from the center of the dome as assumed by Burke! See Fig. 11 where the Tawilah Group crops out mostly north of the basalt cover in the Yemen just south of Şan'ā'). These laterites have an average thickness of ~100 m at most outcrops (Al-Subbary et al., 1998) and would thus have required an average time of ~3 m.y. to be denuded completely. If one considers that the onset of volcanism is thought to have been almost simultaneous with doming, the laterites could not have been swept away at all. Now, if one considers that it is duricrusts that are denuded, then the rate of denudation was possibly considerably slower than the 30 m/m.y. average. One would not need to lower the rate even by half to allow the laterites to be preserved until the onset of normal faulting (let us say 5 m.y. after the onset of basalt eruptions) *even if they had no covering flood basalts to protect them.*

Timing of doming. The somewhat long-winded digression concerning laterites was necessary to show the weakness of the often-used laterite argument against a prebasalt and prerifting doming. *From the foregoing, I must conclude that with reasonable rates of deformation and denudation, laterites say nothing for or against doming.* That they have been so considered is largely owing to neglecting the role of duricrusts in landscape evolution especially in the framework of Büdel's etchplanation (see Summerfield, 1991, p. 462, and especially Fig. 18.4). But, as Burke (1996) pointed out, the absence of laterite cover at the apex of the present-day Afar dome in an area that is hundreds of kilometers in diameter requires explanation, which can most comfortably be provided by the presence of a dome before the basalts were erupted. I now look elsewhere to see whether there was such a dome before the onset of rifting around the Afar triple junction.

Figure 14 shows the basalt cover in Ethiopia, in the apical region of the Afar dome, and the stratigraphic levels on which the basalt rests (see Fig. 11 for the larger surroundings). The uppermost Jurassic—lowermost Cretaceous Amba Aradam Formation is the main unit underlying the Oligocene—Miocene basalts. In limited regions, the basalts overlie the Upper Jurassic Antalo Limestone (Juch, 1975; Schubert, 1975; Zanettin and Justin-Visentin, 1975; WoldeGabriel et al., 1990). Data from neighboring Somalia (Bosellini, 1989) and Yemen (Al-Subbary et al., 1998) indicate that the core region of the Afar dome has not been flooded by the sea since the Paleocene. Juch (1975) argued that the age of the Amba Aradam Formation, the so-called "Upper Sandstone," may be as young as Eocene. I interpret, however, that though this unit may be as young as the Paleocene (see the location of the Paleocene shoreline in Fig. 12), an Eocene age is most unlikely (see also Beauchamp, 1977). I infer that the maximum elevation of this region above sea level during the late Eocene was barely 50 m, as previously argued. What is now needed is some indicator of elevation during the Oligocene. There are two pieces of data that may provide this information.

One concerns the Ahwaz Sandstone Member of the Asmari Formation in the Oligocene—Miocene stratigraphy of the Persian Gulf (James and Wynd, 1965, p. 167 and 169, especially Fig. 76; see Fig. 15 herein). This unit consists of calcareous sandstone, sandy limestone, and minor shale. In the unit's type area of the Ahwaz no. 6 well, it is 213.3 m thick. It is considered to be the eastern continuation of the Ghar Formation of Kuwait and eastern Iraq. During the deposition of the lower and middle Asmari carbonates in Khuzestan, sands were being deposited on a delta draining the Arabian Shield (Setudehnia, 1972; Koop and Stoneley, 1982; Jones and Racey, 1994; Alsharhan and Nairn, 1997, p. 460 and 517, and especially Fig. 9.53). The member contains the index taxa of the *Nummulites intermedius—N. vascus,* and *Austrillina howchini—Peneroplis evolutus* zones, indicating an age in the range Oligocene—earliest middle Miocene. The presence of overlying strata of an age in the range latest early Miocene—earliest middle Miocene age would restrict its age, however, to Oligocene—early Miocene, in agreement with previously assigned brackets in the literature (Jones and Racey, 1994). It is underlain in its type area by late Paleocene to early Oligocene Pabdeh shales (see Fig. 76 of James and Wynd, 1965).

The Ghar Formation in Iraq (Owen and Nasr, 1958; Buday and Tyracek, 1980, p. 263–266) and neighboring Kuwait (Milton, 1967) consists of current-bedded, coarse-grained to pebbly sandstones that show a nodular sugary-weathering surface. Some of the sandstones are firmly cemented and calcareous, and a few green clay beds are scattered through the section, although the sands are generally unconsolidated. A few sandy limestones and rare clay and anhydrite streaks occur in the sequence, but these do not form more than 5% of the sequence. In the Bahrah no. 1 well of the Kuwait Oil Company (Fig. 16), the Ghar is seen to be ~240 m thick and directly overlies the lower Eocene-middle Eocene Dammam Formation. In its type locality to the north, in the Zubair no. 3 well, it is only 130 m thick (Fig. 16). Reported thicknesses from two other Iraqi wells are as follows (Owen and Nasr, 1958, Fig. 1; Falcon, 1958, Fig. 6; Fig. 16 here): Nahr Umr no. 1 has 106 m, and Rumalia no. 7 has 165 m. The Ghar has no fossils and is assigned an Oligocene—early Miocene age on the basis of its stratigraphic position between the Dammam and the Lower Fars Formations. Buday and Tyracek (1980) confined the Ghar Formation to the lower Miocene without giving any reasons, despite the rejection of a similar, earlier attempt in 1972 by V. Ditmar and others in

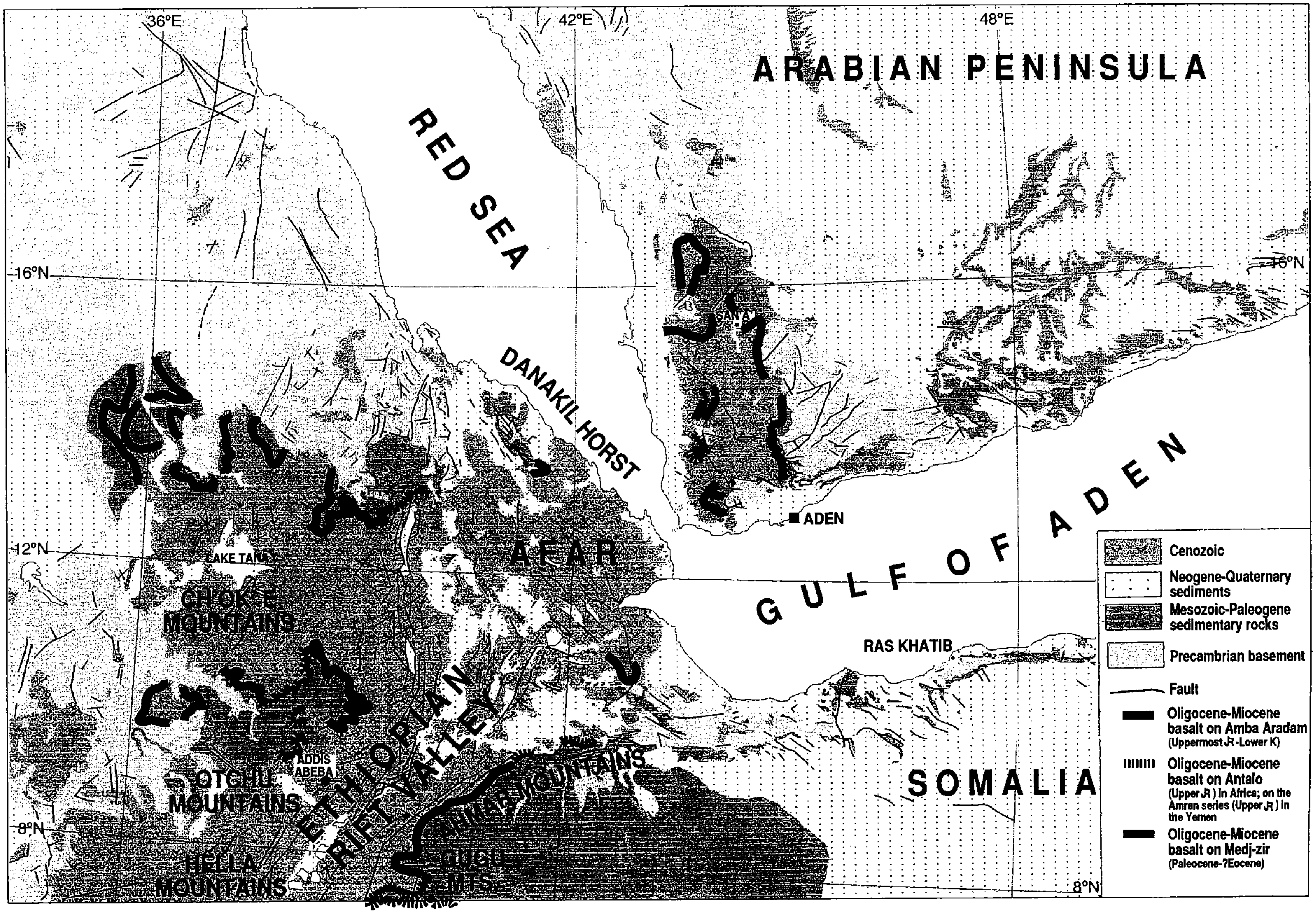

Figure 14. Sketch geologic map of the Afar dome showing the subbasalt unconformity on the Antalo, the Amba Aradam, and the Medj-zir Formations and the Amran Series. Compiled from Juch (1975), Choubert and Faure-Muret (1985), and Al-Subbary et al. (1998).

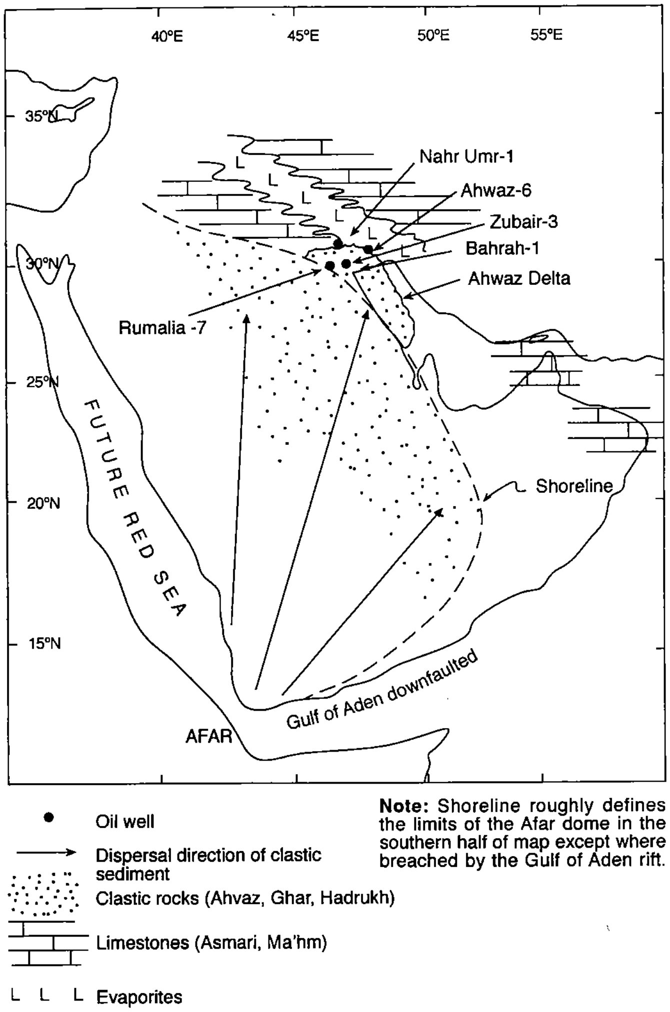

Figure 15. The location of the Ahwaz delta in the Persian Gulf and the Iraqi wells from which the Ghar Formation is known (after Jones and Racey, 1994, 13.14, and Alsharhan and Nairn, 1997, Fig. 9.53).

an unpublished Technoexport report. I follow Ditmar and his collaborators here in accepting the original age assignment of Owen and Nasr (1958) as it is amply corroborated by the correlation with the Ahwaz Sandstone Member. The clastic materials constituting the Ghar Formation in Kuwait were fed from the southwest (Owen and Nasr, 1958; Fig. 16 here), and the thickness variations in the Iraqi wells support this inference.

This sudden appearance of the Ahwaz delta, which was unequivocally fed from the Arabian Shield, i.e., from a place whose elevation above sea level during the Eocene was so little as to induce doubt whether it really formed dry land at all, obviously indicates post-Eocene upheaval. *This upheaval was clearly before the onset of the basaltic volcanism* from the facts that (1) no detritus of basaltic provenance has been reported from the Ahwaz Sandstone Member or its correlatives; (2) the Ahwaz delta deposits consist predominantly of calcareous sandstone, sandy limestone, and minor shale, as previously noted; and (3) its terrestrial correlatives consist predominantly of rounded quartz grains with subordinate limestone pebbles cemented together usually by a poorly sorted, earthy, sandy matrix (Powers et al., 1966, p. D97). The upheaval may even have been before the onset of normal faulting, but because normal faulting in the eastern part of the Gulf of Aden east of Mait (Fig. 12) commenced with the deposition of the Daban series of Oligocene age (Fig. 12; Bosellini, 1989; Bott et al., 1992), it is difficult to know whether upheaval or normal faulting began first. One may approach the problem of the relative age of the normal faulting with respect to the uplift along two indirect paths. One is to question the nature of the early Oligocene faulting. Was it new faulting so as to lead to new shoulder uplift, or was it reactivation of a basin belonging to the late Eocene rifting of the Central African rift system? That the Oligocene clastic rocks seem conformable on the middle Eocene Taleh Evaporites may indicate a basin here before the Oligocene, though none is shown on Bosellini's paleogeographic map of the late Eocene (Bosellini, 1989, Fig. 56).

The other way is to follow the Ahwaz sands to their source. This can be done, though with some difficulty, owing to discontinuous outcrop and lack of characteristic fossils (see especially U.S. Geological Survey and the Arabian American Oil Company, 1963). The Hadrukh Formation is the oldest presumed Miocene unit in Saudi Arabia lying directly and unconformably above the middle Eocene Dammam. The Hadrukh starts with calcareous sands (presumably because it was deposited adjacent to an eroding limestone surface) and upward becomes a sandstone and in places a sandy claystone. It is continental except in an area outlined by Al Qatif, Al ^{c}Alâh, Jabal Qurayn, and Al Jubayl al Barri (Fig. 12). In this area, a few layers near the top of the formation contain poorly preserved marine mollusks (Powers et al., 1966). It is correlated with the Ghar Formation in Iraq and the Ahwaz Sandstone Member of the Asmari Formation in Iran (Jones and Racey, 1994; Alsharhan and Nairn, 1997, p. 460, especially Fig. 9.52). Because only the Ahwaz is bracketed paleontologically by its own fossil content, I am inclined to carry over that bracket to the Ghar Formation (as has been done by many before) and to the Hadrukh as well. That the Hadrukh's lower limit is the middle Eocene Dammam (as is the Ghar's) is certainly compatible with this correlation.

The base of the inland clastic rocks is presumed to be Miocene owing to the original presumptive age assignment to Hadrukh, but there are neither fossils nor bracketing horizons to substantiate this age. The Ahwaz delta shows that its clastic sediments were fed from the Arabian Shield, and the Ghaz and the Hadrukh show how such sequences may pass laterally onto the Arabian Shield continental clastic deposits (U.S. Geological Survey and the Arabian American Oil Company, 1963; Alsharhan and Nairn, 1997, Figs. 9.52 and 9.53). Therefore, some of the so-called continental "Mio-Pliocene" rocks around the Arabian Shield must in fact contain Oligocene clastic sediments also. Parallel with the proposed correlation of the age bracket

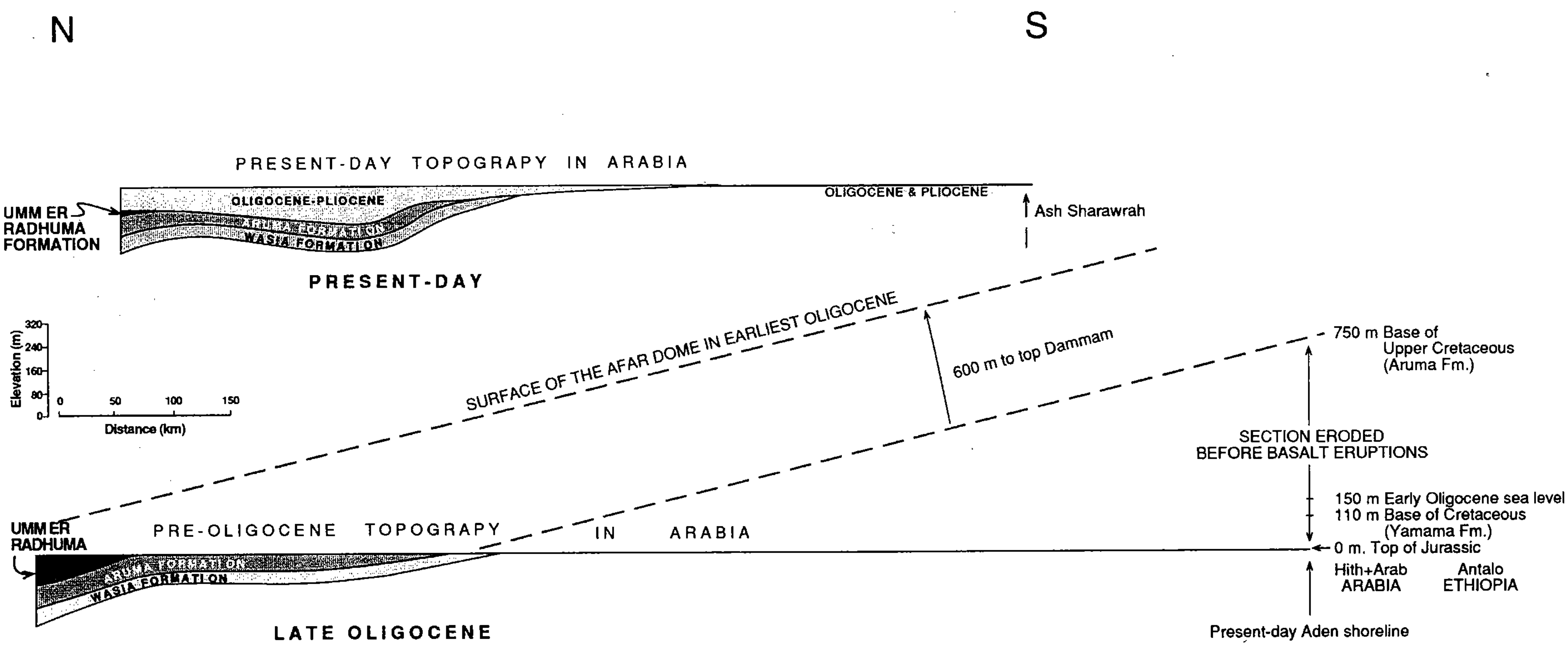

Figure 16. Cross section skirting the Arabian Shield to the east and showing the structural position of the Wasia (Cenomanian-Turonian), the Aruma (Coniacian-Maastrichtian), and the Umm er Radhuma (Paleocene-early Eocene) Formations as their dip projections ascend into the air around Sharawrah. If the dip of the top of Wasia is extrapolated simply to the Yemeni shore along the northern margin of the Gulf of Aden, a height of ~750 m above the present-day sea level is obtained. It is probably that structural level (top Amba Aradam) which now underlies the basalts on the Ethiopian plateau. Because the dip of the top Wasia is now unconformably overlain by Hadrukh (and thus Ahwaz = Ghar; see text) equivalents, the extrapolated elevation probably represents the top Wasia = top Amba Aradam elevation above sea level just before the Oligocene. That means 150 m must be added to the present zero level (Haq et al., 1988). If I now add the ~600 m total thickness of the Umm er Radhuma to the top of the Dammam, the result is $\sim 750 - 150 + 600 = 1200$ m minimum surface uplift.

of the Hadrukh with that of the Ahwaz, one must then take the bottom of the presumptive "Mio-Pliocene" of Powers et al. (1966) as an upper topographic datum to any upheaval that preceded it, but consider its base as Oligocene.

In Saudi Arabia, the southernmost extension of these clastic rocks is in Wadi al Ghirân (Fig. 12), and they are suspected to include Hadrukh equivalents (U.S. Geological Survey and the Arabian American Oil Company, 1963: see text under the signature Tsm). These clastic deposits do not continue into Yemen, where the highest unit beneath the volcanic rocks are laterites and sapropels, indicating prolonged washing and runoff (Geukens, 1966; Grolier and Overstreet, 1978; Al-Subbary et al., 1998). In eastern and northeastern Yemen (the former Aden Protectorate), although the mainly coastal Shir Group (Ghaydah, Hami, and Taqa Formations, consisting of a diverse array of rock types ranging from gypsum and reefal limestone to conglomerates, sandstones, and shales) is approximately age equivalent, tectonically it belongs to the rift trough and not to the shoulder and apron settings of the risen domal structure (Beydoun, 1964, p. 73–75; Beydoun, 1966, p. H37–H39; Greenwood and Bleackley, 1967, p. 53; Hughes and Beydoun, 1992; Bott et al., 1992, especially Fig. 11).

Powers et al. (1966) published an extremely useful longitudinal section essentially skirting the Arabian Shield and ending at Qalamat ash Sharawrah (see Fig. 11 for location). The section shows how the older strata rise and become truncated at an angle of barely 0.1° beneath what I suggest to be Oligocene—Miocene and Pliocene clastic rocks (Fig. 16). If the dip of the top of Wasia Formation (or base of the Aruma Formation) is extrapolated simply to the Yemeni shore along the northern margin of the Gulf of Aden, a height of ~750 m above the present-day sea level is obtained. It is probably that structural level (top of the Amba Aradam Formation) that now mostly underlies the basalts on the Ethiopian plateau. Because the dip surface of the top of the Wasia is now unconformably overlain by clastic deposits that are in part Hadrukh (and thus Ahwaz = Ghar; see previous discussion) equivalents, the extrapolated elevation probably represents the top Wasia = top Amba Aradam elevation above sea level just before a certain Oligocene possibly younger than the Pabdeh shales. That age assignment

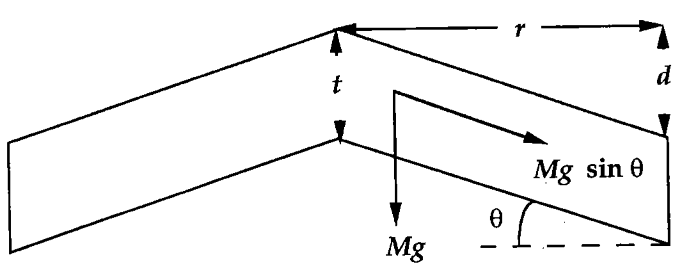

Figure 17. Cross section across a conical hotspot uplift showing the parameters used in calculating the stresses generated by the gravitational potential of the uplifted crust. See text for discussion.

means that 150 m (± 50 m) must be subtracted from the present elevation owing to sea-level lowering since then (Haq et al., 1988). If I now add the ~600 m cumulative thickness of Umm er Radhuma to the top of the Dammam, I get ~750 − 150 + 600 = 1200 m minimum surface uplift. This uplift, which should perhaps be thought of as being ~1 km given the source of errors in the reasoning, is entirely comparable with that of the present *unrifted* African cupolas (hotspots and high spots, Table 4: Thiessen et al., 1979; see especially Burke, 1996, Fig. 6, for the high spots). The uplift formed clearly before the onset of basaltic volcanism at ca. 30 Ma ($^{40}Ar/^{39}Ar$ ages cluster at 31–28 Ma: Hofmann et al., 1997; for a 40–35 Ma population in Ethiopia south of the main traps, see George and Rogers, 1997, but those older volcanic rocks clearly belong to an older episode of rifting in the Lake Rudolf area, more related to the Sudanese rifts than to the Afar dome: see Burke, 1996, p. 375; see also White and McKenzie, 1989, and Burke, 1996, for further references to the literature on the Yemeni-Ethiopian basaltic volcanism).

Watson and McKenzie (1991) showed that the amount of melt produced in a plume is related to the thickness of the mechanical boundary layer above it. They modeled this relationship by changing the pressure in the melting region. Though not a rigorous method, it does give an indication of what would happen. I assume here that the mantle and crustal lithosphere remaining on top of a plume following detachment of a part of the mantle lithosphere is equivalent to the mechanical boundary layer. The elevation of ~1200 m that the Afar dome attained before the onset of voluminous magmatism (as already described) indicates that almost none of the mantle lithosphere had remained below if the initial crustal thickness was closer to 30 than to 40 km (Lachenbruch and Morgan, 1990; cf. Fig. 8). A crust thinner than 40 km is likely owing to the Early Cretaceous extension. If the crust was ~35 km thick, one would expect a melt-production rate of ~0.5 km^3/yr above the plume. Yet even if all the ~400 000 km^3 of melt production (cf. White and McKenzie, 1989) is restricted to the period of peak effusive activity between 31 and 28 Ma (Hofmann et al., 1997), a melt-production rate of at most 0.1 km^3/yr is determined. But if this were all the melt produced, it would correspond with a mechanical boundary layer thickness of ~80 km or more (Watson and McKenzie, 1991, Fig. 16), and that thickness would give an elevation of ~1000 m if a temperature of 1000°C and a plume-head density of 3.22 g/cm^3 are assumed (Lachenbruch and Morgan, 1990).

This elevation is clearly within the limits of error of my method of measuring it during the latest Eocene. But if the melt effusion rate at the surface was lower (which is likely), then a method of producing more magma and then hiding it must be found. I think a solution is to be found in magmatic underplating (McKenzie, 1984). If, by the time of the eruption of the voluminous Ethiopian-Yemeni plateau basalts, the lithosphere had indeed been almost completely robbed of its mantle part, a ~9-km-thick layer must have been magmatically underplated below the Afar dome, if it is assumed that the layer's area is coextensive with the surface basalts (by using the approach of Watson and McKenzie, 1991, Fig. 16). This is a small amount of melt when compared with much thicker igneous sections underplated at volcanic Atlantic-type continental margins (White et al., 1987; White, 1989), but it would still produce an ~1 km surface uplift (assuming a density of 3 g/cm^3 for basalt), which would almost account for the present average height of the Afar dome! It would also require an asthenospheric potential temperature clearly closer to 1400 °C than to 1200 °C (cf. White et al., 1987, Fig. 5; White and McKenzie, 1989, Fig. 3; see also Watson and McKenzie, 1991, Fig. 7), which is compatible with the presence of a mantle plume under the Afar dome.

It is less easy to say anything about the relative timing of rifting and uplift within the Gulf of Aden, because both events happened simultaneously with respect to the resolution of the available data. The earliest rift-related sedimentary rocks are clearly early Oligocene (Ghaydah Formation in the Yemeni side: Hughes and Beydoun, 1992; Daban Series on the Somalian side: Bosellini, 1989, and the references therein; see Fig. 11), but so may be the earliest sands deposited in the Ahwaz delta (Jones and Racey, 1994). I can only question, on the basis of the extremely eccentric position of the Ghaydah and the Daban rocks with respect to the Afar dome (Fig. 11; according to Burke, 1996, they lie completely outside the area of the dome), whether the rifting they record was related to the tectonics of the dome.

The history of rifting in the Red Sea may assist in solving this quandary. It is clear that there was uplift in the future Red Sea area in the Oligocene. The middle Eocene to Oligocene Dabaa shales (which are conformable on the Eocene Apollonia Formation: Hantar, 1990; Said, 1990) in Egypt thicken away from the Red Sea shoulders and the coarser clastic Mamura (late Oligocene) goes from zero thickness near the Red Sea around East Mubarak to ~300 m westward in the Dabaa area (Said, 1990, especially Fig. 24.5). By contrast, no downfaulting took place until the later late Oligocene (Savoyat et al., 1989; Hughes and Beydoun, 1992; Abou Ouf and Gheith, 1998; Plaziat et al., 1998). Owing to poor dating of the Oligocene sequences of the Red Sea margins, no diachroneity has been recognized along its length. Omar and Steckler (1995) have shown by fission-track analyses that the Red Sea in fact opened at once along its entire length, although at its northern end, no preparatory uplift has been recognized in contrast to its southern end (Garfunkel, 1988). In the Ethiopian Rift Valley, WoldeGabriel et al. (1990) pointed out that in the Ethiopian rift, the earliest phase of normal faulting also commenced in the late Oligocene, possibly coevally with the rifting along the Red Sea margins. Though WoldeGabriel et al. (1990) also cited thinning of the Mesozoic section beneath the plateau basalts as evidence for prerifting doming, it is clear that the observed thinning was a consequence of pre-Aptian tectonism that affected northeastern Africa and the Arabian platform in latest Jurassic-Early Cretaceous time (for a general reference, see Beydoun, 1988, p. 52–53; for particular regions, see Bender, 1974: pre-Aptian-Albian tectonism affecting Jurassic rocks in Jordan, creating an angular unconformity; Wolfart, 1967: unconformity below Aptian rocks in western, northwestern, and northern Syria all the way into southeastern Turkey; Wolfart, 1967, Lang and Mimran, 1985, Garfunkel and Derin, 1988, Garfunkel, 1989: extensional [strike-slip-related?] tectonism and volcanism along the coastal regions of Syria, Lebanon, and Israel; El Ramly and Hussein, 1985, and Vail, 1985: latest Jurassic-Early Cretaceous ring complexes in Egypt and in the Sudan; Bosellini, 1989: deformation and uplift in Somalia). The observed thinning had nothing to do with the activity of the Afar plume.

The preceding discussion shows that a *probably* pre-a certain Oligocene to post-middle Eocene uplift affected the region of the future Afar dome. The amplitude of doming was ~1 km about halfway down its eastern slope and thus may have exceeded 1 km at its apex. This doming was succeeded by normal faulting in the Red Sea area and in the Ethiopian rift valley. Only in the Gulf of Aden is the question of primacy of doming or faulting undecided. Early Oligocene faulting in the *eastern* Gulf of Aden region, however, looks as if it was not related to the doming, but represented a propagation from the Carlsberg Ridge in the direction of the apex of the Afar dome as, and for the reasons, suggested by Manighetti et al. (1997, esp. Fig. 11).

GREGORY RIFT VALLEY

The purpose of this brief section is only to show that the Gregory Rift Valley in Kenya and northeastern Tanzania also followed the formation of a dome as in the Afar area, although the dimensions of the Kenyan dome were smaller than the Afar dome and resembled more the other Neogene lithospheric domes in Africa (Burke, 1996). The development of the Kenyan dome and the subsequent rifting mimicked in detail the evolution of the corresponding features in the Afar dome region.

The rift arm extending down from Lake Rudolf through the Gregory rift (Fig. 12) is well studied with respect to relative timing of doming and rifting (see Smith, 1994, for a review; also Burke, 1996). Saggerson and Baker's (1965) careful study of the erosion surfaces in Kenya document that the sub-Miocene surface rose to ~300 to 500 m elevation near the future rift valley before the rise of the end-Tertiary surface commenced, i.e., probably before rifting and attendant shoulder uplift began. Saggerson and Baker noted that the initial faulting began after the warping of the sub-Miocene bevel. The warping of this sub-Miocene bevel is not seen in the recent fission-track analyses (Foster and Gleadow, 1996), although there is some geologic evidence for it, such as the infilling of the late Oligocene-early Miocene rifts in the Lake Rudolf area by detritus derived from the south (Morley et al., 1992). Paleodrainage also indicates the presence of a high in the future Gregory rift region during the Paleogene (see Smith, 1994, p. 7 for references). The discrepancy between the geomorphology and geochronology probably results from the extremely low rates of denudation over a very flat and broad domal surface covered by hard-to-erode laterite, similar to the situation over the Afar dome. (Foster and Gleadow, 1996) pointed out that even the Miocene to Holocene rates of denudation in Kenya are very low, having nowhere accomplished more than ~50 °C cooling, and they ascribed this effect to the dry climate. Because the dry climate is a result of uplift, it is likely that as the Kenya dome rose, it has become progressively harder to denude. Modeling of the age and track-length data from samples that yield ca. 65 Ma ages suggests that regional denudation rates increased again during the Miocene. I suggest that this increase happened because fault-related uplift rates are faster than those that are purely falcogenic, as already emphasized in relationship to the Afar dome, and fault-related elevations have much steeper slopes conducive to rapid erosion. Smith (1994) concluded that a dome of less than 1 km elevation formed before the rift faulting commenced in the Gregory rift. He placed its origin between 40 and 21 Ma.

An extremely important observation by Saggerson and Baker (1965) is that the doming was probably contemporaneous with the development of an *apical depression* that did not exhibit major faulting, but was more like a very broad synclinal trough. Smith (1994) corroborated this inference and showed the presence of a sag basin both in the Lake Rudolf area and in the region of the Elgeyo Escarpment in the northern part of the Gregory rift. As the dome rose, the trough was progressively converted into a rift valley as Williams and Chapman (1986, Fig. 2B) illustrated.

In the Gregory rift, both faulting and volcanicity marched from north to south. Faulting started in the Lake Rudolf area even earlier than the 30 Ma onset of volcanicity related to the current episode of rifting (Williams and Chapman, 1986; Smith, 1994), which is consistent with Watkins' inference that the Lake Rudolf basin may be significantly older than the Koobi Fora section in the northeastern shoulder of the lake near the Ethiopian frontier, the oldest tuff bed of which was dated at 4.5 Ma (Watkins, 1986). Baker (1986) argued that faulting started in the early middle Miocene (ca. 15 Ma) in the Gregory rift and was almost coeval with volcanicity, although Smith (1994) stated that major faulting commenced at 10 Ma. All of this timing is compatible with Baker and Wohlenberg's (1971) result and Smith's (1994) corroborative conclusion that the available evidence indicates the presence of an area of <~1 km uplift in the area of the future Gregory rift before the commencement of rift faulting as well as the north to south progression of both faulting and volcanicity.

MECHANISM OF RIFTING

The preceding detailed discussion showed that in the rifts radiating away from the Afar center and in those bisecting[8] the Kenya dome farther south, rift faulting most likely was preceded by uplift of an amplitude of ~1 km. In both cases, volcanicity succeeded some faulting since the prevolcanism sag basins were presumably also fault controlled (see Smith, 1994, Fig. 3). The sequence of events in these cases was therefore doming-faulting-volcanicity, which is in complete agreement with Cloos's (1939) original suggestion formulated as *Hebung-Spaltung-Vulkanismus.* This sequence modifies Burke's view, which was doming-volcanicity-rifting (Burke and Whiteman, 1973; Burke and Dewey, 1973). Burke's prime examples of riftless, volcano-crowned domes were the Saharan domes. Thus I next take a fleeting look at the geology of at least two of them.

Rifting and volcanicity on the Ahaggar and the Tibesti domes

Perhaps the best studied of the Saharan domes is the large Ahaggar massif straddling the Algerian-Niger frontier in the central Sahara (Fig. 3). The massif consists of annular bands of Mesozoic and Paleozoic rocks around a basement core of Pan-African age. Cenozoic deposits are more irregularly disposed, older ones being toward the periphery and the youngest Pliocene–Quaternary sedimentary rocks in small depressions in various parts of the massif. Elevations in the massif are typically between 1500 and 2000 m; the maximum elevation is 2918 m.

The Pan-African basement has a pronounced north-trending fabric and consists predominantly of variously metamorphosed subduction-accretion material with arc magmatic rocks and later granites making up ~50% of the exposed area (Bertrand and Caby, 1978; Bertrand et al., 1986). Large shear zones slice through the Pan-African collage and represent the final stages of what I interpret as a Turkic-type orogeny reminiscent of the Altaids of central Asia (Şengör and Natal'in, 1996a, 1996b). Following a Cambrian peneplanation, an Ordovician-Carboniferous Gondwana-Land* succession was laid down, but it was largely swept away during a late Paleozoic upheaval, the effects of which may have lasted into the Jurassic. This sequence of events is corroborated by the surfacing and destruction of Paleozoic oil reported by Boote et al. (1998, p. 64). Deposition of Cretaceous terrestrial sediments—and marine sediments in deeper depressions associated with the central African rift system—was in places accompanied by tholeiitic to subalkalic basalts. The volcanism may have lasted into the Eocene (one ring complex was dated by K-Ar method into the Eocene). Thick basin fills of Eocene-Oligocene age indicate rifting in the Paleogene. Finally, volcanism greatly resembling that along the East African rift system and interpreted to be a product of crustal stretching commenced in the Miocene and remains active.

Dautria and Lesquer (1989), from whom I have summarized the preceding brief account, have shown that there was

[8] The Kavirondo rift is both much smaller and younger than the Gregory rift (Shackleton, 1951; Jones and Lippard, 1979). It is unclear to me whether the Kavirondo rift has any significant extension across it. That is why here I do not follow Burke and Whiteman's (1973) rift star model for the Kenya dome.

*The author insists on this spelling of the term. GSA publications normally use Gondwana.

normal faulting in the Ahaggar at least in the Paleogene, most faults following trends established in the Cretaceous. These normal-slip fault systems form a three-armed protorift star, and their bounding faults have been claimed to have been intermittently active through much of the Cenozoic. The Pan-African basement fabric of the Ahaggar has proved very susceptible to reactivation, especially its late strike-slip fault systems. The important thing I wish to emphasize here is that the volcanicity on one of Burke's prime examples of a "riftless" dome has in fact occurred in the wake of rifting, if not concurrently with it. Dautria and Lesquer (1989) interpret that the volcanicity *always* followed the faulting. I should have followed them with no reservations had their descriptions been a bit more precise.

Their observations on the ultramafic nodules in the Pliocene-Quaternary basalts (but not in the Miocene ones) are very detailed and very worthy of note: The nodules resemble those from many zones of acknowledged lithospheric extension in the world and display significant deformation. Those samples from the western Ahaggar (Tahalra and Manzaz) are less depleted than those of eastern Ahaggar (Eggere and Adrar N'Ajjer). Dautria and Lesquer (1989) ascribed this difference to Cretaceous partial melting, because the Miocene lavas have a very narrow range of composition and represent a small (6%) degree of partial melting. Dautria and Lesquer (1989) concluded that the nodules represent a mantle underlying rifts that are actively extending and probably reactivating older shear zones of lithospheric penetration.

In Tibesti, Vincent (1970) also showed that volcanism followed an episode of normal faulting that rejuvenated a set of faults that had formed between the Cretaceous deposition of the Nubian sandstone and the Eocene transgression. The rejuvenated faults tilted the northeastern part of the Tibesti massif in the same direction and "the center of vulcanism is at the steepest corner of the block and it is clear that the two facts are related" (Vincent, 1970, p. 304).

Prerift sags upon domes

In all cases reviewed in this paper, faulting always preceded volcanicity, and both succeeded a phase of uplift. But also in all cases, doming was accompanied by a local depression with little faulting, whose area is always larger than the final rift trough. The Umm Himar and the Usfan Formations in Saudi Arabia and the estuarine infrabasaltic sedimentary rocks at Jabal Oharba in Sudan compelled me to draw a large, south-directed gulf along the axis of, but wider than, the future Red Sea. The age of this gulf could have extended even into the early Oligocene. In the Gulf of Aden, the earliest Oligocene Daban Series, resting on a pan of evaporites mimicking the distribution of the middle Eocene Taleh Gulf and only a little displaced northward with respect to the late Eocene Karkar Gulf (Bosellini, 1989), is not obviously related to the later, narrower rift trough, for it rests on older successions with apparent conformity cf. (Bosellini, 1989, Fig. 39). In east Africa, there is clear evidence of an axial depression just predating, or concurrent with, the doming (Williams and Chapman, 1986; Dawson, 1992; Smith, 1994). In Ethiopia, WoldeGabriel et al. (1990) documented a similar timing by mapping basalt distribution and the different episodes of normal faulting and depression formation (see their Fig. 11). In the Ahaggar, the Amadror and Tafassaset Oued depressions form apical troughs that are superimposed on older Cretaceous rifts (Dautria and Lesquer, 1989). Similar observations have been reported from many rift basins that overlie domes. In western and northwestern Australia, Falvey and Mutter (1981) recognized an "infrarift" stage of broad downwarping before the rise of hotspot domes, an interpretation followed by Veevers (1984, p. 190–191). An explanation is required for the facts that (1) those rifts overlying mantle plumes, which have been inferred from independent data, open following formation of a dome with an average radius of <1000 km and a height of ~1 km and (2) this episode of doming is commonly accompanied by the formation of a shallow sedimentary basin with little faulting lying along the axis of the future rift.

Figure 8 shows how the syndoming trough and the dome might form as already discussed. The sequence of events illustrated there is reminiscent of the lithosphere-detachment mechanism proposed by Houseman et al. (1981), but differs from it by introducing an anomalously low density material (i.e., the plume head) below the part of the lithosphere to be detached, rather than extending that part itself into the asthenosphere by convergence-related thickening. The sag basins may be one test to distinguish the model developed here for lithospheric detachment above plume heads from that proposed by Yuen and Fleitout (1985) and Dalloubeix and Fleitout (1989), which does not require them. This sort of sagging also provides one possible explanation, distinct from the model of Manighetti et al. (1997), of the "anomalously early" rifting of the eastern part of the Gulf of Aden and the other, nearly coeval sag basins that formed in Ethiopia and in the Red Sea.

Gravity-induced rifting

But it is also clear that no amount of doming within reasonable limits allowed by observations is sufficient to create anything like the east African rift troughs by the extension resulting from areal increase consequent upon doming. The Cloos (1939) model and, following this, the Burke model (Burke and Whiteman, 1973; Burke and Dewey, 1973) do not work, and an alternative explanation is needed for the close association of doming and rifting. One possibility is to consider gravitational stresses generated by the topographic potential of the dome, as originally suggested by Bott and Kusznir (1979) and applied by many others since (Artyushkov, 1981; Neugebauer and Temme, 1981; Houseman and England, 1983). The influence of gravity on a density-layered outer shell is complex, but I here take a simple-minded end-member case, that of gravity sliding, to see what sort of magnitudes of stress may be in-

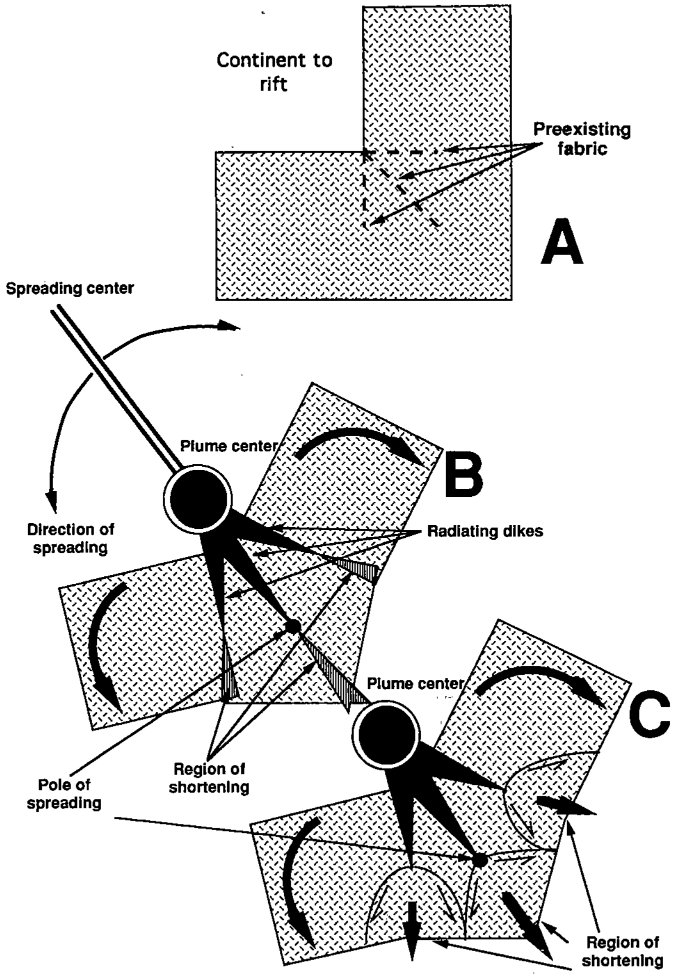

Figure 18. A, Model of a continent with an indented continental margin similar to the present Gulf of Guinea or the Laptev Sea in northeast Asia and a preexisting fabric forming weak zones as illustrated. B, If a spreading center forms at the reentrant (as happened in the Benue Trough in the middle Cretaceous: Burke et al., 1972; Burke and Dewey, 1974) or propagates into it (as in the present Laptev Sea: Fujita and Cook, 1990), the extensional deformation is likely to become diffused along the preexisting weak zones as it did in Cretaceous Africa, and compression must result across the poles of rotation of the spreading center and the individual loci of extension. This extension may take the form of compressional structures such as thrust faults and folds. C, Alternatively, compression within the continent may be accommodated by strike-slip faulting that would also follow old lines of weakness.

volved if only topography were governing the deformation. If I take a conical uplift with an elevation $d = 1$ km and a basal radius $r = 1000$ km (Fig. 17), then the stress σ generated by the gravitational potential along a one-half cross section (owing to symmetry, only one half needs consideration) can be calculated as $\sigma = Mg \sin \theta$, θ being the angle of elevation of the cone, M the mass of the crustal slab to slide downhill, and g the acceleration due to gravity (Fig. 17). $M = rt\rho$, where t is the thickness of the slab to slide and ρ is the density (Fig. 17). Because $\sin \theta \sim \theta \sim \tan \theta \sim d/r$, σ can be expressed as ρdgt. After substituting $\rho = 3.3 \times 10^3$ kg/m^3, $d = 10^3$ m, $g \sim 10$ m/s^2 and $t = 40$, the result is $\sigma = 1320$ bars. If I use only the continental density of 2.8×10^3 kg/cm^3, this result reduces to 1120 bars. This value is similar in magnitude to that which Bott and Kusznir (1979) obtained for a 2 km uplift, which was 2 kbar. Now Bott and Mithen (1983) have shown that to generate a 40-km-wide and 5-km-deep, sediment-filled rift trough, tensile stresses on the order of 1 to 2 kbar are necessary. Clearly both Bott and Kusznir's 2 kbar and my ~1 kbar are just sufficient to break a continental lithosphere if the brittle thickness is 40 km (~30 km in the southern part of the East African rift system: Jackson and Belkinsop, 1993; ~40 km at Lake Baykal: Diament and Kogan, 1990). However, McKenzie and Fairhead (1997) have shown that the elastic thickness is commonly considerably less than the thickness of the brittle seismogenic layer. They also showed that if free-air anomalies are used that fit with the topography, in the places they had data (North America, Australia, east Africa, India, the former USSR), the elastic thickness nowhere exceeds 25 km! In east Africa, they found the elastic thickness to be only 6 km, with compensation occurring at a depth of 8 km. Though these values refer to the present-day, if doming is a result of the reduction of lithospheric thickness and the consequent elevation of heat flow, the elastic lithosphere just before rifting was probably thin enough to concentrate the stresses and accomplish failure (cf. Bott and Kusznir, 1979; Bott, 1982a). This may also be the reason why on many domes, the first volcanism is extensive rhyolitic ignimbrites and granitic ring complexes (e.g., Kampunzu and Mohr, 1991, Mohr, 1992), as the lower boundary of the mechanical layer retreats upward into the continental crust atop the recently bared asthenospheric mantle-continental crust boundary. As elevation (d) increases, tensile stresses increase proportionally. High domes with thin elastic layers are thus prime regions for rifting (Bott, 1982a). That is why the high domes in Africa are the ones that are fractured by rifts, but the lower ones display little evidence for rifting. This argument also makes dike injection into domes possible by opening dike fractures through gravity. Once a high hydrostatic head is attained, it would then be easy to propagate long dikes away from a common center fed by a voluminous-magma-producing plume head. In the Mackenzie giant radiating-dike swarm, for example, magma was injected vertically within 500 km of the axis of the feeding plume and traveled horizontally for at least 2100 km outside that region (Ernst and Baragar, 1992). This inference from a fossil example is compatible with a 1200-km-diameter active plume head inferred in Hawaii (Watson and McKenzie, 1991). Basal drag away from spreading plume centers must certainly contribute to the disintegration of plume-related domes (e.g., Westaway, 1993), and the fact that thick lithosphere with not much doming also breaks by normal faulting (Déverchère et al., 1991; Jackson and Belkinsop, 1993) clearly shows that gravity alone may not always be enough.

If swell shells slide off uplifts, where do they slide to? Upper continental crustal rocks can generally withstand compressional stresses of ~5 kbar (e.g., Bott, 1982b); in shield areas such as Canada or in cratonic regions such as North America, this strength is greater and is applicable to depths greater than just 20 km (see Zoback and Zoback, 1997; Şengör, 1999, Fig. 11). Because the peripheral areas of plume uplifts are not necessarily thermally weakened, it is unlikely that annular thrust belts or fold ridges around them will develop. In fact, no such thrust belts exist on terrestrial plume highlands now. They do not seem to have existed in the past either. None of the African or Australian domes mentioned in this paper (cf. Table 2) has associated peripheral thrust belts. Those that are known from the past, for example, the Atlantic domes (e.g., in northwestern Iberia: Wilson, 1975; Argentina: Burke, 1976; Uliana and Biddle, 1987; Uliana et al., 1989; see also Şengör, 1995) or the much older Scandinavian domes (Kumpulainen and Nystuen, 1985, especially Fig. 4) are devoid of thrust garlands around them. As McKenzie (1994) pointed out, on Earth, rift extension is taken up along plate boundaries. Rifting in east Africa must be compensated along the global plate-boundary network, as must be that along Lake Baykal or along the Upper and the Lower Rhine Valleys. The "rigid" opening of the Red Sea (Omar and Steckler, 1995) further corroborates this.

By contrast, on Venus it is suspected that hotspot plateaus generate strains that are locally compensated by circular to elliptical thrust belts that appear as the coronae. McKenzie (1994) showed that the absence of correlation between gravity anomalies from plumes and the coronae is consistent with the idea that coronae mark places where former plumes existed and now are recognized only by the presence of their peripheral thrust belts. The absence of coronae on Earth (except on a very small scale partially around large individual volcanic edifices, such as Merapi, Merbabu, and Soropati-Telemojo in Java [gravity-slide model]—van Bemmelen, 1949, p. 560–565; Hawaii [gravity-push model]: Dieterich, 1988) further corroborates the view that plume-related uplifts do not generate thrusts and folds in the immediate periphery of the plume highland. This statement further strengthens the case for the existence and local preponderance of nongravity components in rift generation on Earth.

Giant radiating-dike swarms and plume-related uplifts: A genetic connection?

If all rifting is integrated with plate-boundary phenomena, so must be the generation of large radiating-dike swarms (Ernst et al., 1995a, 1995b, 1996; Ernst and Buchan, 1997; Buchan et al., 1998; Ernst and Buchan, this volume). The large amounts of magma involved clearly show that their initiation is a plume-related phenomenon, but the amounts of total extension observed (e.g., in the Mackenzie swarm, the estimated aggregate dike thickness 400 km away from the radiation center and along 180° (arc length is 3.06 km: Baragar et al., 1996) and the minimum related uplift (calculated by Baragar et al. with a method equivalent to Élie de Beaumont's, namely, 17.5 km!) are excessive (similar to the case in most rifts as already shown herein).[9] Given the regular pattern of the dike swarms about an apparent focus gravity will not do either. A radiating-dike swarm similar to a rift star, such as the one illustrated in Ernst and Buchan (1997, Fig. 10 III: "subswarms of subparallel dikes which radiated from a common point"), may be explained by gravity gliding of crustal panels, but not a dense swarm such as the Mackenzie (Baragar et al., 1996) corresponding with the "continuous fanning pattern" of Ernst and Buchan (1997, Fig. 10I).

Instead, one may initiate a dike swarm about a plume head, which then leads to continental separation with a geometry resembling that in Figure 18A. As the spreading center then attempts to propagate inland, it may branch out following the older fabric within the continent and give rise to dike swarms. This process is in many aspects what did happen in north-central Africa as rifting propagated (instantaneously within the resolution of the stratigraphic and isotope chronological data as in the case of the Red Sea: see previously described timing from the opening Benue Trough into the area of the Chad basin and beyond into the central Sahara, cf. Burke and Dewey, 1974; Burke, 1976; Pindell and Dewey, 1982; Fairhead, 1988). This process is also what seems to be now happening in the case of the propagation of the Nansen (or the Gakkel) Ridge into the body of Asia in the Laptev Sea and leading to the formation of the Moma-Zoryansk rift system and the Balagan-Tas alkalic volcano[10] (Fujita and Cook, 1990). Two end-member possibilities (Fig. 18, B and C) could accommodate the compressional strain that would form as a consequence of the rifting and/or dike propagation in the manner described. This manner of extension would enable the dike focus to be tied to a plate-boundary system so as to make continuous extension possible and would provide a realistic analogue with the only active large-scale radial dike system in the world, namely, the Red Sea-Gulf of Aden-East African rift system. Not only have the Gulf of Aden and the Red Sea branches developed into fully operational plate boundaries, but even the East African rift branch has a nascent spreading center in the Wonji fault belt in the middle of the Ethiopian rift valley; that center must be propagating southward now. Though Richard Ernst (1999, personal com-

[9] However, had Baragar et al. (1996) used a more realistic spherical-cap model on a spherical Earth, the uplift might have come out to be not so wholly unreasonable: I redid their calculation by using the spherical-cap assumption, plus an uplift radius of 900 km and an average extension of 1.6 km (their estimated aggregate dike thickness along 392 km of arc at that distance from the radiation center) uniformly distributed for the entire 900 km diameter, and I obtained an uplift of 3.6 km. It is still somewhat excessive but not entirely unreasonable. So Baragar et al.'s pessimistic conclusion that their "assumption of a simple, centrally based uplift is not valid" may not be completely justified. However, it should be borne in mind that I used all of their minimal assumptions in the equation $h' = \delta Aa/\pi r^2$. The required uplift may still remain excessive when values close to average magnitudes for the involved variables are used.

[10] Some of the Nansen Ridge extension seems to be transferred sideways to the east along a strike-slip fault. The plate-boundary geometry where the Nansen Ridge apparently propagates into the continent is more complicated than the schemas illustrated in Figure 18, B and C, indicate (B. A. Natal'in, 2000, personal communication).

munication; also see Rainbird and Ernst, this volume) has suggested that the dike systems of the Central Atlantic magmatic province, which all formed within a few million years of each other (Marzoli et al., 1999), constitute a complete, 360°, radial giant dike swarm, I remain unconvinced that all of these dikes radiated from a single center somewhere in Florida. Judging from the small-scale map in Marzoli et al. (1999, Fig. 1), at least four centers can be identified, although it is entirely conceivable that some of the dikes may have formed independently of any mantle plumes. Bertrand and Westphal's (1977, Fig. 1) larger-scale map corroborates these impressions. The more schematic maps in Deckart et al. (1997) and Ernst and Buchan (1997, Fig. 1, and this volume) look more convincing, but a detailed examination leads to mounting suspicion. This suspicion is strengthened by the apparent absence of a large magmatic province in the proposed focal region (Hill, 1991). I think it is premature to claim the Central Atlantic magmatic province for any model, before adequate amounts of reliable structural data have been obtained. Looking at the marvelous map of Ernst et al. (1996) showing the world distribution of giant dike swarms, I see no other 360° radiating-dike swarm on planet Earth.

Rift propagation and the "steering" of propagating rift tips

This discussion of incorporating any extensional structure that has been stretched for more than 10 km into the Earth's plate-boundary system and the role of propagating spreading centers and rifts brings me to the question of the steering of the propagating rift tips. It is clear that the Carlsberg Ridge in the Indian Ocean somehow "sniffed" the presence of the Afar plume and steered its propagating tip in that direction (see especially Fig. 37 in Burke, 1996 and Manighetti et al. 1997, Fig. 11). In Africa, rifting seems to have propagated from north to south, generally from dome to dome, and is still doing so (see Şengör, 1995, Fig. 2.7). It not only seeks out zones of preexisting weakness, but mainly also zones of gravitational potential that could be exploited, although the Malawi rift seems to have avoided the Katanga dome (Burke, 1996, Fig. 6). I think a chain of broad domes may steer a propagating rift tip along itself owing to their gravitational potential, as if it were "handing the propagating fractures from one dome to the next," provided the crustal zones of weakness and the overall global plate kinematics are compatible with it. This process is precisely what seems to have happened in Africa between the mouth of the Gulf of Aden and Tanzania and beyond. Crustal factors clearly weighed more heavily in the choice of the Red Sea course—the Sinai corner stress concentrator (Burke, 1996) and the late Pan-African Najd trend (cf. Şengör and Natal'in, 1996a) being perhaps the most important.

The idea that a single superplume may have controlled African topography and rifting (Ebinger and Sleep, 1998) cannot, by contrast, explain the near simultaneity of the uplift of the entire continent at ca. 30 Ma (Burke and Wilson, 1972; Burke, 1996) as seen reflected in the marine sedimentary rocks all around the continent (cf. Burke, 1996), plus the pervasive +500 m residual depth anomaly of the ocean floor belonging to the African plate (e.g., Lithgow-Bertelloni and Silver, 1998). That the Afar plume is a very large one and that its effects may have dominated its neighbors of shallower origin are certainly compatible with what is known of its manifestations (Burke, 1996), but it is most unlikely to have been the only plume under Africa since the Eocene.

TWO EXAMPLES OF THE APPLICATION OF THE STRUCTURAL-UPLIFT CRITERION TO PLUME SEARCH IN THE PAST

There are no plume-related falcogenic domes active today on continental crust that lie below sea level. Neither were there any in the geologic past as far as I know. Even the most generous estimates of sea-level rise in post-Archean history seem insufficient to cover the heads of the plume plateaus (cf. Dewey and Pitman, 1998). A useful rule of thumb then is to consider *any* sub—sea-level volcanism on approximately normal-thickness continental crust unrelated to plume activity. This statement then brings me to an example of how my criterion works for identifying past plumes: The Givetian to Famennian volcanicity and rifting in the Russian craton has been attributed to the activity of mantle plumes (Wilson and Lyashkevich, 1996; Kusznir et al., 1997; Wilson et al., 1997) or even to one megaplume under the craton (Kusznir et al., 1997). However, a glance at the paleogeographic maps presented by Nikishin et al. (1996; also see the second volume of the Vinogradov Atlases, 1969, that give more detail) shows that this proposal cannot be applied to the Vyatka rift. There, the volcanicity commenced underwater, and only after rifting very narrow strips along bounding rift faults did rise to surface, most likely as a consequence of isostatic shoulder uplift (of the kind discussed by King and Ellis, 1990). The Devonian Soligalich rift just to the west of Vyatka also developed entirely underwater. Neither can the proposal be valid for the Pre-Caspian Depression, whose volcanism was wholly underwater. The Late Devonian Timan-Kolva and the Pechora volcanism and rifting also did not cause any part of the affected area to rise above sea level (Nikishin et al., 1996, p. 49). The Kola-White Sea rift system and the Dnyepr-Donets rift are possible candidates to be plume related owing to their paleoelevation, but the Donetz rifting and volcanicity commenced in the east entirely underwater. The Ukrainian Shield had been a dry area during the Eifelian and earlier, but the Voronezh High, north of the Dnyepr-Donetz, clearly rose above water after rifting, most likely as a consequence of shoulder uplift. Since Vyatka and the Pre-Caspian were definitely not related to plume activity, I find it unlikely that the Dnyepr-Donets was, because it occupies a very similar peri-Russian craton position. The Kola-White Sea rift system is the only good candidate to be plume related. In any case, a more detailed reconstruction of its paleogeography is certainly necessary before a detailed model can be constructed for the Devonian rifting and magmatism on the Russian craton.

By contrast, a glance at the Mesozoic disintegration of Gondwana-Land shows that the centers of volcanism identified as plumes on independent grounds (e.g., Morgan, 1981; Burke, 1996) were not only all above water, but they were also located in areas of erosion surrounded by terrestrial sedimentation. Figure 19 shows this on a simplified reconstruction. The flagrant contrast with the Late Devonian Russian case is obvious especially when the higher Mesozoic world-wide sea level than the Devonian sea level is considered.

CONCLUSIONS

The aim of this study was to identify *one* criterion that may be uniquely reliable in recognizing mantle plumes in the geologic record. That criterion is uplift. But not just any uplift. The uplifts must be falcogenic (*epeirogenic* in the sense of Stille, i.e., uplifts whose wavelengths far exceed their amplitudes and that rise essentially free of faulting); they must have commonly elliptical to circular map outlines, radii ranging from several hundred kilometers to more than 1000 km, and amplitudes of 1–2 km. Adjacent epeirogenic basins may augment structural amplitudes to a few kilometers. "Uplifts" can only be recognized if erosion does not erode them away as quickly as they rise. Kevin Burke (1996, and 2000, personal communication), for example, pointed to the Dakar plume in west Africa with no visible topographic uplift and warned me that if I make *topographic* uplift a criterion, it would not work for plume identification. *Structural* uplift ought to be the criterion for the identification of a former plume, not topographic. Such uplifts are easy to recognize (1) if the paleogeography preceding their birth can be reconstructed in the detail that, for example, the fortunate circumstance of proximity to the sea allowed in the case of the Afar dome (Fig. 11), (2) if their erosion products are preserved especially in a marine environment to enable precise dating (such as the Ahwaz delta: Fig. 15), and (3) if their structural expression escapes erosion as just north of Qalamat ash Sharawra in Saudi Arabia (Fig. 16). Extreme care should be exercised in not confusing uplifts of other kinds with plume-related uplifts. WoldeGabriel et al. (1990), for example, were misled into believing that the reduced Mesozoic section beneath the Afar dome was somehow related to doming. In reality, it was a product of a totally unrelated tectonism that affected the whole of northeast Africa and Arabia in the Early Cretaceous (as previously discussed herein). The slow retreat of the sea from the site of the future dome in the early Cenozoic is equally deceptive. What resembles a broad uplift was nothing more than a response to eustatic sea-level lowering, which in fact provided a guide to paleorelief (Fig. 12). To recognize uplifts related to ancient plumes by using the uplift argument, the following steps should be followed:

1. Locate broad uplifts of ~1000 km radius from stratigraphy and/or geomorphology (depending on the antiquity of the uplift).

2. Reconstruct paleogeography around the uplifts. Note eustasy. Note timing of uplift (Afar rose to ~1 km within ~5 m.y.). Make sure ice and/or water unloading was not involved (if data are permissive, time scales of 10^3 yr as opposed to 10^6 yr will usually give away ice and water unloading effects). Uplift rates much faster than ~250 m/m.y. are unlikely to be related to plume falcogeny.

3. Reconstruct uplift slopes. Slopes exceeding 1° are unlikely to have been related to plume falcogeny. Note that after faulting commences, much steeper slopes may be generated by shoulder uplift. Make sure that initially rising uplifts are free of *major* fault families. Falcogenic uplift generally must not change the mesoscopic and microscopic fabric of the rocks it affects (i.e., outcrop and hand-lens scale: "epeirogenic" of Stille, 1919, 1924); and it should thus be structurally *reversible* (though it may not be isostatically, unless underplated basalts turn into eclogites). This requirement naturally does not exclude minor faults atop rising domes, such as the ones described by Ajakaiye and Burke (1973). The role of preexisting fault sets may be most deceptive as the example of the Ahaggar dome shows.

Other than the activity of mantle plumes, there is no process on this planet that creates falcogenic domes (i.e., due to bending of the lithosphere and not faulting) of ~1000 km radius and 1–2 km amplitude within several million years at a rate of ~100 to 200 m/m.y.. That plume-related domes eventually and perhaps inevitably give rise to extensional deformation with dominant normal faulting is not a part of their definition. In fact, such taphrogenic faulting commonly confuses the record of the earlier falcogenic history of the domes and can give rise to misinterpretations about the relationship between doming and rifting.

So far as the present-day plume-related uplifts such as the ones shown on Figure 3 are indeed plume related, reconstruction of former epeirogenic uplifts with characters similar to the present ones is the most powerful tool I know in recognizing old plumes.

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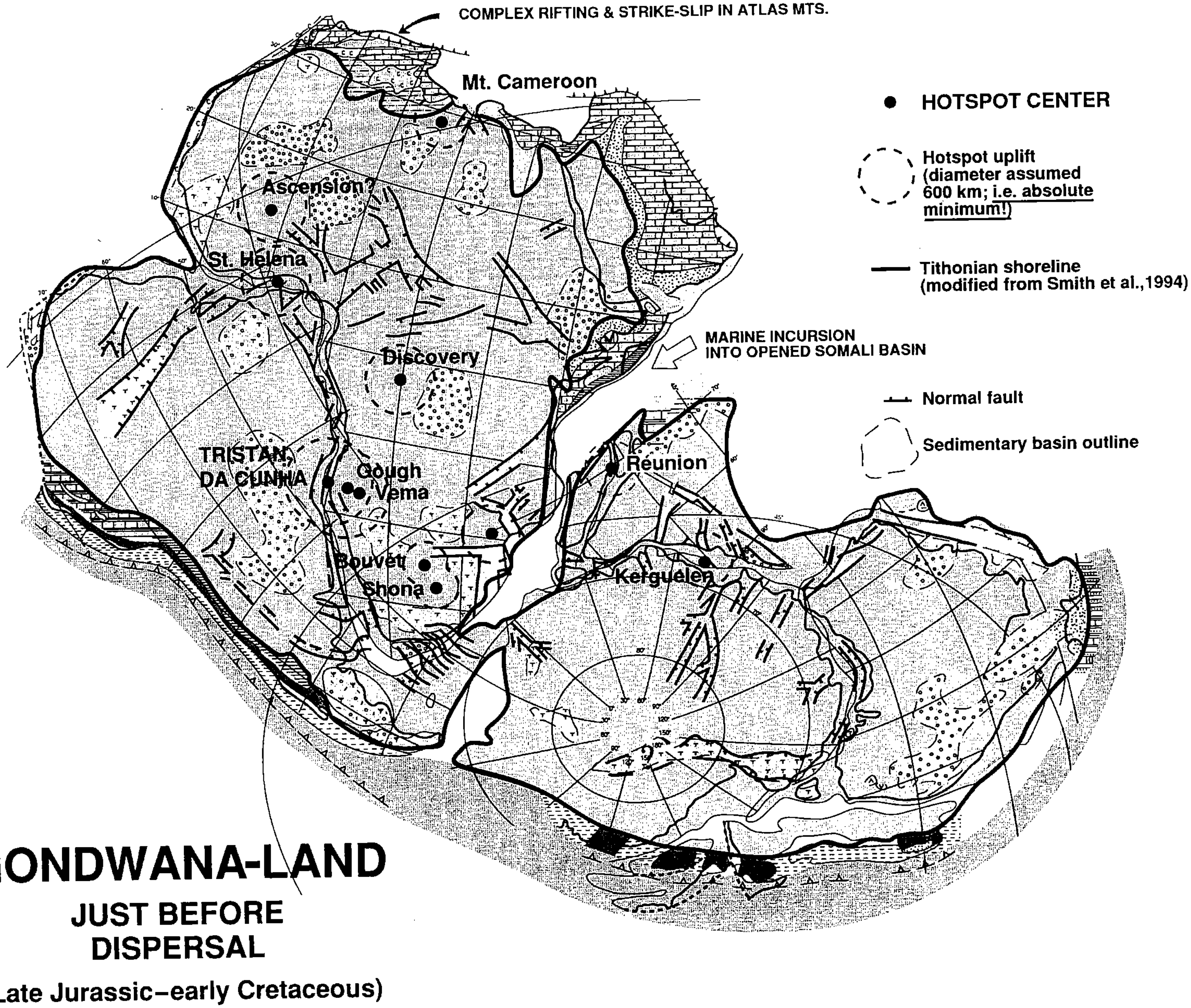

Figure 19. Gondwana-Land as it may have appeared at the end of the Jurassic and the beginning of the Cretaceous, ca. 145 Ma. Note that around reconstructed hotspots, areas of terrestrial sedimentation (light gray, dotted areas) seem to enclose round regions of erosion. This situation is consistent with the round regions being gently dipping falcogenic domes. Gondwana-Land reconstruction and sedimentary centers are modified from Voshko and Khain (1987). Shoreline position is modified from Smith et al. (1994). Rifts are from Şengör and Natal'in (this volume).

with buoyancy calculations (which would have had a more satisfactory form, had she actually written her own paper on the topic that I could have used!); she also kindly supplied me with those U.S. Geological Survey (USGS) publications that I was unable to get in Istanbul and provided a great review. Sirri Erinç taught me much of what I know about pedology. Warren Hamilton gave me his own great collection of USGS Arabia publications, which saved me from much trouble, and shared with me his own impressions of the tectonics of the Afar triple junction. Bill Dickinson has regularly fed me with offprints, preprints, and letters on the topics related to those here discussed. Boris Natal'in has been ready, as ever, with useful advice whenever I needed it. Xavier Le Pichon criticized and led to the improvement of my lithosphere-detachment model. I thank Luce Fleitout and Isabelle Manighetti for informative discussions. Ziyadin Çakir helped me with the construction of Figures 4, 5, and 14, and both he and Kadir Eriş assisted me with some library research. Erden Soysal drafted some of the figures. I thank Naci Görür and Fazli Oktay for tolerating my prolonged absences from the department and Mehmet Sakinç for making them possible and for checking my paleontology. Robert H. Rainbird and Marjorie Wilson reviewed the first draft and offered helpful suggestions. A version of this paper was presented as a conference in a special session of the Austrian Geological Society in the University of Vienna, which enabled me to benefit from the comments of my audience. Last, but not least, I thank my late regretted friend Professor Ziad Rafik Beydoun, the dean of all Middle Eastern geologists, for having enriched me with presents of offprints, books, and numerous letters containing detailed hand-drawn maps concerning the geology of the Middle East and northeast Africa ever since I was a student.

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Cover: The model of mantle plumes is modified after Figure 1 in N. Arndt (2000, Hot heads and cold tails, Nature, v. 407, p. 458–459), and is reprinted with the permission of *Nature*, copyright 2000, Macmillan Magazines, Ltd. Plumes are shown rising from the core mantle boundary or from an intra-mantle boundary. The former can either ascend directly into the upper mantle or stall at an intra-mantle boundary and spawn "plumelets." The Earth image used as a background was obtained from NASA (National Aeronautics and Space Administration). The bar diagram shows the distribution of well-established (in red) and probable (in black) mantle plume head events through time (modified after Figure 2 in R.E. Ernst and K.L. Buchan, Chapter 19, this volume).

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