Origin of the Bermuda volcanoes and Bermuda Rise: History, Observations, Models, and Puzzles

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Abstract

Deep-sea drilling on DSDP Leg 43 and on Bermuda itself, together with geophysical data (anomalies in basement depth, geoid and heat flow) and modeling have long suggested the uplift forming the Bermuda Rise, as well as the initial igneous activity that produced the Bermuda volcanoes, began ca. 47-40 Ma, during the early to middle part of the Midde Eocene. Some authors attribute 65 Ma igneous activity in Mississippi and 115 Ma activity in Kansas to a putative "Bermuda hotspot" or plume fixed in the mantle below a moving North America plate. While this is more or less consistent with hotspot traces computed from 'absolute motion' models, the hotspot/plume must resemble a lava lamp, turning off for up to 25 million years at a time, and/or be heavily influenced by lithosphere structure. Moreover, Cretaceous igneous activity in Texas and Eocene intrusions in Virginia then require separate mantle "blobs".

The pillow lavas forming the original Bermuda shield volcano have not been reliably dated, and the three associated smaller edifices have not been drilled or dated. A well-dated (ca. 33-34 Ma) episode of unusually titaniferous sheet intrusion in the Bermuda edifice was either triggered by plate-wide stress changes or reflects local volcanogenic events deep in the mantle source region. The high Ti and Fe of the Bermuda intrusive sheets probably relate to the very high amplitude magnetic anomalies discovered on the islands. Numerical models constrained by available geophysical data attribute the Bermuda Rise to some combination of lithospheric re-heating and dynamic uplift. While the relative contributions of these two processes cannot yet be wholly separated, three features of the rise clearly distinguish it from the Hawaiian swell: 1) The Bermuda Rise is elongated at right angles to the direction of plate motion; 2) There has been little or no subsidence of the Rise and volcanic edifice since its formation—in fact, Rise uplift continued at the same site from the late Middle Eocene into the Miocene; and 3) The Bermuda Rise lacks a clear, age-progressive chain.We infer that the Bermuda Rise and other Atlantic mid-plate rises are supported by anomalous asthenosphere, upwelling or not, that penetrates the thermal boundary layer and travels with the overlying North America plate.

The elongation, along crustal isochrons, of both the Bermuda volcanoes and the Rise and Rise development mostly within a belt of rougher, thinner crust and seismically "slower" upper mantle—implying retention of gabbroic melts at the ancient MAR axis—suggests the mantle lithosphere may have helped localize rise development—in contradiction to plume models. The Bermuda Rise area is seismically more active than its oceanic surroundings, preferentially along old transform traces and possibly reflecting a weaker upper mantle- lithosphere.

We attribute the "Bermuda event" to a global plate kinematic reorganization, triggered by the closing of the Tethys and the associated gravitational collapse into the lower mantle of subducted slabs which had been temporarily stagnant near the 660 km mantle discontinuity. The widespread onset of sinking slabs required simultaneous upwelling for mass balance. In addition, the global plate kinematic reorganization was accompanied by increased stress in some plate

interiors, favoring magma ascent along fractures at structurally weak sites. We suggest the Bermuda event and concomitant igneous activity in Virginia, West Antarctica, Africa, and other regions were among such upwellings, but structurally influenced by the lithosphere, and probably originating in the upper mantle.

Drilling a transect of boreholes across and along the Bermuda Rise—to elucidate turbidite offlap during rise formation— might discriminate between a widely distributed mantle source and a narrow plume whose head (or melt root) spreads out radially over time, generating an upward and outward expanding swell.

Introduction

The Bermuda Rise (Fig. 1), capped by a cluster of extinct volcanoes (Fig. 2, bottom) dominated by the Bermuda Pedestal, dominates the western North American Basin (western central North Atlantic). The rise forms an oval basement swell that apparently rose, with concomitant volcanism, in middle Eocene through Oligocene, perhaps into Miocene times. Geoscientists from the 19th to middle 20th centuries were interested in the geology of Bermuda and its pedestal (See Foreman, 1951, for review of early interpretations). However, with the advent of plate tectonics, attention was directed to the associated Bermuda Rise, which has become in many ways the archtypical—and to date best studied—regional basement swell developed in pre-existing and apparently **slow-moving** (ca. 15-30 mm/a) oceanic lithosphere, variously contrasting and comparable with the Hawaiian-Emperor island/seamount chain and its associated basement swell, formed and still developing within **fast-moving** (ca. 90 mm/a) Pacific lithosphere. While much of the **geological history** of the igneous activity is reasonably well known, the underlying geophysical processes are subjects of long-standing and ongoing debate**:** For example, is the apparently "unifying hypothesis" valid--that both features are caused by mantle plumes/hotspots, differing only insofar as the Bermuda "plume" is more like a lava lamp, with intermittent blobs of hot mantle (e.g., Morgan, 1972; Crough, 1978)? Of the many plumes that have been postulated, Bermuda, although throughout the last 30 years appearing in most "hotspot catalogues" (Steinberger (2000) is a rare exception), rates practically at the bottom in terms of consistency with the deep mantle plume hypothesis (e.g., Anderson, 2005; Anderson and Schramm, 2005; Courtillot et al., 2003). Is the Bermuda Rise an expression of thermal expansion of hot underlying mantle, low density materials intruded into the plate, a low density buoyant root of partial melt residue, dynamic uplift maintained by convection, or some combination (e.g., Sclater and Wixon, 1986; Detrick et al., 1986; Louden et al., 1987; Sheehan and McNutt, 1989; Sleep, 1990; Phipps Morgan et al., 1995)? There are troubling inconsistencies between data and any of these mechanisms, and even problems with the data themselves, specifically heat flow measured from shallow penetration deep-sea probes (e.g., Ruppel, 1996; 2000). Problems extracting a mantle thermal signal from heat flow data on the Hawaiian swell (e.g., Harris et al., 2000; McNutt, 2002) probably complicate the heat flow interpretation issue on the Bermuda Rise as well.

In this paper we offer an in-depth review of what is known of the geology and geophysics of Bermudan igneous volcanism (first) and the associated Bermuda Rise (second). With these data as backdrop, we then examine the geophysical models that have been presented for the deep (i.e., mantle) processes. The possibility that the postulated "Bermuda hotspot" has a trace that extends into and accounts for the Cretaceous igneous intrusive activity in the Mississippi Embayment area (Cox and Van Arsdale, 1997, 2002) is treated separately from the geophysical models for the present Bermuda Rise. We propose some new ideas to reconcile the differences between Hawaii and Bermuda, and between observations and models. Our organization is partly historical—we review the observational database for each parameter, and modeling efforts more or less in their historical order.

The primary time scale used in this paper to convert from stratigraphic to absolute ages is the one of Gradstein et al. (2004). We also used this time scale to recompute the age of the original oceanic crust below the Bermuda edifices, and, where we cite specific magnetic anomalies or chrons, their ages based on the 2004 time scale. However, when referencing older work which directly or indirectly depends on the stratigraphic time scale (e.g., spreading rates, heat flow relative to crustal age, etc.), we did not adjust the published values. For the most part, such adjustments are small compared to other uncertainties in the values.

Bermuda: Igneous birth, Erosion and Subsidence

Setting and Morphology

The Bermuda Islands (often referred to simply as “Bermuda”, sometimes as “the Bermudas”) rise up to 76 m above sea level along the southeastern margin of a ca. 50 km long, NE-elongated oval bank (Fig. 2), long considered the truncated stump of a large, extinct shield volcano, generally referred to as the Bermuda Pedestal. A thin (15-100 m) mainly Quaternary-aged carbonate cap covers the truncated top of the volcanic pedestal (see further below). Bermuda is the largest of four evidently related seamounts arranged in a 100 km long, NE trending line (Fig.1, bottom), perched on the summit of the broad Bermuda Rise (discussed separately below). The four edifices, starting in the southwest, are Plantagenet Bank (or “Shelf”), later also known as Argus (or “Argos”); Cox, 1959) Bank, Challenger Bank, Bermuda, and Bowditch Seamount. Applying the same terminology used by Vogt and Smoot (1984) in their analysis of the Cretaceous Geisha Guyots (Japanese Seamounts) in the western Pacific, we would call the short Bermuda chain a “cluster”—as further discussed later.

The present Bermuda islands, which occupy only the southeastern 7% of the ca, 665 km^2 Bermuda platform, are almost entirely composed of limestones-- variously indurated calcareous dune deposits (aeolianites) originally eroded from biogenic, primarily coral reef limestones during low sea levels of several Pleistocene glaciations, when the entire Bermuda summit platform was exposed. A variety of corals presently grow on the platform—thus making Bermuda the most northerly coral reef habitat in the modern Atlantic Ocean. The present Bermuda platform consists of a central lagoon averaging about 18 m in depth (greatest lagoon depth is 25 m in Devil’s Hole), surrounded except at the islands by a coral reef 2 to 10 m deep (Fig 2, top). Just beyond and bordering this narrow (ca. 1-1.5 km wide) reef is a ca.1-3 km wide, ca. 20 m deep “Terrace Reef Zone”. Scattered coral patches within the lagoon rise to within a few meters of sea level. The lagoon, reef and outer terrace are known collectively as the Reef Platform (Logan, 1988). At a smaller scale, the pattern of reefs and islands (Fig. 2, top) include a number of ring-like or semicircular features with diameters ca. 500-1500 m. The relation, if any, of this morphology to the underlying volcanic “basement” is unknown. Although resembling typical Pacific atolls in morphology, the widespread occurrence of aeolianites, especially in the islands, have prompted some authors to call Bermuda, at 32.3º N one of the northernmost living coral reefs on the planet, a “peudoatoll”.

Four Interglacial soil horizons occur within the Pleistocene carbonate sandstones (Livingston, 1944; Harmon et al., 1978). Surficial deposits are mostly eolianite, with isolated sublittoral marine and beach deposits (Vacher et al., 1989). Dated Late Pleistocene sea levels perhaps 1-2 m above present (ca. 80 ka; Vacher et al., 1989; Ludwig et al., 1996) and at +20 m (ca. 420 ka; Hearty et al., 1999) constrain not only eustatic sea level fluctuations and global ice volumes, but also geophysical models for the Bermuda Rise (see below), most of which make predictions about elevation and later subsidence of the rise. Data collected to date suggest that reef limestone began to accumulate on the slowly subsiding, wave-eroded summit igneous plateau in the early Miocene—thus indicating that net post-volcanic subsidence has been negligible, averaging only about .004 mm/a. At least under parts of Bermuda, the sequence of limestones and interglacial soils is underlain by a lateritic clay horizon ("Primary Red Clay") derived from subaerial weathering of the volcanic basement in a humid tropical or semi-tropical environment (Moore and Moore, 1946; Foreman, 1951).

The 100 km long chain (or "cluster", using the terminology of Vogt and Smoot, 1984) of four edifices trends 035 (Johnson and Vogt, 1971; the oval Bermuda edifice itself trends somewhat more easterly; 055) paralleling and ca. 30-40 km southeast of reversed Chron M-0 (Fig. 1; Rice et al., 1980). Based on the most recent age estimate for M-0 (125.0 +/- 1.0 Ma; Aptian-Barremian transition; Gradstein et al., 2004) and an assumed but poorly constrained spreading half-rate of 2 mm/a after M-0 time, we estimate the age of the oceanic crust under the Bermuda edifices at ca. 123-124 Ma. Thus the Bermudan volcanoes (ca. 45-35 Ma in age, see below) grew from oceanic crust ca. 78-88 Ma in age at the time of volcanism. The apparent control by old seafloor structure on the linear arrangement, paralleling crustal isochrons, of the Bermuda volcanoes contrasts with a near total absence, along the Hawaiian chain, of any evidence for such control by pre-existing crustal structures (e.g., paralleling or perpendicular to Cretaceous transform trends crossing the chain).

2) *Early Bermuda and Bermuda platform geophysics*

Charles Darwin first suggested that the many isolated Pacific coral reefs are underlain by stumps of eroded former volcanic islands. A similar origin for Bermuda must have occurred to many scientists and naturalists before the 20th century. The idea of an ancient volcano under Bermuda goes back at least to Verrill, who also believed it was of Triassic age (as quoted by Pirsson, 1914a). The first hint of a volcanic edifice at depth came indirectly: At least as far back as the 1873 Challenger expedition, local declination (compass variation) anomalies were apparently known at Bermuda. Given the dependence of navigation on compass bearings, there was a good practical reason to investigate these anomalies. The first to do so was JF Cole (1908), who mapped the magnetic declination throughout the islands in 1905. His "isogon" map revealed high spatial frequency anomalies, which Cole merely attributed to "sources of considerable local disturbance". In retrospect, we can consider these data the first sign of highly magnetized igneous rock at shallow depth below Bermuda.

The Dept. of Terrestrial Magnetism of the Carnegie Institution then sent two expeditions to Bermuda (1907 and 1922) to "study the magnetic anomaly known to exist there" (Fisk, 1927). Attempts were made to reoccupy some of the 1873 Challenger station sites. At the time it was unknown that local declination anomalies could be caused by remanent magnetism, hence the emphasis on repeat measurement. While Cole had only measured variation of the compass (declination), Fisk included inclination and horizontal intensity, measuring these at five primary and 78 secondary stations. Two decades prior to the invention of the proton precession magnetometer, measurements typically took half an hour per station! By the time of the 1922 measurements, a 1912 borehole had recovered igneous rock from shallow depth (see below) and Fisk (1923) could declare that "the island is a submerged mountain of volcanic origin". Fisk reported horizontal changes as much as 700 nT over a horizontal distance of only ca. 6 m. Determining that the magnetization of the limestone and topsoil were inadequate to account for the anomalies, Fisk (1923) concluded that "There are, therefore, a major or primary source of disturbance lying deep in the lower structure of the submerged mountain".

The next geophysical advance was that of Woollard and Ewing (1939), who ran two seismic refraction profiles in the islands, returning depths of 83 and 74 m to the top of volcanic rocks. The experiment yielded a P-wave speed (4.88 km/s for this "basement"), with the overlying calcareous sandstone yielding 2.68 km/s. These may have been the first seismic refraction experiments on a buried igneous basement in an oceanic venue.

In addition to the two seismic lines, Woollard and Ewing (1939) analyzed both the isogon chart of Cole (1908) and Fisk's measurements (1907 and 1922) of horizontal field strength on the islands, concluding from depth-to-source calculations that the basement depths are "all about the same and less than 1200 ft (366 m)". The authors went on to conduct "a series of observations on the vertical component of the earth's field and find very high local disturbances both positive and negative in character". The 1939 Nature paper is a very short summary, and to our knowledge the actual seismic and magnetic data were never published, owing perhaps to the onset of World War II.

As part of an extensive seismic reconnaissance of the Atlantic, Officer et al. (1952) reported four marine seismic refraction profiles shot across the north-central part of the platform (off Hamilton Island) in the summer of 1950 and spring of 1951. In water depths from 12 to 15 m, the experiments returned "basement" depths from 35 m to 96 m, and two volcanic basement P-wave speeds of 5.13 and 5.34 km/s. In a footnote, Officer et al. (1952) cite unpublished 1949 (M. Ewing, G.R. Hamilton, F. Press, and J.L. Worzel) unreversed refraction profiles in the same area west of northern Bermuda—returning a basement depth of 90 m, a sediment P-wave speed of 2.7 km/s, and two basement speeds of 5.43 and 4.21 km/s. Historically, probably the most significant discoveries of Officer et al. (1952) were not the structure just below the Bermuda platform, but the relatively thin, albeit overestimated, sediment layer and oceanic crust (Moho at only 10 km bsl, with no "granitic layer") in the North America basin, on the east flank of the Bermuda Rise. Starting a decade later, these surprising results would be understood in terms of seafloor spreading and then plate tectonics. Even though borehole microfossil analysis had found nothing older than Eocene (Pirsson, 1914a, 1914b; see below), Officer et al. (1952) preferred a Triassic age for the Bermuda volcanic pedestal—completely ignoring the evidence for continental drift (Wegener, 1929) which implied a younger age. Officer et al. (1952) anticipated the still unexplained lack of edifice subsidence by concluding "that the volcanic platform was planed to its present depth during a Pleistocene glacial lowering of sea level" and that "It is not necessary to hypothesize subsidence to explain the known geology of Bermuda".

Bermuda boreholes

Long before seismic methods first measured the thickness of the Bermuda carbonate cap, the latter had already been directly measured when an exploratory well ("Gibbs Hill Boring") was drilled in 1912 to 1283 ft (391 m) bsl for the owners of a local hotel, in the hopes of discovering potable water supplies. Although the borehole failed in this objective, it did prove the presence of a volcanic "stump" under the island. The shallowest volcanogenic rock appeared at 230 ft (70 m) bsl (Pirsson and Vaughan, 1913; Pirsson, 1914ab; Livingston, 1944; Foreman, 1951). Pirsson's (1914ab) borehole section noted the oldest microfossils on the volcanic basement as Eocene. Remarkably, this age has stood the test of almost a century! Moore (1942) and Moore and Moore (1946) present syntheses of Bermuda geology prior to the discovery of plate tectonics.

The second borehole to reach volcanic basement was drilled in 1958 (Rice et al., 1980). Later seismic reflection surveying showed the volcanic "basement" surface to vary from 200 m to 25 mbsl, averaging about 76 mbsl (Gees and Medioli, 1970), close to the two refraction results of Woollard and Ewing (1939). The first attempt at age-dating Bermudan volcanism (Gees, 1969) yielded 34 and 52 Ma for two chips of igneous rock from the 1958 borehole.

A new phase of research on Bermudan volcanism was initiated by the continuously cored, research-driven "Deep Drill 1972" borehole, which penetrated 802 m below the ground surface, reaching volcanics at only 26 mbsl and continuing 767 m into the igneous basement (Reynolds and Aumento, 1974; Aumento and Sullivan, 1974; Hyndman et al., 1974, 1979; Rice et al., 1980). The cores revealed “431 submarine pillows or flows of hydrothermally altered tholeiitic composition” and “493 interfingering intrusive limburgitic sheets” of unusual composition (Rice et al., 1980). The flows were thin, averaging only 1 m in thickness, and the sheets were preferentially intruded between flows or into pre-existing sheets. The 1972 cores have recently been reexamined by M.-C. Williamson of Dalhousie Unviverity (M.-C. Williamson, personal communication, 2005; Williamson et al., 2006). A paper initially rejected for publication is now available on the Internet (Aumento and Gunn, 2005). A new geochemical analysis of dike rocks recovered in both the 1972 and 1980 (see below) boreholes was performed by S. Olsen (2005).

Reynolds and Aumento (1974) reported a minimum (K-Ar whole rock) age of 91 ± 5 Ma for the flows, and a more reliable phlogopite (mica) age of 33.5 ± 2 Ma for the sheets. The latter date has been confirmed by new Ar-Ar dating of samples from the same cores (M.-C. Williamson, personal communication, 2006; Williamson et al., 2006)

The intrusives have been variously described as “lamprophyre” and “nephelenite” dikes, and (Aumento and Gunn, 2005, based in part on their 1970s work on core samples from Deep Drill-1972) alternately as “very titaniferous alkaline rock”, “melilite melanephelinite” or “Bermudite”. The radiometric ages were averaged from whole-rock K-Ar ages of 47 and 91 Ma for the flows and 32, 14, and 30 Ma for the lamprophyre intrusives. In addition, mineral dates of 27, 36 and 33 Ma were obtained for “phlogopite” samples from the intrusives. The flows are “albitized and chloritized basalts of the OIB (Ocean Island Basalt) type” (Aumento and Gunn, 2005), resembling. Mid-Oceanic Ridge basalts of the “EMORB” variety, i.e., somewhat enriched in large-ion lithophile elements. The intrusive the sheets (dikes) are highly evolved rocks, with very high TiO_2 concentrations (5 to 6.2 %), FeO* exceeding 16 %, high P_2O_5 (to 2%) and K_2O, high La/Sm (up to 7.8) and Sr (550-1050 ppm), and very low SiO_2 (30-40 %) and Zr/Nb (as low as 2). The above compositions were taken from (Aumento and Gunn, 2005; see also Olsen, 2005; Aumento et al. , 1974, 1975). Significantly, no subaerially extruded flows were cored, not even at the top of the igneous section, now at 26 mbsl. This observation strongly suggests the flows were uplifted after they were erupted.

Two more recent research boreholes (1980) were drilled at Government Quarry, Bailey’s Bay on the south shore of Castle Harbor—where the carbonate cap was known to be less than 50 m thick—and over the gravity anomaly maximum at the Naval Air Station Annex (Peckenham, 1981). The latter borehole was abandoned in undrillable unconsolidated sediment at 42 m depth. However the first borehole was successfully drilled to 98 m depth, well into the igneous basement. Nearly 50m of submarine melilitic, variously altered pillow lavas were cored, associated with volcaniclastic rocks cut by narrow melilite-bearing intrusives. The igneous rocks, resembling those recovered by Deep-Drill 1972, are silica-undersaturated, titanium- and volatile-rich (Peckenham, 1981; Olsen, 2005). Paleomagnetic studies of the lavas were unsuccessful, as inclinations were either not stable or not reproducible. Forammininfera (Orbulina and Globigerinids) in interpillow sediments are mid-Tertiary in age, consistent with other evidence on the age of Bermuda volcanism (Peckenham, 1981).

Based on a careful new analysis, Olsen (2005) concludes the intrusive dikes (sheets) can only have originated by partial melting of very large-ion enriched, pyroxenite-veined depleted upper mantle at high pressures of 5 GPa (ca. 150 km depth) or more. Depths that great clearly put the magma sources in the asthenosphere—in the putative plume head.

The 1980 drill hole was supplemented in 1981 by submersible (Pisces IV) exploration (Peckenham et al., 1982). Three rock samples, two of them igneous (volcaniclastic breccia and porphyritic limburgite), were collected at 600-700 mbsl depths on the southern flank of the edifice. The sample composition is consistent with core materials and volcaniclastics previously recovered at DSDP Site 386 (discussed below).

Borehole results from the surrounding ocean floor

In 1975 R/V Glomar Challenger drilled a hole (DSDP Site 386; The Shipboard Scientific Party, 1979) ca. 140 km SE of Bermuda (Fig. 1), adding another dimension to the story of island volcanism (The Shipboard Party, 1979; Tucholke and Vogt, 1979; Tucholke and Mountain, 1979). The seismic reflecting horizon A^v was found to correspond to the top of coarse volcaniclastic turbidites originating from subaerial (and subtropical) weathering and erosion of the volcanic islands that existed on the Bermuda pedestal and its two satellite banks to the southwest. A^v extends up to 200km away from Bermuda; upon approaching the platform the reflector merges with the acoustically opaque archipelagic apron around the platform base. As far as known to date, the Bermuda volcanoes did not erupt large amounts of ash: No discrete ash horizons were recovered at DSDP sites 386 or 387, and although typically a few % "altered ash" was reported in most of the volcaniclastic samples described at Site 386, these were always followed by a "?" (The Shipboard Party, 1979).

Volcaniclastic turbidites, with intermixed shallow-water detritus, began to arrive at Site 386 in late middle or early late Eocene times (ca. 40-36 Ma), indicating that the Bermuda volcano (or volcanoes) had risen to sea level no later than that. The volcaniclastic turbidite sub-unit (clay, silt and sand) is more than 162 m thick, with the thickest and coarsest individual turbidites, suggesting the most intense erosion and/or increased volcanism and/or improved turbidite dispersal pathways to the site, happened during the middle Oligocene.

Interpreting the volcaniclastic turbidite record at Site 386 in terms of volcanic episodes is questionable, however. Independent evidence suggests sea level fell by 50-75 m during the middle Oligocene (e.g., Haq et al., 1987; van Sickel et al., 2004), so the apparent increase in turbidite frequency and thickness at that same time may reflect increased exposure of the volcanic edifice to erosion, not increased volcanic activity. Furthermore, the lack of subaerially erupted basalts—even at 26 m below sea level—in the 1972 drill hole (Aumento et al., 1975; Hyndman et al., 1979) is consistent with uplift of an initially deeper volcanic pile, due to 1) edifice inflation associated with intrusion of the sheets, or other intrusions (as has apparently happened in the Cape Verde and Canary Islands, resulting in subaerial exposures of original oceanic crustal basalts) and/or 2) continuing (subsequent to the eruption of the tholeiitic lavas) uplift of the Bermuda Rise, and/or 3) exposure or re-exposure by lowered sea levels.

In their initial interpretation of drill core results, Aumento et al. (1974) envisaged " a massive buildup of tholeiitic lavas near the axis of the Mid-Atlantic Ridge at least 91 m.y. ago", at or soon after the formation of the oceanic crust itself at the MAR axis. This "huge seamount (or island) remained dormant for almost 60 m.y.", sinking below sea level and experiencing protracted low-temperature hydrothermal alteration. The 33 Ma intrusion of lamprophyre sheets then, according to Aumento et al.(1974), inflated the edifice by almost 40 %, causing it to emerge (perhaps for the second time) above sea level. This early interpretation never appeared in a full paper—the 91Ma date being generally disregarded as unreliable by most later authors—and an early (pre-91 Ma) formation of a large seamount or even island near the MAR axis was scarcely supported by Site 386 drilling results, and is inconsistent with the isostatic seamount-height limitation developed by Vogt (1974,1979a).

In any case, the lack of subaerial eruptives at shallow depths below Bermuda is atypical of oceanic volcanic islands, where loading-induced subsidence has depressed subaerially erupted rocks up to 1 km or more below sea level (e.g., Hyndman et al., 1979, p.98). (The Cape Verde and Canary archipelagos show evidence for significant uplifts, however). While subaerially erupted Bermudan rocks may well have once existed at higher levels in the edifice—and were then eroded---better evidence of this must await additional coring and analysis of the volcaniclastics in Horizon A^v. So far the only evidence for subaerial or shallow-water eruptions are the few % "altered ash?" noted in shipboard descriptions of the volcaniclastic sands at Site 386 (The Shipboard Scientific Party, 1986), and the altered tuff described by Foreman (1951), who inferred "volcanic eruptions of the explosive type, both andesitic and basaltic".

None of the publications describing the 1972 and 1980 drill cores reported even increased degrees of fracturing or increased vesicle volumes in the shallowest basalt flows, effects that would be expected had those flows been erupted in water depths of less than a few hundred meters (e.g., Moore, 1970; Moore and Schilling, 1973; Duffield,1973; Schilling, 1986). Although vesicularity was not reported in the drill core descriptions, unusually high vesicle volume probably would have been noted. Alkalic basalts generally contain more volatiles and would show systematically higher vesicularity at any eruption depth (Moore, 1970; Moore and Schilling, 1973), but even the tholeiites cored at Bermuda should be significantly (say, more than ca. 5 %) vesicular—particularly in pillow interiors—if they had been erupted at water depths less than ca 1000 m (see, e.g., Fig. 2 of Duffield, 1973 or Fig. 10 of Schilling, 1986). Thus, the shallowest recovered flows can scarcely represent the very top of the submarine part of the lava pile, but rather must have been uplifted, whether or not they were covered by additional submarine lava flows subsequently removed by erosion.

The age of the youngest volcaniclastic sediments at Site 386 is late Oligocene (ca. 25 Ma), so the volcanic foundation of Bermuda must have been emergent or shallow water at least until that time. As is true for Hawaiian volcanism, directly erupted materials such as pumice or ash are not abundant, and no discrete ash horizons were cored either at Site 386 or the more distant Site 387. “Muscovite” mica is abundant in the youngest volcaniclastic turbidites, but with the rarity of associated quartz is extremely unlikely to have originated from distant continental sources. Very likely this mica is the same phlogopite recovered and dated (see above) in the 1972 drillhole on Bermuda, its stratigraphic age at Site 386 (middle to late Oligocene, ca. 30-25 Ma) in reasonable agreement with the 33.5 ± 2 Ma radiometric age, and further attesting to a later stage of Bermudan igneous activity, the intrusion of the lamprophyre sheeted dikes. The dike ages would be expected to be somewhat older than the stratigraphic age of their erosion products, as observed). If the radiometric age of the altered flows (91±5 Ma) is ignored as unreliable—i.e. the edifice-building stage is dated by via the Site 386 stratigraphic occurrence of edifice erosion products, the intrusions post-dated the shield-building stage by ca. 5 to 10 million years: A better estimate of hiatus length awaits an accurate age determination of the lavas.

Turbidites continued to arrive at the site to the end of the early Miocene, however, and although primarily composed of carbonate, still contain appreciable amounts of altered ash, heavy minerals, and zeolites, suggesting continued erosion of local outcrops and possibly minor, late-stage explosive activity.

DSDP Site 386 cored one other interval---much deeper and older—with abundant volcanogenic components (The Shipboard Scientific Party, 1979). A 29 m thick, primarily Upper Cenomanian (ca. 95-94 Ma) bed of zeolitic claystones contains alteration products of volcanic ash, with glass shards indicating the volcanism was primarily pyroclastic. While this date agrees with the older of the two whole-rock ages (91 Ma) reported by Reynolds and Aumento (1974) for the Bermudan lavas—leading those authors and Peckenham et al. (1982) to suggest Bermudan volcanism began ca. 90-110 Ma, near the ancient axis of the Mid-Atlantic Ridge, the agreement is most likely fortuitous: The cored lavas and Bermuda pedestal morphology suggest passive, flow-dominated shield volcanism, not pyroclastic activity. Furthermore, the present elevation of the wave-eroded igneous rock platform under the islands, above the adjacent oceanic basement (ca. 4.5-5 km) is too high for an origin near the axis of the MOR (Vogt, 1979), and independently supports an off-axis origin for the volcanoes.

Eroding the original Bermuda volcanic islands

The original maximum elevation of Bermuda, prior to erosion, was estimated at 3.5 km above sea level, based on extrapolation of submarine slopes by Pirsson (1914ab). A rough comparison with modern volcanic islands led Vogt (1979) to reduce this to 3 ± 1 km for Bermuda, and roughly 1 km for Plantagenet and Challenger Bank. A more careful analysis of active volcanic island maximum elevations vs exposed area is shown in Fig. 6 of Vogt and Smoot (1984). Using the 116 km^2 Bermuda bank area and the best-fitting line in their figure yields an original Bermuda island elevation of ca. 1000 m. The volume of material eroded can be estimated from Fig. 7 of Vogt and Smoot (1984) as 30 km^3, and the time to reduce the island to sea level from 3 to 10 million years, for shoreline erosion rates 2 to 0.5 km/million years, respectively. The magnitude of the erosion time is consistent with the length of time Bermuda erosion products are stratigraphically important at Site 386, as discussed above.

The estimate of 30 km^3 for the total eroded rock volume, if distributed evenly over the region where Horizon A^v (Fig. 1; Tucholke and Mountain, 1979) can be identified in seismic records, is equivalent to a layer only ca. 0.3 m thick. This is comparable in magnitude to the minimum detectable thickness for a reflection (1/30 of the dominant seismic wavelength, i.e. ca. 0.5 m for 100 kHz sound traveling through a 1500 m/s medium; B.E. Tucholke, personal communication, 2005). However, the actual thickness of A^v is likely to exceed this minimum, and in any case is vastly less than the cored thickness of volcaniclastic material is 162 m at Site 386; even upon reduction to original rock volume and allowing for dilution by non-volcanic sediments, this is two orders of magnitude greater than 30 cm). Allowing for the inevitable concentration of volcaniclastics in valleys (as at Site 386) and near the islands would make the calculated thickness even less elsewhere in the A^v area mapped by Tucholke and Mountain (1979). One interpretation of this paradox is to infer that the Bermuda volcanoes were active long enough to produce the equivalent of many successive emergent islands. This however would invalidate the above erosion time calculation, which assumed the island was constructed and then eroded.

The total volume of igneous intrusive and extrusive rock produced by the four Bermuda volcanoes is unknown, due to the uncertain volume contained in A^v, the uncertain mass intruded into preexisting lithosphere, and the amount of lava, debris, and intrusives filling the flexural depression below the edifices (The latter could in principle be estimated by flexural models such as Sheehan and McNutt (1989) or by seismic reflection and refraction methods described in Purdy and Ewing (1986)). We calculated a minimum volume of ca. 760 km^3, the edifice relief above the 5 km depth level. Of this total, the main Bermuda edifice accounts for about 72 %, Plantagenet and Challenger Banks 8 % each, and Bowditch Seamount 12 %.

The Bermuda volcanic cluster: Comparisons with the Japanese Seamounts

Compared to volcanoes along the Hawaiian chain, the Bermuda cluster (including the three satellite edifices) is a mere pimple, with a total volume only about 10 % of a typical Hawaiian edifice. Guyots and seamounts along the 1100 km long Geisha (now called the Japanese Seamount, hereafter abbreviated to JS) chain (Vogt and Smoot, 1984), provide a much better morphological match with Bermuda. In length and degree of linearity, the 100km long Bermuda cluster (four edifices) closely resembles the Winterer cluster (six edifices) in the JS. Summit plateau break depths along the these guyots (ranging from ca. 104 to 94 Ma in age) rise on average 4650 m above the basement, indistinguishable from the relief of Bermuda's wave-eroded igneous basement above the regional oceanic basement. The platform area of Bermuda (116 km^2) is within measurement error identical to that of Charlie Johnson Guyot. The ca 600 km^3 volume of the Bermuda edifice above the regional basement is comparable to that of Thomas Washington Guyot (800 km^3), and the volumes of the three Bermuda satellite edifices are comparable to the ca. 100 km^3 volumes of some of the smaller JS, e.g., WC-3 and W-2 of Vogt and Smoot (1984).

There are nevertheless four significant differences between the JS) and Bermuda—and at present it is unclear if these differences are related. First, as already implied, the JS comprise a number of guyot/seamount clusters, arranged in a linear chain, whereas Bermuda constitutes just one cluster. Second, the Bermuda cluster (Fig. 1) appears structurally related to the underlying oceanic basement—trending parallel to crustal isochrons (as is also true for the Bermuda

Rise—see below): By contrast, neither the overall trend of the chain, nor the trends of any clusters or subclusters, bears any relation to the pre-existing plate tectonic fabric as defined by magnetic lineations and fracture zones (Fig. 2 of Vogt and Smoot, 1984). Third, almost all the JS guyots (except Makarov Guyot; Fig. 3a of Vogt and Smoot, 1984) exhibit four to six flank rifts per edifice, giving the guyot physiography a starfish character not seen at Bermuda. Finally, the most obvious difference between the two volcanic provinces is post-volcanic subsidence: The JS evidently erupted on the summit of a swell of similar relief as the Bermuda Rise, but this swell has, not surprisingly, in the 100 million years that have passed, long since subsided into the noise. The Bermuda volcanoes and Rise are only half the age of the JS, so would not be expected to have subsided as much. However, neither Bermuda nor the Bermuda Rise appears to have subsided at all since 40 Ma—a major difference with the JS and other oceanic volcanic lineaments and associated swells (as discussed further below).

A major difference between Bermuda and Pacific atolls and guyots is the lack of subsidence of Bermuda (Other Atlantic "hotspot" type volcanic provinces share this lack of subsidence, or have even been elevated; e.g., Holm et al., 2006). However, this difference must relate to the nature and evolution of the Bermuda Rise (discussed below), not to the Bermuda volcanic edifices.

Other seamounts in the Bermuda Rise area

Besides the Bermuda cluster, no volcanoes or other igneous activity of Cenozoic age has been documented anywhere in the western North Atlantic, including the Bermuda Rise (Fig.3). The nearest sizeable seamount on the BR is Muir Seamount (Fig.1), an elongated (more or less transform-parallel, unlike Bermuda) edifice ca. 250-300 km northeast of Bermuda. Like Bermuda, Muir Seamount was constructed on post-M0 crust of ca. 120 Ma age, and is located near the crest of the rise. However, Muir is part of a chain of mainly small and widely separated seamounts, paralleling the much more conspicuous New England Seamount chain, evidently unrelated to Bermuda and therefore not discussed in this paper. At least eight sediment cores (the earliest described in Black, 1964) have been recovered from Muir Seamount (unpublished compilation by L. Raymond, WHOI; B.E. Tucholke, personal communication, 2005). Core-bottom sediment ages range from Pliocene to Late-Cretaceous—the latter ruling out a Middle to Late Eocene age like that inferred for Bermuda (An Late Cretaceous age is shown on the Geologic Map of North America; Reid et al., 2005; Fig. 3). No radiometric dates on Muir Seamount itself are available, so the feature may even be Lower Cretaceous (but of course no older than the Aptian-aged basement).

Bermuda Rise

Residual depth anomalies and crustal age

The NE-SW-trending Bermuda volcanoes rise from the crest of a similarly trending, ca. 1500 km long, 500-1000 km wide basement rise (or swell) called the Bermuda Rise (Figs. 1, 3 and 4; Johnson and Vogt, 1971). The axis and western limits of the rise lie respectively ca. 900-1200 and 600-800 km ESE of the continental/oceanic crustal transition zone. As a bathymetric feature, the Bermuda Rise (hereafter abbreviated BR) is approximately outlined by the 5500 m depth contour (e.g., Fig.1 of Jaroslow and Tucholke, 1994). The rise (swell) is bordered by connected abyssal plains on three sides (Fig. 1) : The Sohm in the northeast, the Hatteras to the west, and the Nares to the south, with depths and source distance increasing in that order (e.g., Pilkey and Cleary, 1986). Besides affecting turbidite depostion patterns, the BR has steered abyssal currents—in geologically recent times the the Antarctic Bottom Water- and influenced sedimentation patterns, e.g., along lower SW (Driscoll and Laine, 1996) and NE flanks (Laine et al., 1994) of the rise.

Detrick et al.(1986) give the dimensions of the BR bathymetric expression as 900 x 600 km, similar to Johnson and Vogt's (1971) 500 x 300 nautical miles (ca. 925 x 550 km). Other papers cite different but similar dimensions. Of course, horizontal dimensions of any kind of anomaly---excepting those with sharp edges—are necessarily hard to pin down, and estimates often depend on the "eye of the beholder"and on what is defined as the background.

Because the BR is clearly a feature of the oceanic crust and upper mantle, its substantial sediment cover (Figs. 3, 4) is of no interest (except as a record of volcanism and uplift, see below) and needs to be "removed" because it affects estimates of residual heat flow and depth anomaly. The total BR sediment cover averages 500 -1000 m in thickness (Fig.4) over most of the central rise but increases to 1000-1500 m on the outer NE, W and SW flanks, while decreasing to 100-500 m over the middle and outer SE flanks (Tucholke, 1986). (Sheehan and McNutt, 1989, cite the thickness "in the Bermuda region [to] vary from 200 to 1300 m"). Rises/swells in oceanic crust far from terrestrial sediment sources, such as many in the Pacific, have accumulated far less sediment than e.g., the BR, making the shape and extent of the features there much easier to estimate from bathymetry alone.

Bathymetry and basement isochrons alone (Fig. 1) suggest that oceanic crust underlying the rise (Fig. 1) ranges from ca. 100-105 Ma (roughly midway between chrons M-0R and 34º N) to about 140 Ma (roughly midway between M-0R and M-25N). Upon allowance for isostatic crustal depression by sediments (e.g., Crough, 1983a), plots of basement depth vs. crustal age suggest the BR depth anomaly (Fig. 1, bottom, and 5) extends from about 80-100 Ma crust west to 150-160 Ma crust, and along the axis of the rise a distance ca. 1700-2000 km, from about 25-27º N to 37-38º N (Figs. 9 and 10 of Sclater and Wixon, 1986). The depth anomaly is somewhat asymmetrical in transverse section, with steeper southeastern vs. northwestern slopes.

The BR basement between anomalies M-0 and 34 was formed at higher average spreading half-rates (ca. 20 mm/a; recomputed, using the Gradstein et al. (2004) time scale, from - Klitgord and Schouten, 1986) than crust formed after 34 or from ca. 135-140 Ma until M-0. However, any variation of spreading rates during the long normal polarity chron between M-0 and 34 is unconstrained by magnetic lineations or other data. It is thus unknown if spreading rates increased at M-0 time or somewhat later. Generally, oceanic crust formed at faster spreading rates is smoother. As a dramatic example, the basement on the upper western flanks of the BR becomes sharply smoother west of a slightly time-transgressive boundary ca. 135-140 Ma in age (Fig. 4; Sundvik et al., 1984), a time when half-rates declined to ca. 7- 10 mm/a (Klitgord and Schouten, 1986). Despite the presumably higher spreading rates, the basement on the eastern flanks of the Bermuda Rise does not become dramatically smoother (e.g., Vogt and Johnson, 1971; Jaroslaw and Tucholke, 1994 and Lizarralde et al., 2004). However, existing detailed mapping along the Kane FZ corridor (Vogt and Tucholke, 1989) and earlier flow-line profiles to the south (Vogt and Johnson, 1971) suggest that topography in the western ca. 30-50 % of the M-0 to 34 interval ("Cretaceous Quiet Zone") is rougher than in the eastern part, and was therefore probably formed at spreading rates less than the average for the interval. This western part of the Cretaceous Quiet Zone also exhibits fewer fracture zones (Fig. 2), based on their gravity expression.In its trend and extent, the Bermuda Rise thus approximately corresponds to the region of roughest basement (Fig.4), an observation whose possible significance is discussed later.

There is no indication from the character of the basement reflections or from within the sediment column, or from the well-recorded magnetic lineations, that intrusive or extrusive activity, if any, penetrated the crust subsequent to its formation by seafloor spreading. The only exceptions are a few scattered seamounts and the Bermuda volcanoes (discussed above). Furthermore, refraction and wide-angle reflection data from the Bermuda Rise (e.g., Purdy and Ewing, 1986) show no evidence that the oceanic crust under the BR is in any systematic way different from—and specifically thicker than-- other oceanic crust of comparable age and spreading rate of formation.

Upon correction for isostatic loading by sediments, the residual depth anomaly, developed in old (from 80-100 to 150-160 Ma, see above) oceanic crust, reaches 800-1000 m (Fig. 1, top ; Fig. 5) in the region around the volcanoes (Sclater and Wixon, 1986; Detrick et al., 1986; Sheehan and McNutt, 1989; Vogt, 1991). However, the precise magnitude and extent of the anomaly depends on model assumptions and parameters, for example sediment thickness and density structure (Crough, 1983), crustal thickness and density structure, and more generally what is considered to be "normal" for the crustal ages represented on the Bermuda Rise. For example, if basement depth is compared to the Parsons and Sclater (1977) boundary-layer model, the anomaly on the BR crest is about 1800m (Sclater and Wixon, 1986; Jaroslow and Tucholke, 1994). An independent estimate for the **minimum** BR uplift can be obtained from the present elevation of formerly abysssal plain pre-BR turbidites on vs off the Rise: The values so obtained (+400m at DSDP Site 387 and +700m at 386, discussed further below) are consistent with other estimates. The formation of the BR displaced about 0.5 to 1 million km^3 of water, which should have increased global sea level ca. 2 m.

Several studies, starting with Davis et al. (1984), have concluded that the corrected basement **even off the Bermuda Rise** is significantly **too shallow**, compared to predictions of both the plate and the cooling half-space (boundary layer) models. Sclater and Wixon (1986) found the average depth anomaly in the western Atlantic (including the BR) to be about +200m relative to the plate model. Detrick et al. (1986) infer a regional depth anomaly of about 300-400 m relative to the plate model, and 1000 m relative to the boundary layer model.

However, if the two flanks of the North Atlantic south of 52º N are compared, the basement, corrected for sediment loading, averages 557 m (i.e., ca. 500 m) **deeper** west of the Mid-Atlantic Ridge vs. to the east (Conrad et al., 2004). This result—inconsistent with earlier studies, e.g., Fig. 2 of Colin and Fleitout (1990) — is independent of lithosphere cooling models, and was attributed by Conrad et al. (2004) to continued sinking of the "Farallon slab", which based on seismic tomography is thought to underlie the western Atlantic at great depths in the mantle. However, those authors made no effort to remove the effects of mid-plate swells, which are more numerous and better developed east of the Mid-Atlantic Ridge (e.g., the Cape Verde Rise) than the Bermuda Rise is to the west. At any rate, the interpreted existence of the Farallon slab below the far western North Atlantic is speculative.

Geoid and gravity anomalies and their relation to topography

Associated with the BR depth anomaly is an oval geoid high (Fig. 1, top; Fig. 5) reaching 5-10 m in amplitude, depending on the model (Crough, 1978; Haxby and Turcotte, 1978; Detrick et al., 1986; Sheehan and McNutt, 1989), and a corresponding gravity high, measured relative to the adjacent seafloor, of about +10 mGals (Rabinowitz and Jung, 1986; Sclater and Wixon, 1986). Differences in the calculated geoid high (e.g., 8 ± 2 m; Detrick et al., 1986, vs 5 ± 2 m; Sheehan and McNutt, 1989; Anderson, 2005 cites +5.5 m) reflect different methods of filtering the raw geoid.

More locally, the Bermuda edifice generates the largest Free-Air gravity anomaly (+350 mGal) in the Atlantic Ocean (Rabinowitz and Jung, 1986), with a local geoid high, above the regional BR high, of +7 m (Fig. 5; Sheehan and McNutt, 1989). The latter authors derived the same the elastic thickness (30 ± 5 km) for the mechanical lithosphere downflexed under the "positive load" of the Bermuda volcanoes by two different methods: 1) the admittance (cross-spectral ratio of observed geoid to observed topography in the wave number domain) and 2) linear filters. Black and McAdoo (1988) had previously used the admittance method to compare various seafloor topographic features, including the Bermuda Rise area, and to test whether the admittances at longer wavelengths better fit uplift by convection or by plate temperature.

Both admittance and linear filtering methods of Sheehan and McNutt (1989) yield similar compensation depths for the "negative load" under the Bermuda Rise (50 ± 5 km and 55 ±10 km). Prior derivations of BR compensation depth, which ignored wavelength dependence of the anomalous geoid/topography ratio, varied from 40-70 km (Crough, 1978) to 80 km (Cazenave et al., 1988) and 100 km (Haxby and Turcotte, 1978; repeated in Turcotte and Schubert, 2002).

Seismicity and Stress

The state of stress in any part of the crust and underlying mechanical part of the mantle lithosphere can be established by a number of techniques, including borehole elongation and earthquakes of sufficient magnitude to allow determination of first motion. In the western North Atlantic, Zoback et al. (1986) reviewed historical and instrumentally recorded earthquakes and suggested that mid-plate seismicity is anomalously high in the crust and/or upper mantle below the Bermuda Rise, compared to surrounding oceanic crust. When augmented with epicenters from events postdating their paper (Fig. 4), the anomalously high Bermuda Rise mid-plate seismicity is reinforced. This is further supported by historical reports of earthquakes felt at least as far back as the late nineteenth century (Sieberg, 1932; Zoback et al., 1986), with an 1883 event of Modified Mercalli level VI to VII relayed in the latter paper. Of course, such earlier observations would have been biased towards shocks occurring in the vicinity of Bermuda.

Three earthquakes in the BR area (labeled A, B and C in Fig. 4) were large enough to allow determinations of fault motion and/or focal depth (e.g., Nishenko and Kafka, 1982; see review by Zoback et al., 1986 and Zhu and Wiens, 1991). The largest and most studied event (A, m_b 6.0; 24 Mar 1978), ca. 380 km southwest of Bermuda, involved thrust faulting at a focal depth of about 8 km. A m_b 5.1 shock (B) ca. 300 km east of Bermuda (24 Nov 1976) showed strike-slip faulting at a focal depth of 10 km, based surface wave amplitudes (Nishenko and Kafka, 1982). The 9 Dec 1987 m_b 5.2 event (C), located about 500 km northeast of Bermuda, involved thrust faulting with a strike-slip component; its focal depth could not be accurately determined. The preferential occurrence of earthquakes in the area of the Bermuda Rise (Fig. 5) shows that the mechanical lithosphere under the rise is under stress, but what is the source of this stress?

The net stress within any part of the mechanical lithosphere, such as under the BR, reflects the combined effects of a) ridge push resulting from oceanic topography (i.e., the Mid-Atlantic Ridge), and greatest at shallow depths in the plate; b) thermal stresses remaining from initial cooling; c) stresses resulting from loading of volcanic edifices, e.g. the Bermuda volcanoes, most effective near the edifices, or sediments; c) membrane stresses due to the change of local curvature; d) basal drag due to motion of the plate with respect to the underlying asthenosphere; and e) thermoelastic stresses due to hotspot reheating. The latter effect –in the assumed absence of other forces--was modeled by Zhu and Wiens (1991), who concluded that if positive temperature anomalies exist, the resultant thermoelastic stress "may be an important contributor to stress fields near hotspots in old oceanic lithosphere". Their modeling predicts a complicated pattern of stresses, with maximum deviatoric stress "on the order of 100 MPa, covering a broad area". Their models further predict extension at depths less than 10-15 km, and compression from there to the base of the mechanical lithosphere. Zhu and Wiens (1991) list three Bermuda Rise earthquakes (labeled A, B and C in Fig.4), whose P axes "suggest horizontal compressional stresses oriented radially with respect to the Bermuda swell, in agreement with our modeling". They also modeled the thermoelastic stress field for the Cape Verde Rise, finding the two first-motion solutions for earthquakes there also in approximate agreement with model predictions.

However, Zhu and Wiens (1991) note that compressive stresses should prevail only deeper than 15 km, in contrast to the 8 km focal depth of the 1978 m_b 6.0 event on the Bermuda Rise (A in Fig. 4), the most energetic shock instrumentally recorded on the rise. The authors suggest that the Bermuda hotspot may have had enough time to heat the upper part of the plate—so as to account for the above discrepancy between model and observation—is not supported by heat flow data discussed at length above.

Alternatively, the BR earthquakes could be occurring largely in response to a NE to ENE oriented maximum horizontal compressive stress, which dominates the central and northern Atlantic in all lithosphere older than about 20 Ma (Zoback et al., 1986) and which is largely due to the ridge-push force. As shown in Fig. 2 of those authors, the stress pattern implied by BR earthquakes is largely consistent with this regional stress field, vs. a more local thermoelastic field as modeled by Zhu and Wiens (1991). The relative concentration of mid-plate seismicity on the Bermuda Rise may at least in part reflect a thinner, more fractured (hence weaker) crust and more gabbro-rich (hence weaker) upper mantle associated with the crust formed at very slow spreading rates (i.e., in the "rough basement" belt in Fig. 4) as deduced by Lizarralde et al. (2004) and discussed in a previous section. However, whatever process formed and maintains the BR must still play some role in the overall pattern of strain release (Zoback et al., 1986)—otherwise enhanced seismicity would be spread northeast and southwest beyond the limits of the rise.

Zoback et al. (1986; See their Fig. 3) further suggested that "many of these [teleseismic] events appear to be located near fracture zones and suggest a simple causal relationship". Given the small data set and the large epicenter location errors, those authors were reserved about this correlation. Remapping fracture zones by satellite radar altimetry-derived Free-Air gravity anomalies, and adding newer epicenters (courtesy of the National Earthquake Information Center), we can now be somewhat more certain that Zoback et al. (1986) were correct in suggesting that BR earthquakes tend to occur along the traces of ancient fracture zones (Fig. 4). However, this observation does not explain how the processes forming the BR promote ongoing fracturing in the oceanic crust or uppermost mantle below the rise.

Heat Flow

If anomalously elevated oceanic crust such as the Bermuda Rise has a thermal origin (e.g., Crough, 1978, 1983b and later authors) then it may, depending on the specific model (discussed further later), exhibit anomalously high heat flow. Heat flow (Fig. 1) has thus come to be considered an important geophysical constraint on the nature and origin of the mantle processes responsible for Bermuda and the BR, motivating several expeditions to acquire new data from the Rise and surrounding parts of the western Atlantic (e.g., Davis et al, 1984; Detrick et al., 1986; Louden et al.,1987), as well as reanalyses of extant databases (e.g., Sclater and Wixon, 1986; Nagahira et al.,1996). Local heat flow variability, typically on horizontal scales of ca. 5-6 km (e.g., Detrick et al., 1986; Harris et al., 2000) mandates that each heat flow "value" used for modeling a large-scale process like the BR be obtained by averaging a number of close spaced measurements of both gradient and thermal conductivity, with local basement topography and sediment thickness mapped by seismic reflection data for the purposes of modeling and correcting for the effects of sedimentation rates, spatially varying conductivity structure, and porewater convection within the sediments and/or basement (e.g., McNutt, 2002; Von Herzen, 2004). The data discussed here (Fig. 1) all represent averages of many individual penetrations—older, single-penetration values, of dubious value, are not shown.

Starting with Langseth (1969), and as late as the paper by Sclater and Wixon (1986), no significant heat flow high, predicted by some models, was resolvable in archival heat flow databases. However, Detrick et al. (1986) measured heat flow at additional stations , which, combined with earlier data and arranged in a ENE-trending transect from the western Bahama Islands to Bermuda, did reveal a modest but significant (8-10 W/m^2) heat flow high over the rise (Fig. 1). The average heat flow was found to be 57.4 ± 2.6 mW/m^2 on the rise, vs. 49.5 ± 1.7 mW/m^2 off the rise (Detrick et al., 1986). However, local spatial variability (notably at the "54.4mW/m^2" site closest to Bermuda, where individual penetrations returned values from ca. 37 to 73 mW/m^2) raises the possibility that the "spatial coverage [may be inadequate] to define the mean heat flux" (Detrick et al., 1986, p.3709). Noting the correlation of local heat flow with basement topography, those authors tried, but failed, to account for the heat flow variability by modeling the heat flow refraction effect predicted from pure thermal conductivity. Effects of rough but buried basement topography could from molecular conduction alone account for local values ranging up to 25 % above the regional mean (Von Herzen, 2004).

With the benefit of more recent studies along the Hawaiian moat and swell (e.g., Harris et al., 2000; McNutt, 2002), we speculate that the high local variability observed close to Bermuda reflects porewater flow within the spatially varying thickness of Bermudan volcanogenic debris (discussed above; Seismic Reflection Horizon A^v; Fig. 1, top; e.g., The Shipboard Scientific Party, 1979, Tucholke and Vogt, 1979; Tucholke and Mountain, 1979). The volcanogenic debris—comprising turbidites—is thicker in valleys. If these sandy sediments serve as recharge zones (downward percolating seawater), the thermal gradients above them would be reduced, thereby perhaps accounting for the apparent correlation of heat flow with basement topography noted by Detrick et al. (1986).

Even without allowance for the effects of the Bermudan volcanogenic sediments, the generally rougher basement topography of the crust which includes the BR (e.g., Sundvik et al.,1984; Fig. 3 of Zoback et al., 1986)) would favor higher heat flow via porewater advection (over basement peaks) in the more permeable basement. Von Herzen (2004) estimates that thermally significant hydrothermal circulation has influenced ca. 20-30 % of heat flow stations on oceanic crust older than 95 Ma. Anomalous heat flux would be expected to extend out to older crust for slower spreading rates (i.e., rougher basement topography), and would involve porewater advection of the order 10 cm/yr (Von Herzen, 2004). Although none of the BR heatflow measurements to date show large deviations from the regional mean, some of the most anomalous (up to an order of magnitude above the regional mean) heat flow values on old crust anywhere lie just south of the BR, on the Nares Abyssal Plain (Embley at al., 1983; Von Herzen, 2004) and are almost certainly the result of porewater advection.

Heat flow measured in the 1972 Bermuda island borehole (53-57 mW/m^2; Hyndman et al., 1974) is comparable to the average value returned from the three ocean floor sites within ca. 200 km of Bermuda (Fig.1, bottom; Detrick et al., 1986) even if radioactive heat sources in the Bermuda volcano (specifically the lamprophyre sheets) contribute up to 4 mW/m^2 to the borehole value as suggested by Hyndman et al. (1974). Borehole temperatures—measured at 8 m intervals to a depth of 360 m—showed a linear increase with depth; the modest gradient and lack of irregularities are consistent with long-term volcanic and hydrothermal quiescence.

In a later study, Louden et al. (1987) found relatively high (53 mW/m^2, 25 % higher than predicted by either plate or cooling half-space models) even on old (ca. 163 Ma) crust northwest of the Bermuda Rise (Fig. 1). Comparing this heat flow with other values in the region, they suggest "the entire Mesozoic northwest Atlantic may have a uniformly elevated heat flow". Louden et al. (1987) corrected all heatflow values for this region by using a higher sediment conductivity for the station on anomaly M-0 (Fig. 1) and by allowance for lithospheric cooling with age (as predicted by the plate model). This reduced the apparent heat flow high on the Bermuda Rise to a mere 5 mW/m^2, compared to the 8-10 mW/m^2 derived by Detrick et al. (1986), and so far only demonstratable on the SW flank of the rise (Louden et al., 1987).

And, while the heat flow southwest of the Rise has apparently reached an equilibrium value (ca. 48-50 mW/m^2), this is ca. 15 % higher than expected for crust of this age, and the basement depths are at least 200-300 m too shallow. These anomalies were first noted by Davis et al. (1984) and further discussed by Detrick et al., 1986, and Nagahira et al., 1996). While the depth and thermal anomalies over supposedly "normal" ocean crust may cast doubt on the interpretation of data from the rise itself, it may be that the geographic "extent" or influence of the process (discussed below) that formed the BR affected a wider area of crust than suggested by the basement swell dimensions. For example, while Nagahira et al. (1996) deliberately excluded heat flow stations on the BR (as defined by them), they concluded that stations within 600 km of the BR crest suffered the greatest amount of thermal rejuvenation, while those least reheated are 900 km distant.

Except for the Bermuda borehole data of Hyndman et al. (1974), all the above measurements were made with shallow-penetration probes. For example, Detrick et al. (1986) used either 3.5 m or 7 m probes, depending on bottom sediment penetrability. Louden et al. (1987) employed a 4 m long probe. Even though temperature gradients were shown to be relatively constant along these thermistor strings, there is always nagging doubt about the effects of long-term (many centuries and longer) bottom water temperature changes, and other effects which may have biased the results. These doubts have been reinforced by a comparison (Ruppel, 1996; 2000) between deep (320-420 mbsf) ODP Leg 164 borehole sediment temperatures in the SW North Atlantic (Blake Ridge area) and shallow penetration data in the same region—underlain by oceanic crust estimated to be 175-180 Ma in age. The borehole data suggest heatflow values of about 35 mW/m^2, compared to 48-55 mW/m^2 for the shallow-penetration data (e.g., Fig. 1). This result casts some doubt on the putative anomalously high western Atlantic heatflow—or at least its magnitude-- inferred by Davis et al.(1984), Detrick et al. (1986), Louden et al. (1987) and Nagahira et al. (1996). Ruppel (1996) concluded that "basement heat flux on the Blake Ridge may be up to 25% lower than the previous estimate of 49 mW/m^2". Given the various error sources, this apparent "excess" heatflow is comparable in sign and magnitude to the result of Detrick et al. (1986) who found heat flow off the Bermuda Rise to be about 15 % higher than expected. Similarly, Louden et al. (1987) concluded that "heat flow values are elevated by 20-30 % throughout the entire region compared to standard one-dimensional conductive cooling models, but are not systematically related to basement depths". Of course, additional borehole thermal data are needed to demonstrate that the discrepancy found by Ruppel (2000) applies to the entire region. However, her borehole results do support a primarily conductive heat transfer, at least in the borehole area.

Deep structure below the Bermuda Rise—Evidence from seismic tomography

High-resolution whole-mantle seismic tomography is still in its infancy (see e.g. Nataf (2000) for a global synthesis of seismic searches for mantle plumes, and Zhao, 2004; Montelli et al. 2004 and van der Hilst and De Hoop, 2005 for more recent analyses). However, several studies have inferred the existence of the "Farallon Slab" deep below the region from the eastern US to the western Bermuda Rise at depths increasing from ca 800 km in the west to ca. 2500 km in the east (e.g., Fig. 19 of Zhao, 2004). As noted previously, Conrad et al. (2004) attributed the generally deeper western Atlantic (vs. the eastern) to the effects of this deep slab. Subducting slabs are now thought to become stagnant at the 670 km discontinuity—and only much later detach and sink into the lower mantle (e.g. Fukao et al., 2001). In the upper mantle, modestly slower (by ca. 0.2-0.5%, perhaps not significant) P-wave speeds extend to about 400 km directly below Bermuda on an EW vertical cross-section (Fig. 19 of Zhao, 2004), suggesting that anomalously warm mantle associated with the BR may extend that deep. Similarly anomalously "slow" upper mantle extends to ca. 1000km depth under a broad region extending from the western BR to the eastern margin of North America. In the shear-wave speed tomography of Ritsema (2005), the mantle below Bermuda is fast at 100 km depth (plausibly the result of the thick, old lithosphere), slow in the 300, 600 and 1100 km depth sections, and fast at still greater depth—broadly consistent with the P-wave model of Zhao (2004). However, Ritsema's model does not resolve a separate anomaly under the Bermuda Rise—which can also be said for most other postulated hotspots/mantle plumes (Fig. 6 of Ritsema, 2005). In general the S-wave speed contrast is highest in the upper most mantle and lowest below ca. 600 km depth.

At the very bottom of the mantle (2850 km, layer D"), most of the North American plate, including the western North Atlantic, generally is "fast", both in shear-wave (Ritsema, 2005) and compressional-wave (Dziewonski, 2005) speeds.

Uplift History of the Bermuda Rise: The Sedimentary Record

Deep drilling on the Bermuda Rise at DSDP Sites 386 and 387 (Fig. 1) recovered numerous mixed bioclastic/terrigenous turbidites that may well have originated along the North American continental margin, with basement uplift—presumably the formation of the BR--shutting off this sediment source some time in the middle Eocene (whose age range is given by Gradstein et al., 2004, as 48.6- 37.2 Ma)(Tucholke and Vogt, 1979; Tucholke and Mountain, 1979). The top of this turbidite unit is marked by seismic reflector A^t, whose seaward extent is shown in Fig.4. The offlapping relation for the youngest turbidites—i.e., the cessation of turbiditic sedimentation happened earlier on what would become the crest of the BR than farther on the flanks—is tentatively suggested by the "younging" of the youngest turbidites from east (middle Eocene at Site 386) to west (post-late Eocene at DSDP Site 8). Minimum (i.e., neglecting possible offlap effects) total uplifts of the Bermuda Rise (700 m at Site 386 and 400 m at Site 387) can be estimated from the present BR elevations of originally presumably nearly horizontal (i.e., abyssal plain) A^t (Tucholke and Vogt, 1979).

At least at DSDP site 386, near the crest of the rise, these turbidites comprise largely biogenic sediments, so whereas an origin along the continental margin—some 1200 km distant--is likely, it is not absolutely certain. At least some turbiditic sediments might also have been derived from adjacent basement highs. If the end of turbidite deposition marks the uplift of theBR, its age immediately preceded, or was coeval with, Bermuda volcanism, because the oldest certain volcanogenic debris shed from the islands (discussed previously) arrived at Site 386 immediately upon cessation of biogenic turbidite deposition.

The BR is crossed by fracture zones, some of which have deep basement valleys that continue for long distances. At least on the central parts of the BR and further east, these fracture zones show up as negative gravity lineations, indicating basement valleys (Fig. 2). The Kane FZ is foremost among those, and due to its depth has accumulated a thicker sediment pile than adjoining oceanic crust. The greater thickness of sediments in FZ valleys reflects mass wasting from local basement highs, but also turbidity flows from distant sources. In any case, the sediment record is generally expanded in deeps, allowing better resolution of seismic stratigraphy. Jaroslaw and Tucholke (1994) took advantage of these circumstances and analyzed the sediment fill within the Kane FZ valley—along which turbidity flows moving southeast (down the flank of the BR) traveled as far as 55.3° W (well beyond the BR), where further travel was stopped by a basement obstruction (dam) within the valley floor. The seismic stratigraphy within the valley fill revealed reflectors, generally associated with former abyssal plain surfaces, post-dating the Bermuda volcaniclastic seismic horizon A^v. Because these reflectors must have been nearly horizontal when deposited, but are now inclined (down towards the southeast) by amounts that increase with increasing subbottom depth, Jaroslow and Tucholke (1994) were able to demonstrate that uplift of the Bermuda Rise **continued** into the Miocene, later than had been assumed from earlier observations. Their analysis confirmed that significant uplift of the BR had occurred by early Oligocene times, but that some **400-500 m additional uplift must have occurred after ca. 33 Ma**, the time of the second (lamprophyre or "Bermudite" sheet) igneous episode discussed earlier. Ca. 200 m relative excess uplift was inferred from the middle early Oligocene to the earliest Miocene, with an additional, albeit perhaps not significant, uplift of 45-50 m during the early-to-late Miocene interval. Notably, the seismic stratigraphy in the Kane FZ valley, as interpreted by Jaroslaw and Tucholke (1994), reveals **no evidence for migratory uplift** such as would be expected from relative westward motion of the plate over a mantle hotspot.

Independent evidence for the initiation of BR formation comes from bottom current-controlled sediment deposition, because such currents are steered by bottom topography. In geologically recent times, sediment cores and seismic profiles show a pattern of sediment redeposition caused by the movement of Antarctic Bottom Water around the BR (Laine et al., 1994; Driscoll and

Laine, 1996). Seismic reflection data suggest that bottom currents began to be steered around the BR as early as Middle Eocene (Ayer and Laine, 1982), consistent with other evidence for initiation of BR uplift.

The Southeast Bermuda Deep (SEBD)

A regional bathymetric low, centered about 900 km southeast of the BR crest (Fig. 1), may be genetically related to the origin of the BR (see below), and is therefore included in this review. The feature was discussed both by Sclater and Wixon (1986), who referred to it as a "pronounced counter-balancing low", and by Vogt (1991), who called it the "Anti-Bermuda Rise". We suggest the more appropriate name "Southeast Bermuda Deep" hereafter shortened to SEBD. The residual depth anomaly of this feature (Fig. 4 of Sclater and Wixon, 1986) is below –600 m but does not reach –800 m, the core of the SEBD anomaly being roughly defined by an area where average water depths commonly exceed 6000 m. Because sediment thicknesses average less than 100 m (Tucholke, 1986), the basement depth and water depth contours define the shape of the feature almost equally. Chron 34 bisects the central part of the SEBD, which is developed in crust ranging from ca. 50 to 90 Ma. Sclater and Wixon (1986) only calculated the depth anomaly east to 40W, i.e. approximately out to the crest of the MAR.

Although the amplitude of the SEBD low is only 2/3 of the BR high, and is somewhat smaller in planform, both are of comparable magnitude in extent and amplitude. Furthermore, if the depth anomaly were to be defined on the basis of average corrected basement depths in the western Atlantic, vs. from theoretical plate cooling models, the BR and SEBD would have about the same amplitudes (Sclater and Wixon, 1986).

A geoid low ca. –5 to –6 m is associated with the SEBD, again comparable but opposite in sign to the regional BR geoid high (Sclater and Wixon, 1986). The corresponding Free-Air gravity low is about –10 mGals, relative to "normal" seafloor in the western Atlantic, and about –20 mGals relative to that associated with the BR (Rabinowitz and Jung, 1986). The archival heat flow database shows only widely scattered stations in the SEBD area. Although these give no indications of a HF low, this result may well be inconclusive.

The SEBD appears to have no seamounts large enough to be resolved in the Smith and Sandwell (1997) bathymetric grid. The finer-resolution seamount (including features less than 1km in relief) location map of Epp and Smoot (1989) does not extend far enough southwards to cover the SEBD.If the SEBD is related to the Bermuda Rise (e.g., Sclater and Wixon, 1986), it presumably, like the BR, developed in pre-existing oceanic crust during Eocene times. However, it is much more likely that the SEBD is unrelated to the Bermuda Rise—reflecting instead formation of anomalously deep oceanic crust at the ancient MAR axis. This is supported by the continuation of relatively deep crust east to the MAR axis, and to the existence of a conjugate region of anomalously deep crust of comparable age east of the MAR. Conjugate deep areas east of the MAR (generally more than 300 m deeper than normal for the ages, but depending on the age-depth model) appear in the global depth anomaly maps of e.g., Colin and Fleitout (1990), Kido and Seno (1994) and DeLaughter et al. (2005). The North Atlantic depth anomaly, calculated from Model GDH2 and kindly shared by the latter authors, is shown in Fig.6. The depth anomaly would probably be even more symmetric about the Mid-Atlantic Ridge, had it not been affected by Tertiary swell development that formed and probably continues to maintain (Courtney and White, 1986) the Cape Verde Rise.

We therefore consider it fairly safe to dismiss the interpretation (e.g., Sclater and Wixon, 1986; Vogt, 1991) of the SEBD as the downwelling counterpart of the Bermuda Rise.

Is there a Bermuda Hotspot Trace?

When Morgan (1972) extended J.T. Wilson's hotspot ideas to propose relatively fixed mantle plumes, he attempted to account for elongate traces of progressive age-varying igneous activity (e.g., the Hawaii-Emperor seamount chain) and to use these traces to develop a model for "absolute" motion (i.e., relative to a reference frame in the deeper mantle) of lithospheric plates. The isolated Bermuda volcanoes, perched on the crest of the broad BR , did not fit this pattern and probably for that reason were not discussed in the early hotspot/mantle plume literature. This well-mapped part of the Atlantic shows no evidence of the seamount chain that should exist, younging towards the east and becoming older to the west, by the "stationary hotspot" or plume model (e.g., Figs.1, 3). In fact, the four volcanoes of the short Bermuda cluster are oriented at right angles to the expected trace, as the BR is also. Furthermore, the uplift history—most recently clarified by Jaroslaw and Tucholke (1994)—shows continuing uplift centered on the BR from the middle Eocene at least into the earliest Miocene. No evidence was found by them for eastward migration, pedicted by the hotspot model, of the uplift center. In fact there is little or no evidence for any kind for eastward migration.

All three hotspot models (Morgan, 1983; Duncan, 1984; Mueller et al., 1993) concur in predicting a present location of a relatively fixed "Bermuda hotspot" at around 32º N, 58º W, ca. 650 km east of the islands, on the outer edge of the present BR. This predicted hotspot location is ca. 1450 km from the closest point (ca. 26º N, 45º W) on the present MAR spreading axis.

The present bathymetry over the region that should have been crossed by a Bermuda hotspot trace from ca. 40 Ma to the present (Fig.7) shows a small seamount at 31.1º N, 58.2º W, not too far from the predicted present hotspot location. Although lack of data on this feature precludes ruling out the young age expected if this seamount reflected present activity of a Bermuda hotspot, the latter appears extremely unlikely: No earthquake seismicity has been reported there (Fig. 4): Of the three BR area earthquakes summarized in Zoback et al. (1986) the nearest to the "predicted hotspot" is a m_b=5.1 reverse faulting event ca. 350 km to the west (33.01º N, 61.66º W), and no seamounts exist between the small seamount and Bermuda (Fig.7). Most probably the small seamount is Cretaceous in age, forming the southeast end of a sparsely populated NW-trending chain that includes the relatively large Muir seamount north of Bermuda.

Sleep (2002) proposed that when plumes are sufficiently close to a spreading axis, the buoyant material flows towards the axis rather than erupting locally, thus forming a gap in the hotspot track. On page 2, Sleep wrote: "The Bermuda track is a less clear example of a ridge-approaching hot spot with an unclear on-axis hot spot between 20 N and 28 N…..[The] activity has ceased near off-axis Bermuda…but has not yet clearly commenced at the axis". We are skeptical about this explanation, in the absence of any evidence for hotspot influence along this section of the MAR. If anything, the MAR is deeper than elsewhere in this region (Fig. 6).

The analysis of BR depth and geoid anomalies (Sheehan and McNutt, 1989) shows a modest southeast displacement of the rise culmination relative to Bermuda (Fig.5). Inspection puts the summit of depth anomaly of the Rise (disregarding Bermuda) at ca. 31.7º N, 64.0º W, 75-100 km southwest of the Bermuda edifice; the summit of the geoid anomaly is near 31.0º N, 63.3º W, ca. 200 km from Bermuda (Sheehan and McNutt (1989) noted this offset, estimating it at 150 km). If these offsets represent the distance the plate has moved since 40 Ma relative to the Bermuda mantle source below the lithosphere, the average rates of migration have only been of the order 0.2-0.5 mm/a, with the implied plate motion towards the NW, vs. W or SW predicted by hotspot models (Figs.1,7). However, this type of calculation assumes the edifice was originally built on the summit of the BR, a plausible but probably untestable assumption. The offset might equally be attributed to the outward spread of the swell root, with magmas utilizing a NE-SW trending zone of weakness (based on the model of Phipps Morgan et al. (1995))

Nevertheless, longer-term "absolute" plate motion models can readily be applied to any spot on any plate so as to predict the trace that point would have made over the mantle below the plates. Morgan and Crough (1979) first applied their plate motion model to a postulated "Bermuda hotspot". They could only constrain their model at one place (Bermuda) and time (40-

45 Ma, as discussed above), i.e., postulating that this hotspot affected the upper crust only for a geologically short interval of time, during which the Bermuda volcanoes and the BR were formed. This hotspot trace postulated a present hotspot location on the lower east flank of the Bermuda Rise, and a track which increased in age towards the west, passing over the area of the "Cape Fear arch", a NW trending structural high along the Carolina coast, about the time this arch was thought to have originated (Morgan and Crough, 1979). However, Winker and Howard (1977) showed that uplift of the arch continued into the later Neogene, and newer studies (e.g., Stewart and Dennison, 2006) show the Cape Fear arching to have occurred too late (late Oligocene to mid-Miocene) to correpond to the predicted time (ca. 50-60Ma; Fig. 8) the area passed over the putative hotspot. Moreover, subsequent plate motion models have predicted a more southwesterly track for the putative Bermuda hotspot (e.g., Morgan, 1983; Duncan, 1984; Mueller et al., 1993), The Cape Fear Arch consequently disappeared from later treatments of the "Bermuda hotspot" idea. Furthermore, the denudation history of the Appalachians—as revealed by sedimentary deposits along the US middle Atlantic margin (Poag and Sevon, 1989)--- does not support any uplift during the time the putative Bermuda hotspot track should have crossed this area (Paleocene).

However, Morgan and Crough (1979) also correlated the middle to late Cretaceous alkalic and ultrabasic igneous activity—particularly kimberlite pipes-- in central Arkansas and around the triple state junction of that state and Louisiana and Mississippi with the predicted position of the North America plate over a fixed "Bermuda hotspot" during that time. This correlation has been further developed in several subsequent papers, from Crough et al. (1980) to Cox and Von Arsdale (1997, 2002). However, as McHone (1996) has emphasized, post-breakup Jurassic and Cretaceous magmatism along the eastern US margin, along and west of the Appalachians, and in New England scarcely fits the basic model of the North America plate passing over even several plumes. Although the hotspot models of Morgan (1983), Duncan (1984) and Mueller et al. (1993) all predict somewhat different traces for the "Bermuda hotspot", particularly prior to 60 Ma, all three pass through the general area of this igneous province at about the right time, with Duncan's providing the best geometric fit (Fig.8; Cox and Van Arsdale, 2002). Because the three hotspot models do not include any data from Bermuda or its putative trace, agreement between prediction and observation could only be considered either support for the fixed hotspot model or fortuitous, but not circular reasoning of any sort.

More recent compilations by Cox and Van Arsdale (2002) indicate the igneous activity (including kimberlites, syenites, peridotites, lamprophyres and carbonatites) increases in age in the direction and approximate rate predicted by the three predicted hotspot traces (Fig.8). The oldest intrusions date to ca. 115 Ma in eastern Kansas and the youngest (ca. 65 Ma) occur in central Mississippi. Cox and Van Arsdale (1997, 2002) also attribute the coeval mid-Cretaceous uplift (ca. 1-3 km) and subsequent erosion of the Mississippi Embayment to the "Bermuda hotspot". Nunn (1990) similarly attributed the lesser Sabine Uplift, immediately southwest of the Mississippi Embayment uplift, to the Bermuda hotspot.

Noting the lack of evidence for "Bermuda hotspot" activity anywhere between the Bermuda Rise and Mississippi, Cox and Van Arsdale (2002) speculate that the Cretaceous igneous activity and uplift "may have been the result of increased hotspot flux of the typically weak Bermuda hotspot during the Cretaceous superplume mantle event (ca.120-80 Ma)".

The youngest known igneous activity in the southern-central US (central Mississippi) is centered about 32.25º N, 90.3º W (Fig.8), ca. 2400 km west of the Bermuda volcanoes (32.3º N, 64.8º W). Even if the putative hotspot trace comprised a great circle, the distance covered in ca. 25 my (between ca. 65 Ma and ca. 40 Ma) implies an average rate of motion of ca. 10 cm/a for this point on the North America plate, a high rate similar to the late Tertiary migration rate of Hawaiian volcanism, which is generally attributed to rapid Pacific plate motion over a relatively fixed Hawaii hotspot. However, there is now compelling evidence that the ca. 81-47 Ma age progression along the Emperor Seamounts, formed roughly (considering dating and other errors) over the same time as the 65-40 Ma "fast motion" Bermuda interval, can be explained only by rapid motion of the Hawaii "hotspot" (locus of melt extraction) (Courtillot et al., 2003; Tarduno, 2005). By contrast, the distance between the ca. 65 Ma Mississippi igneous rocks and the ca. 115 Ma eruptions to the NW in Kansas (ca . 40º N, 96º W, based on Fig. 1 of Cox and Van Arsdale, 2002) is 1000 km, implying an average migration rate of only 2 cm/a.

The high rate of average North American plate motion from 65 Ma to 40 Ma required by the separation of 40 Ma Bermuda and the 65 Ma Mississippi igneous activity is inconsistent with predicted ages along the late Cretaceous-early Tertiary parts of predicted hotspot traces. For example, the Duncan (1984) model trace, which best fits the igneous events of Cox and Van Arsdale (2002), predicts that the Bermuda hotspot should have been located near 32º N, 76º W at 65 Ma. This is 600-700 km east of the central Mississippi igneous centers, a significant misfit even if the hotspot is assumed to be a broad area of upwelling.

A major difficulty with the above attempts to link widely separated igneous activity and uplifts to the same, fixed hotspot or plume in the mantle is that tens of millions of years (ca. 25 my for Bermuda) have to pass between "pulses" of activity. I.e., we are asked to imagine a "plume" resembling a lava lamp that only rarely produces a hot blob. In the case of the igneous activity and uplift in the Mississippi Embayment area, earlier and later igneous and tectonic events in the same region must then be explained in other ways—for example, Triassic and Lower Jurassic igneous rocks are attributed to rifting in the Gulf of Mexico; late Paleozoic intrusions are found around the north end of the Missisissippi Embayment (Cox and Van Arsdale, 2002). Meanwhile, the Kansas igneous rocks were intruded from a region of basement faults of the Proterozoic Central North American Rift System, while the late Cretaceous intrusions could have followed reactivated faults of the Late Proterozoic-early Paleozoic Mississippi Graben. To account for such correlation with older underlying fault systems, hotspot proponents like Cox and Van Arsdale have had to postulate that igneous intrusions of hotspot origin can only penetrate the crust when and where a hotspot happens to underlie such structures. It might be simpler to attribute pulses of mid-plate tectonic and igneous events to episodes of increased stress, associated with reorganizations of plate motion. In any case, no-one attributes the historic and ongoing seismicity and tectonics in the New Madrid area to an underlying hotspot!

Finally, attempts to correlate the Mississippi-Kansas igneous belt (extending from the 65 Ma and 115 Ma points in Fig. 8) to a Bermuda hotspot trace must necessarily and somewhat arbitrarily disregard or attribute to another hotspot the ca. 49-35 Ma igneous activity in Virginia/West Virginia (Fig. 8; Fullagar and Bottino, 1969; Southworth et al., 1993; Tso and Surber, 2002) as well as the ca. 100 km wide, 400 km long Late Cretaceous (87-77 Ma) Balcones igneous belt in southwestern Texas (Fig. 8; Griffin et al., 2005). In the western Balcones (Uvalde County), $^{40}Ar/^{39}Ar$ dating suggests igneous pulses at 82-80 Ma and again at 74-72 Ma (Miggins et al., 2004). The NE-SW trend and lack of age progression in with Balcones belt are inconsistent with predicted North America-mantle motion in the late Cretaceous, and scarcely consistent with the Bermuda trace. If interpreted according to 'mantle plume' type models, both the Virginia and Balcones igneous episodes reflect isolated mantle blobs. Alternatively they reflect episodes of intra-plate stress (e.g., Anderson, 2002).

Bermuda Rise: Quantitative Models for Current State and Origin

The Bermuda Rise as a bathymetric feature was known for decades before any attempts were made to explain its origin. More or less ignoring or discounting the early 20th century works of Alfred Wegener (1966), most US and Soviet marine geoscientists prior to the 1960s presumably considered the rise a Pre-Cambrian relic of the early earth. The first model put forth specifically for the BR was that of Engelen (1964), who proposed that the Wegener-type continental drift that had opened the North Atlantic in Mesozoic times had exposed the sub-Moho level of the American shield. The pressure reduction in the lee of the drifting continent would have induced, by pressure reduction, a phase transition from eclogite to gabbro, thereby generating an "intumescence" and hence a rise.

After the "New Global Tectonics" had burst on the scene in the later 1960s, Menard (1969) proposed the first modern explanation for "mid-plate rises", attributing them to small transient convection cells or other localized disturbances in the mantle, directly below the lithospheric plates. Models creating mid-plate rises by thermal expansion date from Crough (1978), although these models and the earlier modeling of the Mid-Oceanic Ridge as an effect of thermal expansion (Langseth et al., 1966; Sclater et al.,1971) were prefigured by Alfred Wegener (1966) in the second (1929) edition of his book. Wegener noted that the average depth of ocean basins was less for those basins he had shown opened more recently. In a rough calculation he showed that depth differences of the observed magnitude could easily be explained by higher temperatures and greater thermal expansion of the younger mantle between continents that had separated more recently.

Aside from thicker sediment accumulations, which can be dismissed as a cause of the rise, there are basically only few ways a seafloor swell like the BR (i.e., a basement swell) can have been generated (e.g. Fig.1 of DeLaughter et al., 2005) and some of those can readily be discounted. Seafloor rises/swells can be expressions of 1) isostatically compensated thickened oceanic crust, i.e., similar to large igneous plateaus, aseismic ridges, or smaller volcanic masses, such as the Bermuda edifice; 2) flexural upwarping or forebulge construction due to adjacent loads, such as volcanic edifices imposed nearby on the mechanical lithosphere or ice sheets; 3) reduced density distributed widely within the mantle lithosphere under the bulge, either by lower-density intrusions (A) or/and by higher temperatures (B, i.e., by thermal expansion) or/and (C) by the mantle melting beneath the hotspot; 4) thinning of the lithosphere, e.g. via heating from anomalously hot upwelling asthenosphere or by delamination and sinking of some of the lower lithosphere, and its replacement by hotter material; or 5) dynamic support by upwelling mantle convection under the swell, either attached to the swell or, as in deep mantle plume models, decoupled from it.

Mechanism #1 cannot account for the BR because seismic refraction data show the rise to be underlain by oceanic crust not significantly thicker than elsewhere (Purdy and Ewing, 1986). In fact, more recent seismic experiments show the crust east of the basement-topographic smooth-rough boundary (Fig.4) to be ca. 1.5 km thinner than normal (Lizarralde et al., 2004). Moreover, modeling of the geoid anomaly over the rise (Figs.1, 5,9) clearly indicates a Pratt, not an Airy type of isostatic compensation (Haxby and Turcotte, 1978). Furthermore, known oceanic rises formed of thickened crust generally have well-marked topographic edges and are not broad and gently sloping.

Mechanism #2 was briefly considered by the senior author ca. 1970 (Johnson and Vogt, 1971), with the idea that the BR and the Appalachians could both be flexural upwarps caused by the thick sediment loads along the North American continental margin. However, subsequent studies of e.g., seamounts and islands showed the flexural characteristics of the mechanical lithosphere to be incompatible with the large distances from the BR and the Appalachians to the sediment pile. A flexural or forebulge- type origin for the BR can safely be dismissed.

One variant of mechanism 3), wide-spread intrusions (Withjack, 1979) can also probably be excluded as the sole mode of rise formation, because, as also noted by Liu and Chase (1989), volcanic activity has only occurred in one relatively small area of the BR (the Bermuda volcanoes). However, Phipps Morgan et al. (1995) attributed much of swell elevations to lens-like, lower-density regions of plume-generated partial melting under the lithosphere. Inspired by the latter paper, Holm et al. (2006; their Fig. 20) suggested that melts generated at depth below the root of refractory melt under the Cape Verde Rise must move around it to reach the surface, accounting for the succession of different extruded compositions and the 200 km displacement of the Cape Verde volcanoes from the rise summit. The corresponding 75-100 km displacement of Bermuda from its rise summit (Fig.5) may have a similar origin—whether by this or some other mechanism.

Most of the papers modeling the BR either alone, or along with other swells (rises), were published in the period 1978 to 1990—a “golden age” of rise modeling. The subsequent decline in swell-modeling papers can be ascribed in part to the lack of significantly improved or new data (i.e., diminishing marginal improvement, from the point of view of observables such as depth and geoid anomalies). Marine heat flow research efforts declined, particularly in the US. Moreover, in more recent years the validity of shallow-penetration probe data for inferring deeper thermal structure has been challenged, as noted above (e.g., Ruppel, 2000; McNutt , 2002).

All papers attempting to model the Bermuda Rise in the “plate-tectonic era” have adduced, modeled or tested some combination of mechanisms 3, 4 and 5. The Bermuda Rise, like many other oceanic “mid-plate swells”, represent departures (depths anomalously shallow) from the predicted increase of crustal depth with crustal age predicted by the cooling plate model (e.g., Sclater et al., 1971) and the thickening boundary layer (cooling half-space) model (e.g., Parker and Oldenburg, 1973). These departures from the predicted subsidence (Figs. 1, 4-6; e.g., DeLaughter et al., 2005), typically begin at crustal ages of ca. 80 Ma, although the plate model fits the data from older seafloor better than the cooling half-space model. This is well illustrated by the Bermuda Rise (See Fig. 2 of Sclater and Wixon, 1986), which is about twice as “anomalous” relative to the latter model. In reconciling the two models, Parsons and McKenzie (1978) distinguished between an upper, cooler and therefore rigid “mechanical” lithosphere (boundary layer), and an underlying viscous region constituting a thermal boundary layer. In their model, both “layers” thicken as the square root of crustal age, but under old crust the thermal boundary layer becomes unstable and begins to convect, transporting heat to the base of the mechanical plate and thereby maintaining its thickness. The combined thickness of the two parts of a lithosphere thus defined is about 150-200 km. Numerical models of shallow convection under the plates (starting with McKenzie et al., 1974) also make predictions about the correlation between gravity and depth anomalies, first shown to be approximately satisfied for the BR by Sclater et al. (1975).

Taking a somewhat different tack, Crough (1978) led the efforts to explain this and other swells by thermal expansion within the plate and/or plate thinning, caused by plate motion over hotspots or thermal plumes below the plates (e.g., Crough, 1978; 1983b; Detrick and Crough, 1978). A simple plume model (e.g., Griffith and Campbell, 1990; Campbell, 2005) explains primarily only the height and horizontal scale of the BR. The relation between geoid anomaly and depth anomaly over e.g. Hawaii and Bermuda (Crough, 1978; Haxby and Turcotte, 1978) implied that these and other rises were supported by mass deficits at relatively shallow depths (40-70km), above the thermal boundary layer of Parsons and McKenzie (1978). Crough (1978) and Detrick and Crough (1978) also examined the rates of crustal subsidence for Pacific swells and concluded that hotspots or plumes had “reset” the thermal age to that of 25 Ma crust; i.e., after passing over the hotspot and being reheated at depth, the plate would subsequently share the subsidence of “normal” 25 Ma seafloor (As reviewed in a previous section, similar swell subsidence has not occurred at Bermuda).

If plate reheating occurred only by thermal conduction, an unrealistically high temperature anomaly would be required at depth (Detrick and Crough, 1978). This problem was first addressed by Parsons and Daly (1983), whose numerical models of cellular convection suggested that seafloor topography and gravity/geoid anomalies largely reflect the temperature structure in the upper part of the thermal boundary layer, which would participate in convective overturn but then lose heat by conduction into the overlying rigid mechanical plate. The rapid observed subsidence of oceanic crust in Pacific swells could then be explained if, as reasonable, the convecting upper thermal boundary layer does not (or only incompletely) share the motion of the rigid plate overhead. In other words, the observed subsidence would include the effect of thermal contraction of material not moving with the Pacific plate. We note that a Parsons-Daly type model might explain the difference between the fast subsidence of Pacific swells and the slow subsidence of the Bermuda Rise, if we imagine that fast-moving plates shear off the underlying thermal boundary layer convection, whilst slow-moving plates like the North American plate carry their underlying convection cells with them. Several authors have previously noted that any mantle convection associated with and in some way responsible for the Bermuda Rise must be traveling with the North America plate (Sclater and Wixon, 1986; Vogt, 1991; King and Anderson, 1998).

Although not doing any modeling of mantle convection, Sclater and Wixon (1986) concluded that the Bermuda Rise could have formed "either by a cooling boundary layer of which the thermal structure is reset by a localized convective jet" (i.e., the type of model proposed by Crough, 1978) or "by a two-layer thermomechanical plate perturbed by upper mantle convection". Sclater and Wixon preferred the plate model because it better explains the residual gravity and depth anomalies, better explains the lack of observed subsidence on older crust, and because it explains the lack of a heat flow anomaly. (At the time of their writing, no heat flow anomaly had been resolved on the Bermuda Rise).

As elaborated earlier, Detrick et al. (1986) collected additional heat flow data, which showed a 8-10 mW/m^2 heat flow high over the rise. They modeled these data and the previously known depth (800-1000 m) and geoid (6-8 m) anomalies and compensation depths ca. 40-70 km with two-dimensional models containing a low viscosity zone immediately below the mechanical plate. Cazenave et al. (1988) supported this model by calculating geoid to depth anomaly (admittance) as a function of crustal age. They suggested that swells like the BR, developed in older crust, are underlain by a thinner or more viscous low-viscosity zone. Following Parsons and Daly (1983), Detrick et al. (1986) concluded that the Bermuda Rise anomalies are "consistent with simple convection models in which the lower part of the thermally defined plate acts as the upper part of the thermal boundary layer of convection".

Louden et al. (1987 reduced the possible BR heat flow high to 5 mW/m^2. They attributed the lack of a larger anomaly, despite the large geoid and depth anomalies, to dynamic uplift of the BR by mantle convection, and the lack of time for the thermal equilibration. However, no convection models were presented.

Robinson and Parsons (1988 and other papers) concluded from numerical modeling of a low-viscosity zone (asthenosphere) below the conducting lithosphere and a high-viscosity underlying mantle that convective instabilities with Bermuda Rise-scale wavelengths and reasonable heat flow anomalies arise spontaneously below old oceanic lithosphere. The magnitude and onset time of convection was found very sensitive to the viscosity contrast (0.4-0.01) between the low-viscosity layer and the overlying lid. In this type of model, the Bermuda Rise is supported by thermally induced buoyancy forces.

In another numerical modeling attempt, Liu and Chase (1989) compared Hawaii with Bermuda, attributing each to Gaussian-shaped thermal perturbations caused by a steady-state plume, over which the plate moves at rates of 99 mm/a (Hawaii; "strong plume") and 15 mm/a (Bermuda; "weak plume"). Their models showed that thermal convection and convective thinning was required for Hawaii, while simple conduction worked for Bermuda. Model parameters assumed by Liu and Chase (1989) were: anomalous heat flux at base of lithosphere, 90-120 mW/m^2; excess plume temperature, 100-150º C; and radial velocity of plume relative to background convection, 2-4. The predicted evolution of heat flow, topographic and geoid anomalies over the Bermuda Rise (Fig. 10) incorrectly assumes the North American plate encountered an existing, steady-state "Bermuda plume" but probably gives a good sense of what simple conductive models can achieve. The models would have to be refined to reproduce the longer-than-assumed history of rise uplift (Jaroslaw and Tucholke, 1993) and probably smaller-than-assumed heat flow anomaly (Louden et al., 1987).

In the same year, Sheehan and McNutt (1989) calculated an effective elastic thickness (30 ± 5 km) of the mechanical lithosphere supporting the Bermuda volcanoes (from the topography and gravity anomaly over the pedestal), as well as the BR compensation depth (55 ± 10km, from the depth and geoid anomalies). Based in part on these results, they presented a new inversion technique (assuming instantaneous reheating from a point source) to put envelopes on the possible geotherms at the time of reheating (assumed to be 35 Ma) and at present. While their thermal expansion/conduction based models fit observations reasonably well without any dynamic uplift component, Sheehan and McNutt concluded that "a thermal convection mechanism is almost certainly required for thinning the lithosphere initially". As with other published models, the minimal or even non-existent post-volcanic subsidence of Bermuda was only marginally predicted—these difficulties are compounded by the discovery that 400-500 m uplift occurred subsequent to 33 Ma (Jaroslaw and Tucholke, 1993).

Assuming Morgan-type narrow plumes, Sleep (1990) analyzed the global hotspot dataset—mainly swell heights and dimensions, and geoid and heat flow anomalies—to constrain uplift mechanisms. Using swell cross sections and shapes, he estimated volume, heat and buoyancy fluxes. His measured values for Bermuda are: cross sectional area of swell, 1300 km^2 (1000 km^2 when corrected for asthenospheric compensation); buoyancy flux, assuming 15 mm/a plate velocity, 1.1 Mg/s (1.5 Mg/s according to Davies, 1988); volume flux (assuming a 225º C excess plume temperature), 48 m^3/s; and stagnation distance, 320 km. Sleep (his Fig.11) fit (by eye) a 'stagnation curve' of 500 km radius to the residual geoid anomaly over the eastern Bermuda Rise. This swell "snout" is interpreted as stagnation streamlines separating Bermuda plume outflow from the ambient asthenosphere below the western Atlantic. In recent years, plume temperature excesses such as 225º C have been challenged, and are not supported by heat flow data. Furthermore the Sleep's model for the Bermuda swell as produced by narrow mantle plume under the crest of the BR is inconsistent with the lack of evidence for migration during the last 30-45 my, unless the plume head is attached to and has moved with the North America plate.

Phipps Morgan et al. (1995) explored the possibility of a hot, partially molten 'swell root' supplying the volcanoes with magma, with the residue spreading out under the plate to elevate the rise. They focused on Hawaii, where the fast-moving plate drags this residue downstream—thus making room for new plume materials to rise and experience decompression melting. Although not mentioning Bermuda, Phipps Morgan et al. (1995) did examine model predictions for a plume underlying a stationary plate, using the Cape Verde rise and volcanoes as the type example. Given the slow (15 mm/a) motion of the North America plate at Bermuda, the latter would be closer to the Cape Verde in terms of model predictions. The model of Phipps Morgan et al. (1995) would attribute the relatively meager (compared to Hawaii) Bermuda edifice volume (760 km^3, see above), relative to the large swell, to the accumulation of residue below the rise, and to the thick, old lithosphere. This accumulation, although spreading laterally to help support the rise, would tend to impede decompression melting by clogging the underside of the plate in the area above the plume (e.g., Fig. 20 of Holm et al. 2006).

The lack of a Bermuda hotspot trace, the elongated shape of the rise (discussed further below), and the lack of subsidence at Bermuda led to another class of explanations—which involved convection attached to the North America plate and moving with it (e.g., Sclater and Wixon, 1986; Vogt, 1991). In particular, the deep boundary between the "new" Atlantic oceanic lithosphere and the old North American continental lithosphere would form a major thermal boundary. King and Anderson (1998) analyzed this possibility with numerical convection models, showing that many oceanic basement swells located ca. 1000 km seaward of continental margins (such as the Bermuda and Cape Verde rises) could be caused by such "edge-driven convection". Such models better account for the NE orientation of the Bermuda Rise, the development of the swell after the ocean had widened, and the lack of a hotspot trace. However, the apparent connection between Bermuda and the igneous activity in the Mississippi Embayment (Fig.8; Cox and Van Arsdale, 2002) would then be fortuitous.

NW Elongation of Bermuda Rise and other geographic coincidences: Accidents of mantle blob shape or also structural control and "Pre-Conditioning"?

If the NE-SW elongation of both the Bermuda volcanoes and the BR cannot be explained as a hotspot trace, how can they be explained? Clearly, the NE orientation of the four volcanic edifices and even the Bermuda edifice itself suggests structural control—for example magma ascent along normal faults formed at the ancient axis of the MAR. The location of this volcanism at the magnetic lineation bend (from NNE south of Bermuda to NE to the north; Fig. 1) also hints at structural control. Is there something special about the crustal or upper mantle structure or composition below Bermuda, or does this igneous activity simply mark the culmination of the underlying mantle blob (source) or thermal anomaly, as might be expected by the near-correspondence of the volcanoes with the BR swell summit (Fig.5)? Interestingly, the next largest volcano on the Bermuda Rise is the much older Muir Seamount (discussed above), Cretaceous in age but rising from oceanic crust of the BR crest about same age as that under Bermuda (Fig.3). Is this chance, or does it imply some atypical 'pre-conditioning' of the oceanic crust and upper mantle formed around M-0 time?

Farther northeast, the same-aged crust that hosts Bermuda and Muir seamounts forms an anomalous, west-dipping escarpment under the Sohm Abyssal Plain (Fig.4). This escarpment, somewhat diachronous (southwestward) towards younger crust, rises above the abyssal plain in the northeast, forming the J-anomaly Ridge (Fig. 4) a drillhole target at DSDP Site 384 (Tucholke and Vogt, 1979). The northeastern, highest part of this basement feature is developed in crust of M-2 to M-4 age (ca. 125-129 Ma), and is associated with anomalously high magnetization—the J-anomaly, a unique feature in the North Atlantic and perhaps globally. However, the timing corresponds to the 122-124 Ma magmatism "concurrent over a large area of the [North American] continent" (McHone, 1996, p. 328). Moreover, within dating errors, the Great Stone Dome lamprophyric pluton (Fig. 8) 1100 km NW of Bermuda, was intruded at the same time (Crutcher, 1983) and on the same NNE-to NE structural bend, replicated by sea-floor spreading from 175 Ma breakup to the 123 Ma crust on which Bermuda is located.

At least 600 km long, the J-anomaly ridge escarpment can be followed in seismic reflection profiles at least to 38.5º N, 54.5º W, ca. 1200 km northeast of Bermuda, and involving oceanic crust the same age as that under Bermuda. Is this pure coincidence, or is the crust and upper mantle along this isochron anomalous (e.g., a zone of weakness), favoring igneous activity both at Muir and later at Bermuda? Indeed, a structural or compositional preconditioning would also be consistent with an early (but not demonstrated, given the isotope dating problems for altered basalt) formation of a "Proto"-Bermuda Seamount, erupting the tholeiite pillows (as originally proposed by Aumento et al., 1974) long before the Eocene uplift of the BR and the intrusion of the lamprophyre sheets. However, the Cretaceous-aged New England Seamount chain does not appear anomalous where the chain crosses crust of slightly post-M0 age (Figs. 1,4).

Also problematic, but similarly hinting at "pre-conditioning", is the BR elongation more or less along crustal isochrons (Figs. 1,4). This correlation may of course be simple chance—in which case the shape of the BR more or less represents the shape of the underlying mantle blob (source) or thermal anomaly. To our knowledge, no other swells developed in pre-existing old oceanic crust have been observed to "follow" crustal structures (As discussed earlier, the BR developed in crust ca. 78-88 my old at the time of rise formation). The plate boundary "imprinting" of structure would be expected to influence only the crust and uppermost mantle, with deeper mantle lithosphere, formed gradually off-axis, not so influenced. However, to the extent mantle lithosphere forms rapidly off-axis, near-axis structural control could be imprinted at greater depth. To a first approximation, the BR happens to involve crust formed at very slow paleo-spreading rates, and with rough basement topography (Fig.4). One possible interpretation (besides mere chance) is that “very slow” crust and mantle lithosphere are somehow more “vulnerable” to swell development and magma intrusion. While little is known about the differences between mantle lithosphere formed at fast vs. slow opening rates, Lizarralde et al. (2004) determined the seismic structure along a line across the southern Bermuda Rise. They found the rough (450-1200 m relief), very slow (8 mm/a half-rate) crust that underlies most of the BR (Fig. 4) along their profile to be ca. 1.5 km thinner than the crust below the smoother (150-450 m relief), not so slow (13-14 mm/a half-rate) oceanic crust in the western part of the profile. Perhaps significantly, they found the mantle lithosphere below the two regimes to differ as well, at least to the 30-40 km depth limit allowed by their experiment (To prove beyond a doubt that this anomalous upper mantle structure was created at the ancient spreading axis, not during BR formation, the seismic experiment would need to be repeated along a conjugate profile in the eastern Atlantic). This mantle difference they attribute to less complete melt extraction under lower spreading rates, at the ancient MAR axis. In any event, the results of Lizarralde et al. (2004) suggest that the mantle lithosphere is distinctive (retaining more gabbro) at least to 30-40 km depth below the “very rough” belt that more or less corresponds to the Bermuda Rise. How strength varies with depth varies with lithology, and we suggest (following Vink et al., 1984), that mantle lithosphere which retains more gabbroic components is somewhat weaker than a more purely peridotitic mantle, and might be more vulnerable to fracturing. With more retained gabbro melts, such mantle would also be more vulnerable to renewed partial melting associated with swell development. At least, reduced strength in the uppermost mantle might help account for the greater seismicity of the Bermuda Rise (Fig. 4; Zoback et al., 1986).

Although the Cape Verde Rise is not elongated in a “structural” direction, it was developed within crust of largely similar age (ca.155-105 Ma; Williams et al., 1990)) as the Bermuda Rise, and should, unless modified by rise development processes, have the same crustal and upper mantle structure as determined by Lizarralde et al. (2004) for the Bermuda Rise. Barring coincidence, this adds credence to a compositional and/or structural “pre-conditioning” role for the lithosphere, and thereby undermines a simple mantle plume model for these features, although Montelli et al. (2004) adduce tomographic evidence for a Cape Verde plume rising from at least 1900 km depth.

The “Bermuda Event”- Symptom of a global plate tectonic reorganization?

Was the middle to late Eocene/early Oligocene Bermuda Rise formation and volcanic activity just an anomalous event in the upper mantle below the western North Atlantic? We suspect not, suggesting instead that the activity was part of a global plate reorganization episode. This is still speculative, and difficult to prove on a planet with ongoing “noise” of igneous episodes and plate reorganizations.

Some degree of global synchronism would be expected if there were episodic reorganizations of mantle dynamics, whether by plumes (e.g., Vogt, 1972; 1975) or simply the effects of episodic and unavoidable plate kinematic/tectonic reorganizations (e.g., Anderson, 2002). For example, Rona and Richardson (1978; see also Patriat and Achache, 1984) listed many Eocene events they attributed to the collision of India with Eurasia, i.e. the closing of the Tethys Ocean.

Plate kinematic reorganizations would generally change intra-plate stress fields as well as asthenosphere flow patterns, and perhaps account for the Bermuda event as originating in the lithosphere and asthenosphere. Although the BR is characterized by somewhat increased seismicity (as discussed above) there have been no specific model predictions of how the Eocene plate reorganization could have increased stress or decreased plate strength in areas like the BR or Virginia. Moreover, as also discussed above, seismic tomography (Zhao, 2004; Ritsema, 2005) suggests that anomalously "slow" mantle extends to at least 400 km, possibly to 1000 km below the BR area, although an anomaly specifically associated with the mantle below the rise has not been resolved. Alternatively, the postulated relatively abrupt, widespread Eocene onset of gravitational instability, as previously subducted lithospheric slabs temporarily stuck (flattened out) near the 660 km discontinuity begin sinking into the lower mantle (Lithgow-Bertelloni and Richards, 1998; Fukao et al., 2001), would require upwelling in other areas for mass balance purposes. While this plausibly explains the timing of the Bermuda event, neither the specific location of the BR, nor its apparent elongation along crustal isochrons, are thereby specifically explained. Whatever the precise mechanism, a global Eocene reorganization of plate motions would be expected to affect plate boundaries and even plate interiors. Compilations of accurately dated tectonic or igneous events are only the first step: Temporal correlations do not necessarily translate to causality.

We first ask if BR formation relates to motion of the plate in which the BR and its volcanoes were developed. As exemplified in Fig. 6, the direction of spreading between the Africa and North America plate, as recorded by the Kane and other transform fracture traces in the central North Atlantic (See Vogt and Tucholke, 1989, for a more detailed map along the Kane FZ corridor), changed by ca. 30º to a more E-W direction about the time of magnetic anomaly 21, now dated at 46.5 Ma—essentially coeval with the Hawaii-Emperor Bend (47 Ma; Tarduno, 2005; see further below). Within error limits, this is also the time of Bermuda Rise and seamount formation. The coincidence in timing may be explained by a changing stress field acting on the plate—with the changed stress fields responsible for opening the lithosphere to intrusions and upwelling (BR formation) and simultaneously forcing a change in plate motion. However, why then did other bends not also cause mid-plate swells and volcanism?

Ages of more distant, -possibly coeval "events" (Fig. 10) can be obtained from stratigraphic evidence, from direct radiometric dates, and from magnetic anomalies in the oceans. Comparison among different types of dates depends on a time scale relating stratigraphic time and magnetic polarity reversals to absolute (radiometric) time. In this paper, we mostly use the time scale of Gradstein et al. (2004), but Fig. 10 is based on the slightly different Palmer and Geissman (1999) scale.Earlier we reviewed the evidence dating the beginning of the Bermuda event (initiation of BR uplift) to some time or time interval within the Middle Eocene, which lasted from 48.6 to 37.2 Ma. The emergence of Bermuda volcano above sea level (based on DSDP Site 386) happened somewhat later-- in the early to middle Late Eocene (i.e., ca. 37-35 Ma). However, since the volcanism must have begun before the first erosion products arrived at Site 386, BR uplift and volcano birth may have begun at the same time in the Middle Eocene. The sheet intrusion event (33-34 Ma) was plausibly, like the Hawaiian post-erosional volcanism, a local phenomenon deep below Bermuda; however, it might also have been triggered by a later global stress reorganization, distinct from the Middle Eocene event.

A limited compilation of more or less coeval events around the world suggests that, indeed, Bermuda was part of a larger "happening" (Fig.10). Atlantic events include 1) the extinction of the Mid-Labrador Sea spreading axis (also called Ran Ridge) at least by C13 time (ca. 33-34 Ma) according to Kristoffersen and Talwani (1977) but possibly as early as C18n (ca. 38-39.5 Ma)—the very slow spreading prevents reliable identification of the corresponding magnetic lineations. In the Norway Basin, a regionally unique linear ridge developed along the Aegir spreading axis at C18n time (Vogt and Jung, submitted), plausibly in response to the "annexation" of the Greenland plate by the North American plate.

On the neighboring continent of North America, felsitic and mafic magmas as well as diatremes—the only Cenozoic igneous activity in the eastern US—were emplaced in one small part of the Virginia and West Virginia Appalachians (Figs.8, 10). De Boer et al. (1989) call this the "Shenandoah Igneous Province", pointing out that the same area had been intruded by diverse rock types in the Jurassic and early Cretaceous, with upwelling mantle material structurally localized at the intersection of a modern geothermal high and the major 38th parallel fracture zone (de Boer et al., 1989). Radiometric dating and paleomagnetic work indicates the intrusions began ca. 49 Ma—with most of the activity occurring during a few million years, latest dates ca. 35 Ma (e.g., Fullagar and Bottino, 1969; Southworth et al., 1993; Tso and Surber, 2002). Nearly all the intrusives are reversely magnetized (Lovlie and Opdyke, 1974), suggesting most may have been emplaced within less than 1 m.y. (de Boer et al., 1989). Within the various dating uncertainties, this unique igneous episode (middle Eocene; 45 ± 5 my; de Boer et al., 1989) was Bermuda's precise contemporary—albeit without the formation of a corresponding rise/swell.

Extensive deep coring and 'backstripping' calculations in New Jersey and the adjacent continental shelf and margin 600-700 km Northwest of Bermuda has revealed a local –probably largely eustatic—sea level history from the early Cretaceous to present (Van Sickel et al., 2004). The results show that the greatest (100-150 m) pre-Quaternary sea level fluctuations in the last 100 my (from a high ca. 54 Ma to a low at or below modern sea level ca. 46 Ma, and new high ca 43 Ma) occurred during the period of interest. Van Sickel et al. (2004) speculate that the Early Eocene high was related to opening of the Greenland-Norwegian Sea, and the succeeding low to the closing of Tethys, which increased ocean basin space by doubling continental crustal thickness under Tibet.

The only spreading-type plate boundary between the North American and Caribbean plates—and the closest of any to Bermuda--- is found in the Cayman Trough. East-west rifting, which continues today, began in the late Eocene (Rosencrantz et al., 1988).

Across the MAR from Bermuda, Africa also experienced enhanced mid-plate igneous activity at many sites about 40 Ma (Bailey and Woolley, 2005). Repeated synchronous activity also occurred at many of the same sites ca. 20, 85 and 120 Ma, strongly supporting the concept of global episodes of increased mid-plate stress effecting tectonic-magmatic reactivation.

Although the more or less persistent volcano-tectonic activity in the North American Cordillera complicates resolution of unique igneous or tectonic events, Oldow et al. (1989; pp. 219-220) state that "A dramatic transition in magmatic type occurred throughout the Cordillera about 40 to 42 Ma. This event coincided with a major (global?) change in plate interaction. Timing of this transition corresponds to the age of the bend in the Hawaii-Emperor seamount chain (42 Ma; Dalrymple and Clague, 1976) indicating a change in Pacific plate motion". Bally et al. (1989) identify five Mesozoic-Cenozoic Tectonostratigraphic events in the geology of North America, and relate them to global plate-tectonic reorganizations. The last such event was the "plate reorganization that followed the collision of India with Eurasia in the early Eocene". This event is marked in North America by "the end of compression of the Cordilleran foreland belt, the inception of extensional tectonics in the Basin and Range Province, and of strike-slip tectonics in California, as well as the establishment of major drainage patterns that are responsible for the accumulation of thick Tertiary clastic sequences of the Gulf Coast". (Bally et al., 1989, p.7).

Winterer et al. (1989) cite the Hawaii-Emperor bend age at 43.1 ± 1.4 Ma. Although estimates of bend age changed little since the early 1970s, Sharp and Clague (2002), using just dates along the Emperor chain, recently suggested the bend may actually be as old as 50Ma, while recent drilling (ODP Leg 197) indicates 47 Ma as the most likely Bend age (Tarduno, 2005). However, given the various uncertainties, the bend age may still correlate with beginning of uplift of the BR some time during the Middle Eocene (49-37 Ma).

An abrupt acceleration of motion of Australia away from Antarctica along the young Southeast Indian Ridge can be confidently dated to the beginning of C18n (ca. 38-39.5 Ma) (Cande and Stock, 2004). This was preceded according to the same authors by the start of the main episode of Cenozoic East-West Antarctic separation and the initiation of seafloor spreading in the Adare Basin, and other plate rearrangements at about Chron 20 time (42-45 Ma).

Cenozoic uplift and igneous activity in the West Antarctic Rift area (Victoria Land and Ross Sea; Rocchi et al., 2005)) bears striking resemblance to what happened far away in Bermuda: A main denudation- uplift in the Middle Eocene was accompanied by igneous activity dating from 48 to 35 Ma, with a geographic shift in activity ca. 33 Ma. However, unlike at Bermuda, renewed igneous activity began in the Middle Miocene and has continued locally to the present. In contrast, spreading in the Dare Trough occurred from 43 Ma to 26 Ma, closely coinciding with the entire Bermudan episode—rise uplift and igneous activity.

Courtillot et al. (2003; their Fig. 2) compiled "time variations of four significant kinematic, geographic or dynamic indicators of hotspot motion", relating to Hawaii and Reunion hotspot traces and true polar wander. All four parameters show a step-like velocity change ca. 45 Ma (at least some time between 50 and 40 Ma). Evidently a major reorganization of mantle motion occurred at 45 Ma—with Reunion and Hawaii hotspots moving rapidly towards the equator prior to that time, and slowly if at all subsequently. The relatively sharp Hawaii-Emperor bend implies that this reorganization of mantle motion was a rather abrupt event.

Finally, Vogt (1975) speculated that if plumes originate near the core mantle boundary (CMB), changes in plume convection might influence geomagnetic reversal frequency, and any time lag between frequency change and hotspot expressions at the earth's surface might yield the rise speed of plume material. He noted an apparent correlation between the time of the Hawaii-Emperor Bend (then dated at 43 Ma) and an increase in average reversal frequency about that time. However, improvements in radiometric dating (Fig. 10) seem to rule out such a correlation: The Hawaii-Emperor Bend is more likely about 47 Ma in age (Tarduno, 2005), whereas reversal frequency appears to have increased ca. 30-31 Ma, with a prior short-lived interval if higher reversal frequency between about 39 and -35 Ma, i.e. of "Bermudan age", probably not statistically demonstratable. Moreover, plume modeling suggests a much longer rise time for new plumes (100 m.y., Campbell, 2005). However, modeling of the geomagnetic dynamo (Glatzmaier et al., 1999) shows that different patterns of geographic variation in heat flux from the core across the CMB can affect reversal frequency as well as dipole moment, secular variation, and the duration over which reversals occur. Whereas the correlation of the ca. 40 Ma igneous event to geomagnetic reversal frequency change is unconvincing, it is more compelling for the 80 and 125 Ma limits of the Cretaceous normal superchron (Vogt, 1975; Larson and Olson, 1991; Bailey and Woolley, 2005).

Discussion and Conclusion

What produced Bermuda and the Bermuda Rise—Why then and why there?

As shown by the length of this paper and the large number of references, Bermuda and its rise (swell) have been extensively studied, and practically every paper dealing with oceanic swells/rises and hotspots at least mentions Bermuda, and in many cases uses it—in contrast to Hawaii-- as the "type example" of a weak hotspot under a slow-moving plate (The fact that Bermuda has actually been quiescent for ca. 30 million years is generally ignored; better "type examples" would be the Cape Verde and Canary Islands, volcanically active and rising from the nearly stationary Africa plate). On balance, almost all models involve some component of heating and thinning of the lower lithosphere, but, as typified by Sheehan and McNutt (1989), some contribution from dynamic uplift (ascending convection) cannot be excluded. The strongest proponents of a dynamic uplift model (Louden et al., 1987) base their case largely on the absence of a significant heat flow anomaly. Most thermal uplift models also have difficulties explaining the lack of subsidence since igneous activity ended, probably prior to 30 Ma—and most such papers predated Jaroslaw and Tucholke (1993), who showed evidence for 400-500 m continued uplift after that time, exacerbating the problem. However, Bermuda is by no means unique in terms of lack of subsidence: In the eastern Atlantic volcanic archipelagos and their swells, not only has there been no subsidence, but old oceanic crust was elevated above sea level in the Cape Verde Islands, and submarine lavas associated with the island volcanism are now up to 400-500 m above sea level (e.g., Holm et al., 2006). The continued accumulation of melt residues below these swells (as in the Phipps Morgan et al., 1995, model) best accounts for this pervasive uplift or lack of subsidence where plates are nearly stationary. However, Bermuda and its swell must have been carried some 650 km west, away from any fixed mantle source, since 35 Ma, according to all the major absolute-motion models (e.g., Duncan, 1984). A Phipps Morgan et al. (1995) type model might attribute this to a melt-residue cushion that travels with the plate. To adapt a simple plume model to Bermuda would require a plume source that either travels with the North America plate or was bent over by the lithosphere or by larger scale mantle motions (e.g., Steinberger, 2000), perhaps finally breaking its connection in the early Miocene, when uplift finally ended (according to Jaroslaw and Tucholke, 1993).

An alternative, non-plume type explanation for Bermuda is edge-driven convection (King and Anderson, 1998), which accounts for several features not predicted by the various other models: 1) formation of Bermuda when the Atlantic basin was already wide; 2) elongation of the BR parallel and ca. 1000 km SE of the oceanic-continental crustal boundary; and 3) the lack of migration of the BR. Radially symmetric upwelling, no matter how wide the plume head (e.g., Griffiths and Campbell, 1991), would not be expected to produce swells elongated perpendicular to plate motion, as is the case for Bermuda—although we noted that the crust and lithosphere under most of the rise may be structurally or compositionally preconditioned (Fig.4), having been formed at very slow spreading rates at the ancient ridge axis (Lizarralde et al., 2004). While edge-driven convection models do not explain the apparent ancient continuation (Fig. 8) of a Bermuda hotspot trace via Mississippi (65 Ma) to Kansas (115 Ma), these suggested traces seem highly suspect, given the occurrence of igneous activity in Virginia and in Texas, and the implied "blob" like behavior of a Bermuda plume.

Courtillot et al (2003) place Bermuda in the category of "A-type plumes", ie those not due to deep mantle plumes. This accounts for the poor "rating" Bermuda receives (Anderson, 2005; and Courtillot et al., 2003 both give Bermuda a zero) when compared with predictions of deep mantle plume models (some predictions, for example the He^3/He^4 ratios for its igneous rocks, have not yet been tested on Bermuda, however). While we concur with this assignment, we find it curious that the otherwise rather similar Cape Verde Rise gets a rather higher rating (1 by Anderson, 2005 and 2 by Courtillot et al., 2003) than Bermuda: It would seem that the main difference between the two is age: the Cape Verde archipelago and rise began developing in the early Miocene and are currently active, while Bermuda's activity ended at that time, and should not be tested against currently active hotspots. Another curious discrepancy involves calculated buoyancy flux (Sleep, 1990), which is only slightly higher for Iceland (1.2 Mg/s) than for Bermuda (1.1 Mg/s), when Iceland gets the highest plume score on the planet: 4+/5 by Courtillot et al. (2003) and 8?/12 by Anderson (2005).

Why did the Bermuda swell and volcanism happen when it did? In this paper (Fig.10), we attribute the timing to a ca. 45 Ma, mantle-wide rearrangement of convection (e.g., Courtillot et al., 2003), perhaps associated with the closing of the Tethys (Rona and Richardson, 1978; Patriat and Achache, 1984) or/and the sinking into the lower mantle of slabs that had accumulated near the 660 km transition zone (Fukao et al., 2001). The location of the BR may simply reflect a geochemically more fertile region of the upper mantle. With the asthenosphere being close to its pressure melting point, even a modest rise, perhaps triggered by edge-type convection, would trigger substantial partial melting, causing partial melting, plate thinning, and uplift. Alternatively, the region of partial melting may have been influenced by the pre-existing composition and structure of the oceanic mantle lithosphere (Fig.4).

The debate about the origin of so-called hotspots continues—and professional opinions have moved in recent years towards more "plate/shallow origin" and less "plume/deep origin" causality (e.g., Anderson, 2005). Nevertheless, some tomographic studies have adduced seismic speed anomaly anomalies as evidence for at least some deep mantle plumes (e.g., Montelli et al., 2004), results which have however been challenged (e.g., van der Hilst and de Hoop, 2005), Nevertheless, seismic tomography shows that the great majority of hotspots are located above two large regions (equatorial Pacific and Africa) of slow S-wave (Ritsema, 2005) and slow P-wave (Dziewonski, 2005) in the lowermost mantle. As the latter author notes, there must be 'some communication between the upper and lower mantle systems'. Most of North America and the western Atlantic is underlain by "fast" lowermost mantle, whereas the eastern Atlantic (Canaries, Madeira, Cape Verdes) is part of the African "slow" lower mantle province. While some have attributed the greater abundance of volcanism in Africa and the eastern Atlantic to the more nearly stationary plate, this explanation does not account for a complete lack of mid-plate volcanism in very slow northern parts of the North America plate, not to speak of the lack of it on many fast-moving plates. Perhaps the movement of the North America plate 650 km further into the region underlain by fast lower mantle in some way accounted for shutting off the anomalous mantle source under Bermuda.

The two stage igneous history implied by available data

If the 33-34 Ma "Bermudite" (Aumento and Gunn, 2005) intrusive sheets—postdating the shield formation—can be shown to be, as we hypothesize, the Bermudan equivalent of Hawaiian post-erosional basalts, Bermuda may affect explanations for the latter. In the Hawaiian case, volcanism continues for up to at least ca. 5 my (Clague and Dalrymple, 1989; Gurriet, 1987), but with 1-2.5 my long quiescence between the early tholeiitic phase and the volumetrically trivial post-shield ("post-erosional") alkalic phase. The lag for Bermuda is clouded by the still-uncertain age of the shield basalt flows; a long quiescent period is possible but not proven: The continuous arrival of volcaniclastic turbidites at Site 386 implies continuous erosion, not necessarily volcanism. If we date the shield formation at 40-45 Ma based on Leg 386 volcaniclastic results (Tucholke at Vogt, 1979), the corresponding time from shield to sheets is ca. 6-12 my, within the "ca. 5 my" range of Gurriet (1987) and Phipps Morgan et al. (1995). Comparable or even greater (up to 65 my) volcanic durations characterize eastern Atlantic volcanoes, e.g. in the Canary and Cape Verde archipelagos.

Clague and Dalrymple (1986) explained the late eruption of Hawaiian post-erosional volcanics, and the 1-2.5 my quiescent period in terms of the time it takes for mature Hawaiians shields to be carried by plate motion over the arches associated with the youngest, active shields. As the old shields passed over the swell, they would experience flexure and thus faulting, thus causing decompression melting at depth, and providing pathways (along faults) for magma ascent. Gurriet (1987) modeled the conductive heating and partial melting of the lowermost lithosphere as it passes over a putative hotspot, and attributed the ca. 5 my spread in eruption ages (Clague and Dalrymple, 1989) to the time required to extract small percentages of partial melt. Gurriet's models could not account for a quiescent period, so he appealed to the swell-overriding idea of Clague and Dalrymple. Phipps Morgan et al. (1995) attributed the ca. 5 my duration to "post-emplacement spreading and thinning of the swell root", and on the basis of this time calculated a viscosity, for their hypothesized depleted swell root, of ca. 1-3 x 10^{20} Pa s. In their view, melting is stopped below shields because vertical motion is impeded, while progressive, laterally-upward migration of the spreading root materials causes decompression melting and volcanism at some distance along the existing chain, thus explaining the observed duration of activity along some chains Ribe and Christensen (1999) were able predict a ca. 4 m.y. quiescent period, between the main shield stage and the arrival of the edifices over a secondary melt zone.

However, we are suspicious of all the above models, because all assume movement across a melting anomaly, and therefore do not explain an apparently similar (albeit longer) duration (and possible quiescent period) of volcanism on Bermuda, which is not a chain, and which sits on a swell that has apparently traveled with the plate.

Why enhanced mid-plate seismicity on the Bermuda Rise?

As first noticed by Zoback et al. (1986) and substantiated by earthquakes that occurred after they went to press (Fig.4), the BR is a region of abnormally high (for a mid-plate oceanic region) seismicity. This is the only known way in which Bermuda and the BRe are in any geophysically/geophysically active today. The reason for this enhanced seismicity is elusive. In this paper we have used the deep seismic reflection results of Lizarralde et al. (2004) to suggest that the upper mantle below the Bermuda Rise is relatively weaker, due (to extrapolate from Vink et al., 1984) to greater gabbro retention under slow spreading rates (Klitgord and Schouten, 1986) at the ancient Mid-Atlantic Ridge. Under this scenario, there is no need to postulate any swell-related stress enhancement under the BR.

Alternatively, the enhanced seismicity reflects unusual stress conditions associated with swells. Neglecting the various other stress sources likely to be affecting the western North Atlantic lithosphere (e.g., Zoback et al., 1986), Zhu and Wiens (1991) calculated thermoelastic stresses caused by putative hotspot heating. They find some some agreement between their predicted deviatoric stress and the few fault-plane solutions and seismicity, both for the Bermuda and the Cape Verde rises. However, it seems unlikely that theBR, whose volcanism and most if not all uplift had ended prior to 30 Ma, would still exhibit the same thermoelastic stresses and mid-plate seismicity as the higher, younger (and volcanically active) Cape Verde Rise. Perhaps the Cape Verde Rise, also developed within “slow-spreading” crust about the same age as that of the Bermuda Rise, is therefore also weaker and more easily ruptured. However, this cannot be the only factor at work, otherwise much of the crust in the North Atlantic, formed at slow spreading rates, would exhibit higher mid-plate seismicity.

Future Research Approaches

How can future research discriminate among the wealth of models so far presented in the literature, and perhaps others still to be developed? First, earthquake-source seismic tomography with even better resolution than e.g. in Fig. 19 of Zhao (2004) may delineate the “slow” region in the upper 400 km of the mantle below the Bermuda Rise. Below some such depth, the mantle must be decoupled from the motion of the North America plate. Explosion-source seismic experiments with OBS arrays on the rise, expensive and perhaps not feasible due to concerns for marine mammals, would probably be needed to test for a buoyant refractory root (e.g., Phipps Morgan et al., 1995; Holm, 2006) which might extend from ca. 50 km (the depth of swell compensation deduced from geoid data) to 200 km below the rise (The depths of origin of the Bermuda sheets is ≥150 km; Olsen, 2005). Airgun-source multichannel profiling of the type conducted by Lizarralde et al. (2004) across the southern Bermuda Rise might detect cooled intrusions which reached the upper mantle lithosphere, thus testing the model of Withjack (1979), but any reduced wave-speed anomalies would have to allow for melt retention under the slow spreading rates forming most of the crust under the Bermuda Rise (Lizarralde et al., 2004). Spreading-rate dependent mantle velocities should change only gradually northwards along isochrons, whereas any anomalies associated with the Bermuda Rise should reach extrema under the rise summit (e.g., the X’s in Fig. 5).

Additional drilling into the igneous basement of Bermuda and its three smaller satellites (Plantagenet/ Argos and Challenger banks, and Bowditch seamount) is essential. Given the geologic complexity we know from volcanic islands (e.g. the Cape Verde archipelago; Holm et al., 2006), it seems highly unlikely that even the Bermuda edifice was formed by a simple two-stage process of an Eocene tholeiitic shield, followed after ca. 6-12 my quiescence by 33-34 Ma

"Bermudite" (Aumento and Gunn, 2005) sheet intrusions. Hints of other igneous units in the edifice are provided by Foreman (1951), who wondered whether the rather abundant zircon and quartz he found in soils derived from the volcanics could have come from "late intrusions of a more acidic nature". Hints of explosive andesitic or basaltic eruptions are provided by altered tuff pebbles (Foreman, 1951) and small amounts of altered ash at DSDP Site 386.

Further, we have assumed—with no direct evidence!-- that the three satellite edifices are of the same age as Bermuda. They may have formed like major Pacific seamount rift zones (e.g. Vogt and Smoot, 1984) extending NE and SW from the main Bermuda edifice, or they may have erupted from separate conduits extending deep into the upper mantle. All this remains pure speculation until they are drilled, preferably to several km depth, and the core samples analyzed. A detailed magnetic and gravity survey of all four edifice summits and upper flanks should be conducted in advance of any drilling—to help map out the structural layout of intrusive sheets, lava accumulations, central conduits and flanking volcaniclastic debris aprons.

Several deep holes—similar to those drilled at DSDP Site 386 (Fig. 1) and 385 (New England Seamounts; Tucholke and Vogt, 1979) should also be placed around the bases of the four Bermuda edifices, as close as possible to the bases, but still practically penetrating the volcaniclastic debris and flows (i.e. the "inner" seismic reflector A^v of Tucholke and Mountain, 1979), to recover and biostratigraphically date the youngest sediments overlain by the oldest Bermudan rocks. Recovery of larger, less altered rock fragments would also be more likely closer to the base of the volcanic edifices.

A transect of boreholes across and along the, just deep enough to sample the oldest hemipelagic sediment just above the Eocene biosiliceous turbidites, should be able to resolve the detailed spatial-temporal pattern of BR uplift initiation. Such boreholes might also recover the time when bottom currents were first steered by the BR (e.g., Ayer and Laine, 1982). Several boreholes should be placed along the Kane fracture zone to calibrate the uplift history deduced by Jaroslaw and Tucholke (1994) from seismic reflection mapping of more local turbidites. Abyssal plains, with gradients of 1:1000 or less (the present BR is surrounded on three sides by modern or at least late Pleistocene abyssal plains; Pilkey and Cleary (1986)), should be extremely sensitive to small elevation changes. As depicted in Figure11, the Mid-Eocene turbidite offlap pattern—in time and space—would depend on the uplift mechanism. A plume model predicts rise uplift migrating radially outwards from a region above the upwelling plume head, whereas a "distributed source" model with simultaneous partial melting or/and temperature rise would predict simultaneous uplift over the entire area of the present Bermuda Rise. A plume-type model (Griffiths and Campbell, 1991; Campbell, 2005) predicts a possible lag of about 2 million years between first uplift above the center of the plume head, and uplift on the outer fringes of the expanding asthenosphere (I. Campbell, personal communication, 2005), and such a lag should be recorded by the offlapping turbidites, and is probably measurable from biostratigraphic dating of the first hemipelagics deposited on the last turbidites. The Campbell (2005) model also predicts a lag of ca 2-4 million years between uplift initiation and onset of volcanism. Current dating (Middle Eocene for onset of uplift; late Middle to early Late Eocene or earlier for the volcanism) make such a lag possible, but not proven. The plume model also predicts a possibly testable early central uplift several hundred meters higher in the beginning, before spreading below the plate flattens the head and reduces swell height in the next ca. 2 million years. The "swell root spreading" models of Phipps Morgan et al. (1995; their Fig.6) make even more specific and testable predictions about the rise uplift and swell radius as a function of time. Uplift resolution might be further refined by correlation of individual turbidites from one borehole to the next. Some of the thicker and compositionally distinctive Quaternary turbidites have been correlated from one core to another in modern abyssal plains surrounding Bermuda (Pilkey and Cleary, 1986)Readers interested in joining the authors in proposing the Bermuda Rise borehole transects discussed above are welcome to join the authors and others in this effort.

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Figures

Fig.1. Bottom: Bathymetry (Smith and Sandwell, 1997) of Bermuda and vicinity, with heat flow stations (W/m^2) from Detrick et al. (1986) and Hyndman et al. (1974), location of DSDP site 386 (The Shipboard Party, 1979), and sea-floor spreading magnetic anomaly M-0 from Klitgord and Schouten (1986). Solid line shows outer limit of seismic reflector A^v—caused by volcanogenic sediments of Bermudan origin (Tucholke and Mountain, 1979). Dashed lines indicate major fracture zones as interpreted from geophysical data.

Top: Greater Bermuda Rise region: Bathymetry (Smith and Sandwell, 1997), key magnetic lineations (Mueller et al., 1993), DSDP Leg 43 drill site locations (stars); heat flow values (Detrick et al., 1986; Louden et al., 1987); residual depth anomaly (white contours at 200 m interval; Sclater and Wixon, 1986); residual geoid anomaly (meters; dashed black lines); trace of Kane Fracture Zone (dashed line, after Jaroslaw and Tucholke, 1994); and predicted track (red) of North America plate over a fixed Bermuda hotspot (Duncan et al.,1984), with predicted present hotspot location shown by large red circle. Heat flow station on southern BR shows Detrick et al. (1986) value and "(49.9*) same original data, but recalculated by Louden et al. (1987). NES, New England Seamounts; SAP, Sohm Abyssal Plain; HAP, Hatteras Abyssal Plain; NAP, Nares Abyssal Plain; MS, Muir Seamount.

Fig. 2. Top, satellite image of Bermuda platform, showing shallow (<25 m) water central "lagoon(blue) reef rims (light blue) and islands (light gray and green) (NASA World Wind v.1.3 at about 40 km altitude); isolated white specks distant from Bermuda are clouds. Bottom: Bathmetry of the Bermuda Pedestal, the associated three smaller edifices, and surrounding seafloor; 100 m contour interval, based on Smith and Sandwell (1997).

Fig. 3. Geologic map of Bermuda Rise area—part of the Geologic Map of North America (Reed et al., 2005). Water depth contours (blue) at 1000 m intervals; Cretaceous/Jurassic and Lower/Upper Cretaceous crustal age boundaries (thin black lines). Note most of surface geology is dominated by Quaternary (white) and late Tertiary (yellow) hemipelagic sediments. Igneous outcrops (small "v" pattern) are limited to a few fracture zone escarpments (original oceanic basement) and seamounts like Bermuda.

Fig. 4. Bathymetry, residual depth contours, and boreholes reproduced from Fig. 1-top; western (after Sundvik et al., 1984) and approximate eastern limits (thick solid lines) of rough oceanic basement topography, created by slow spreading; eastern limits of seismic reflector A^t (dotted lines; Tucholke and Mountain, 1979), turbidites that covered part of the region prior to uplift of Bermuda Rise and spread eastwards inside the Fracture Zone valley (Jaroslaw and Tucholke, 1994); JAR, J-Anomaly Ridge and associated basement ecarpment (Tucholke and Vogt, 1979; Tucholke and Ludwig, 1982); small open circles, earthquake epicenters (NEIC database as of 12/2005; with additional older events from Zoback et al., 1986); A, B and C denote epicenters whose first motion and/or depth could be determined. For events with aftershocks, only the location of the main shock is plotted.

Fig. 5. Residual depth anomalies (left, in meters) and geoid anomaly (right, in meters) over the central part of the Bermuda Rise, with our pick of anomaly maxima (slightly southeast of the more dramatic local highs associated with the Bermuda edifice cluster). Modified from Figs. 2 and 3 of Sheehan and McNutt (1989).

Fig. 6. Depth anomaly contours from the western North Atlantic/Bermuda Rise region across the Mid-Atlantic Ridge to the conjugate region in the eastern Atlantic. Some fracture zone traces are

shown (The Kane FZ is the southern of the two transforms shown) to aid linking conjugate crustal parcels. Note the Southeast Bermuda Deep (SEBD) discussed in the text (Fig. 1, bottom) corresponds do a conjugate deep region east of the MAR axis, and was therefore probably formed at the ancient MAR axis. The Cape Verde Rise (swell) probably formed during Tertiary times (volcanism since at least 19 Ma; reviewed in Holm et al., 2006), elevating part of the eastern SEBD conjugate. Based on global depth anomaly grid derived by DeLaughter et al. (2005) from their model GDH2. Isochrons are from Mueller et al. (1997), with the locations (dotted) of anomalies 21 (46.5 Ma) and 13 (33.5 Ma) added to show the transform trends (Africa-North America plate motion directions) at about the time of the formation of the BR and edifice, and the time the sheets were intruded into the lava flows. Note the ca.30º bend recorded by the western flank of the Kane FZ and the adjacent Northern FZ at about anomaly 21 time.

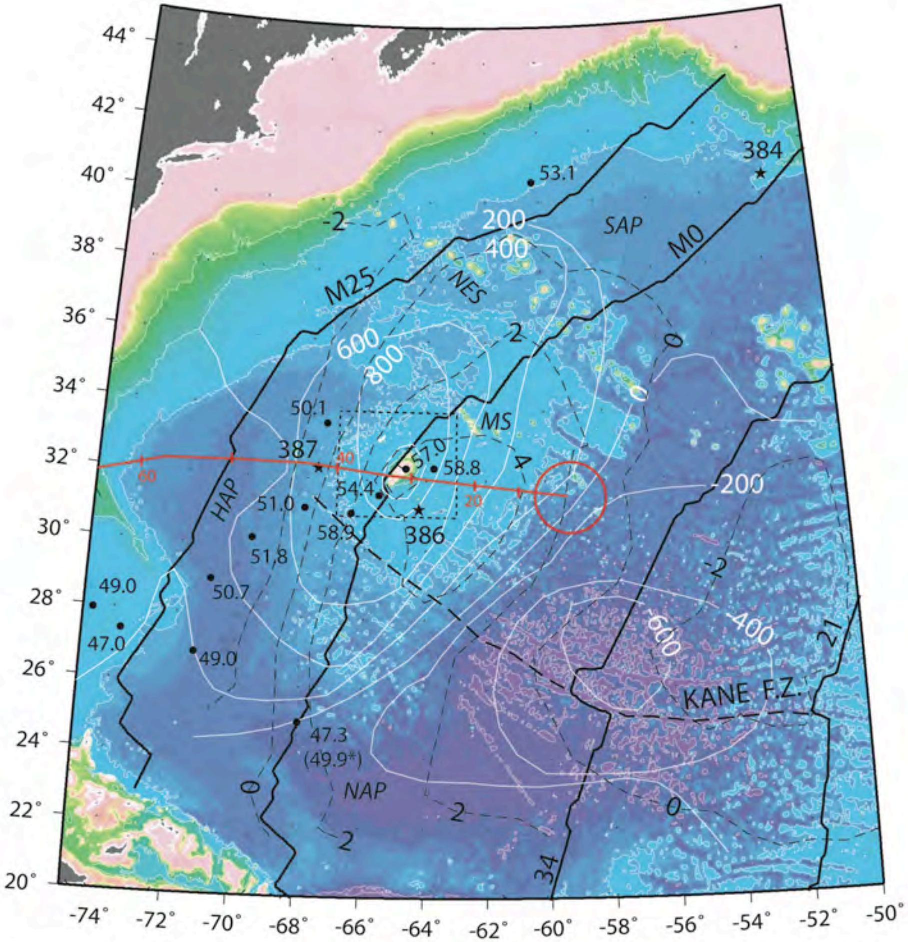
Fig. 7. Bathymetry (200 m contour interval, based on Smith and Sandwell, 1997) of eastern Bermuda Rise region, showing the predicted trace (red; Duncan et al., 1984 and other models) of igneous activity expected from a continuously active, fixed Bermuda hotspot model. Models predict a present hotspot location about 32º N, 58º W (red circle) a region lacking seamounts large enough to be detected via radar altimetry. The small seamount at 31.1º N, 58.2º N has not been investigated, but is very likely of Cretaceous age.

Fig. 8. Alternative traces of the Bermuda "hotspot" relative to the North America plate, as predicted by the plate/mantle motion models of Morgan (1983), Duncan (1984) and Mueller et al. (1993), with filled circles indicating the predicted hotspot location at 0, 50 and 100 Ma for each trace. Thin lines within the Mississippi Embayment are the Mississippi Valley graben faults, and open squares show the average location of the earliest (ca. 115 Ma) and latest (ca. 65 Ma) eruptions along a migratory path of igneous activity. CFA and GSD denote Cape Fear Arch and Great Stone Dome. Modified from Cox and Van Arsdale's (2002) Figure 1. See de Boer et al. (1988) and McHone (1996) for more detailed maps of Jurassic and younger igneous activity in the eastern US and adjacent Canada.

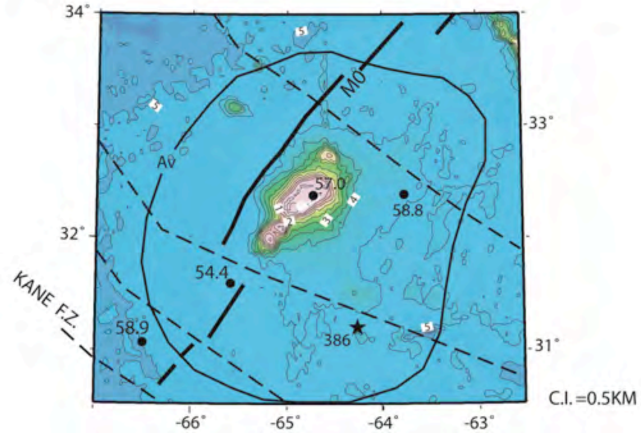
Fig. 9. Predicted temporal evolution in anomalous topography (top), heat flow (middle), and geoid (bottom) for a model lithosphere 90 km thick, moving at 15 mm/a across an underlying Gaussian-shaped heat source, with first contact at t=0. Dynamic uplift is assumed negligible. Adapted from Fig. 9 of Liu and Chase (1989). Note this "steady-state" model assumes the moving plate first encountered and then was modified by a fixed source—actual evidence shows no sign of plate migration over a fixed source, and no hotspot trace.

Fig. 10. Time correlation diagram, showing selected, more or less coeval tectonic or magmatic events at Bermuda and elsewhere in the world. Assembled from numerous sources, including Bailey and Woolley, 2005; Cande and Stock (2004), Dalrymple and Clague (1976), Kristoffersen and Talwani (1977), Oldow et al. (1989), Palmer and Geissman (1999), Rocchi et al. (2005), Rona and Richardson (1978), Sharp and Clague (2002), Southworth et al.(1993), Tarduno (2005), Winterer et al. (1989), and various data relating to Bermuda, as discussed in this paper.

Figure 1



44°
42°
40°
38°
36°
34°
32°
30°
28°
26°
24°
22°
20°
-74°
-72°
-70°
-68°
-66°
-64°
-62°
-60°
-58°
-56°
-54°
-52°
-50°
384
53.1
SAP
M0
200
400
M25
NES
600
800
2
0
50.1
MS
387
57.0
58.8
54.4
-200
HAP
51.0
58.9
386
51.8
50.7
49.0
47.0
49.0
-2
-400
600
21
KANE F.Z.
47.3
(49.9*)
NAP
34

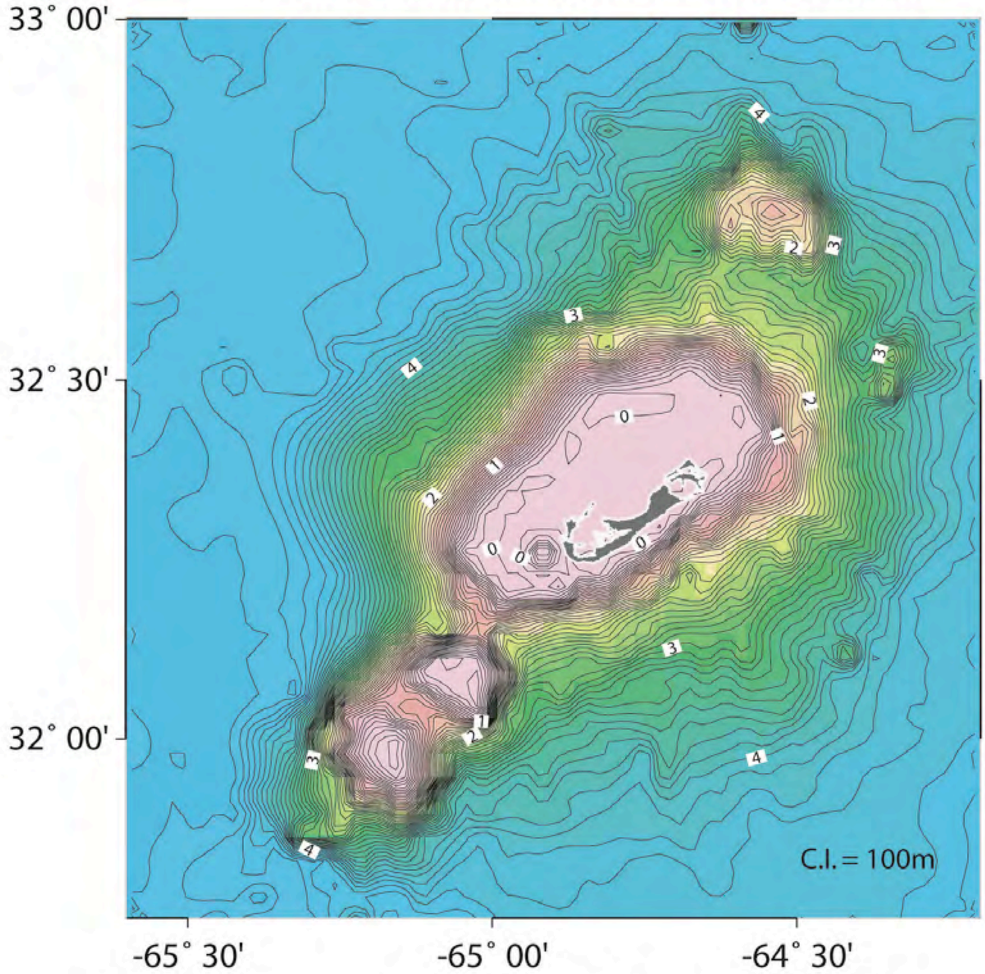


34°
33°
32°
31°
M0
Av
57.0
58.8
54.4
58.9
386
KANE F.Z.
-66°
-65°
-64°
-63°
C.I.=0.5KM

Figure 2

LANDSAT 7 VIEW OF BERMUDA

DETAILED BATHYMETRY AROUND BERMUDA



Numbers in km, based on Smith and Sandwell (1997)

Figure 3

NEAR SEA-FLOOR GEOLOGY
Sediment age, basement outcrops and crustal isochrons

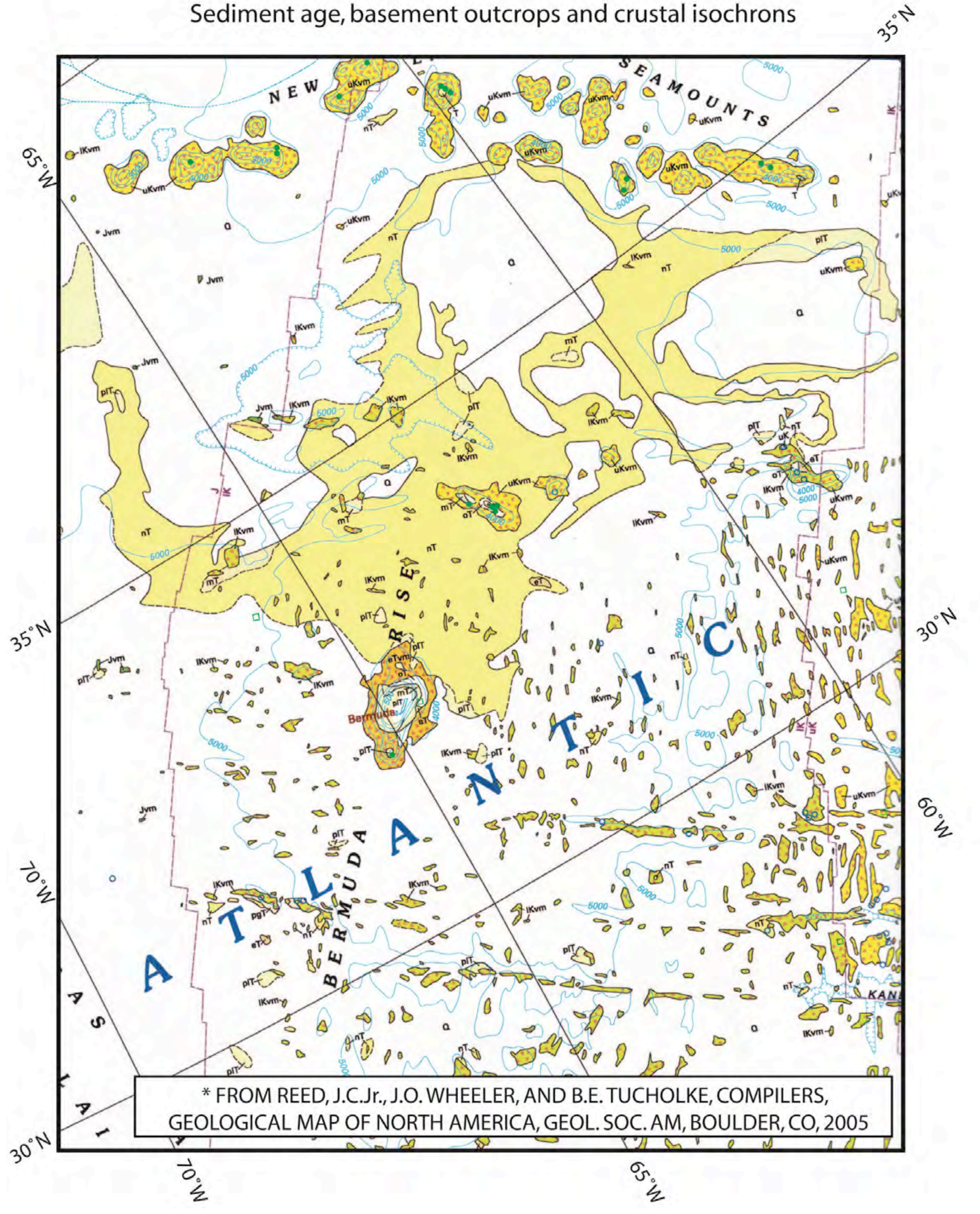
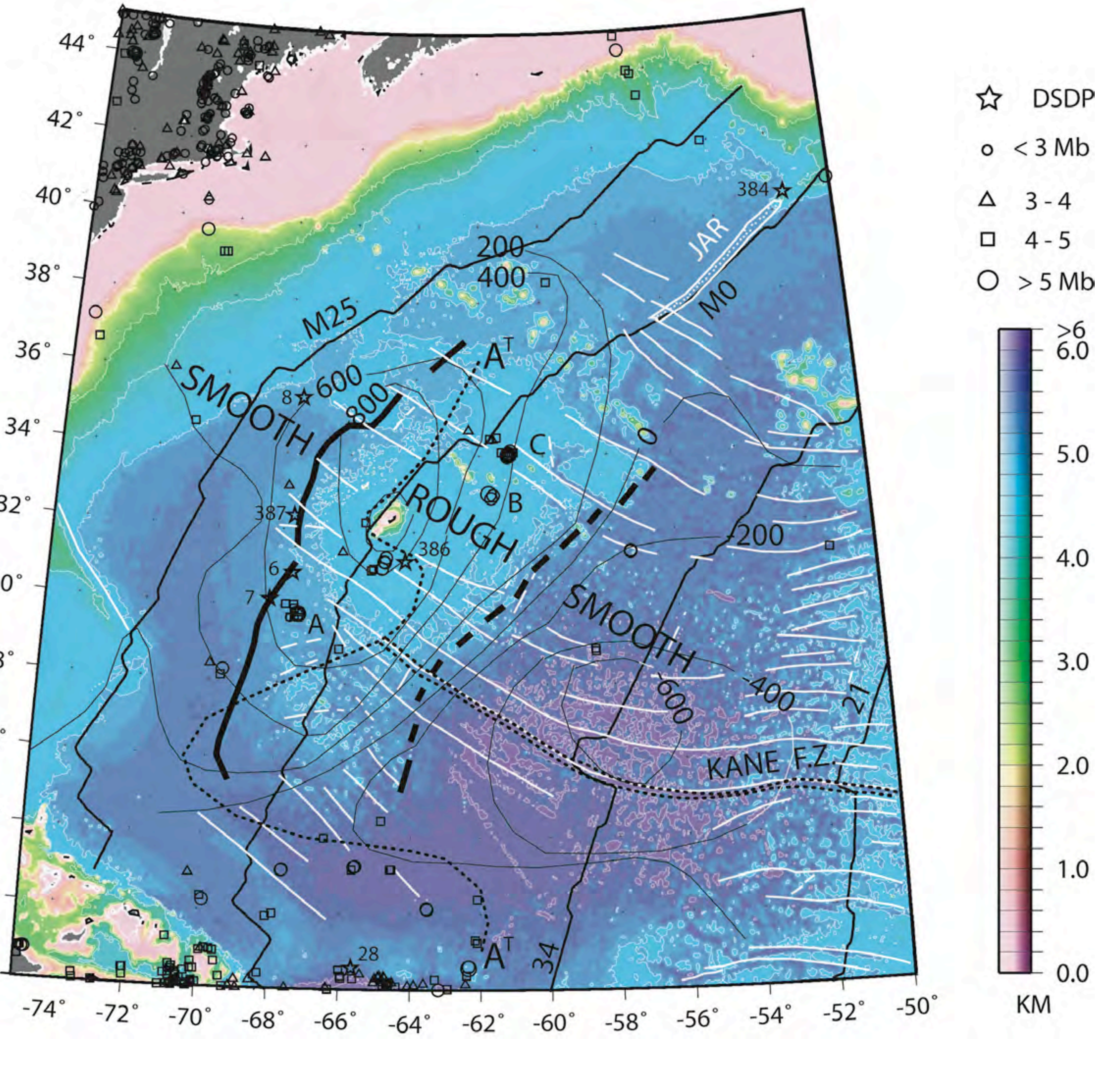


Figure 4



DSDP
< 3 Mb
3 - 4
4 - 5
> 5 Mb
>6
6.0
5.0
4.0
3.0
2.0
1.0
0.0
KM
44°
42°
40°
38°
36°
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32°
30°
28°
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-74°
-72°
-70°
-68°
-66°
-64°
-62°
-60°
-58°
-56°
-54°
-52°
-50°
M25
M0
JAR
384
200
400
600
800
SMOOTH
ROUGH
SMOOTH
A
B
C
8
387
6
7
386
28
A^T
A^T
KANE F.Z.
21
34
200
400
600

Figure 5

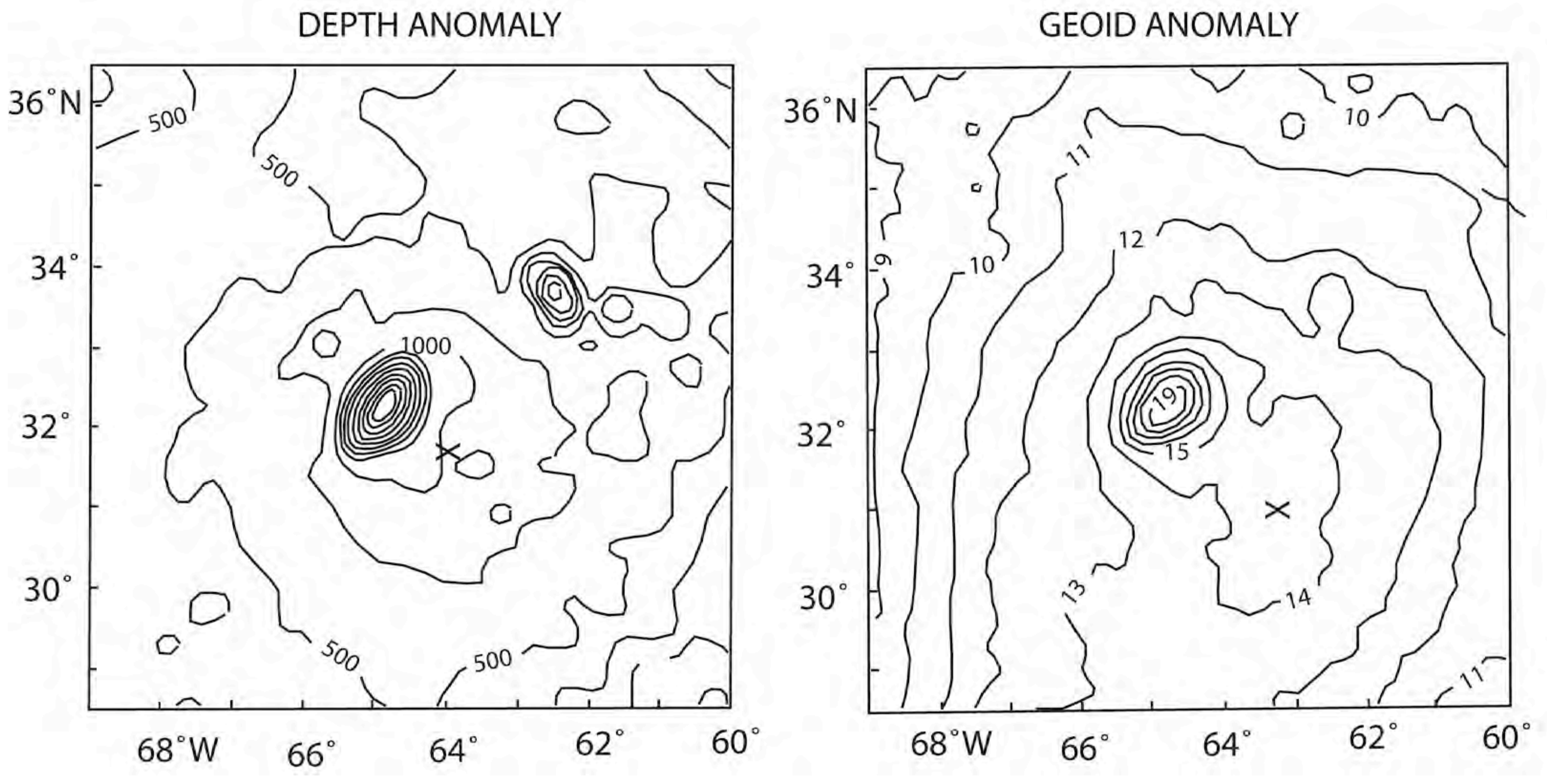

DEPTH ANOMALY
500
500
1000
500
500
36°N
34°
32°
30°
68°W
66°
64°
62°
60°
GEOID ANOMALY
10
11
12
9
10
19
15
13
14
11
36°N
34°
32°
30°
68°W
66°
64°
62°
60°

Figure 6

Depth Anomalies (Model GDH2 of DeLaughter et al., 2005)

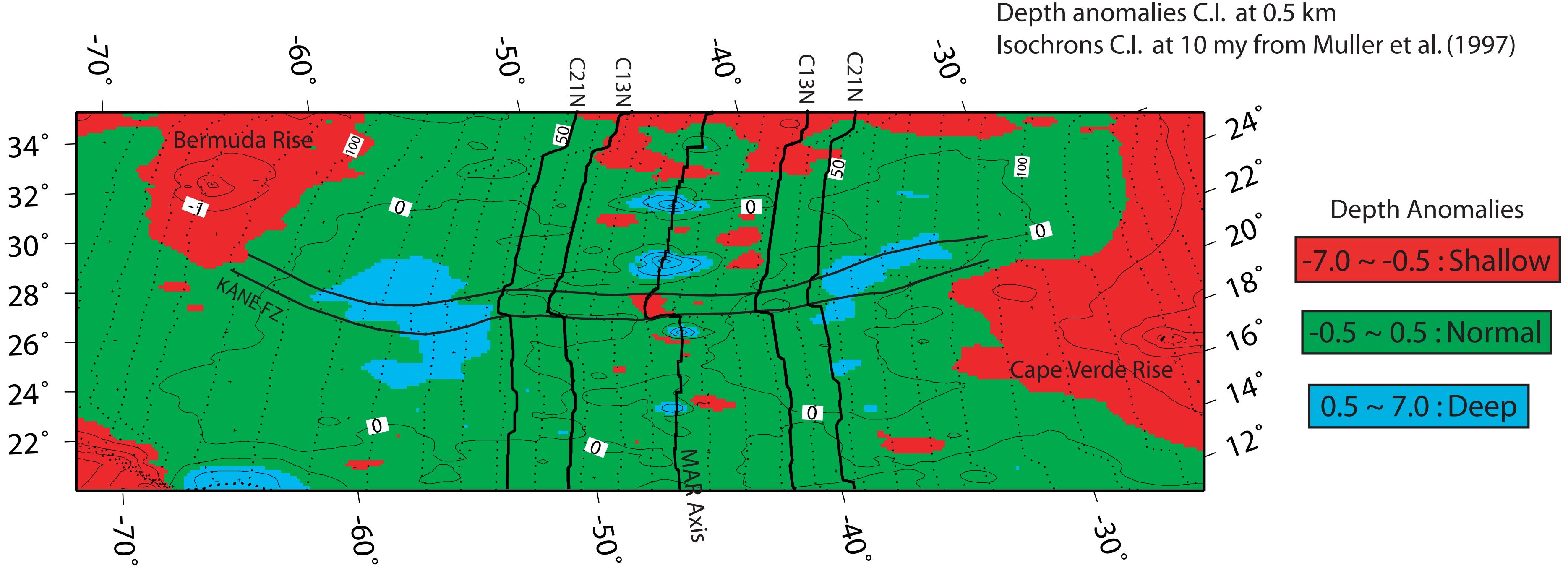


Figure 7

BERMUDA HOTSPOT-WHERE IT SHOULD BE TODAY

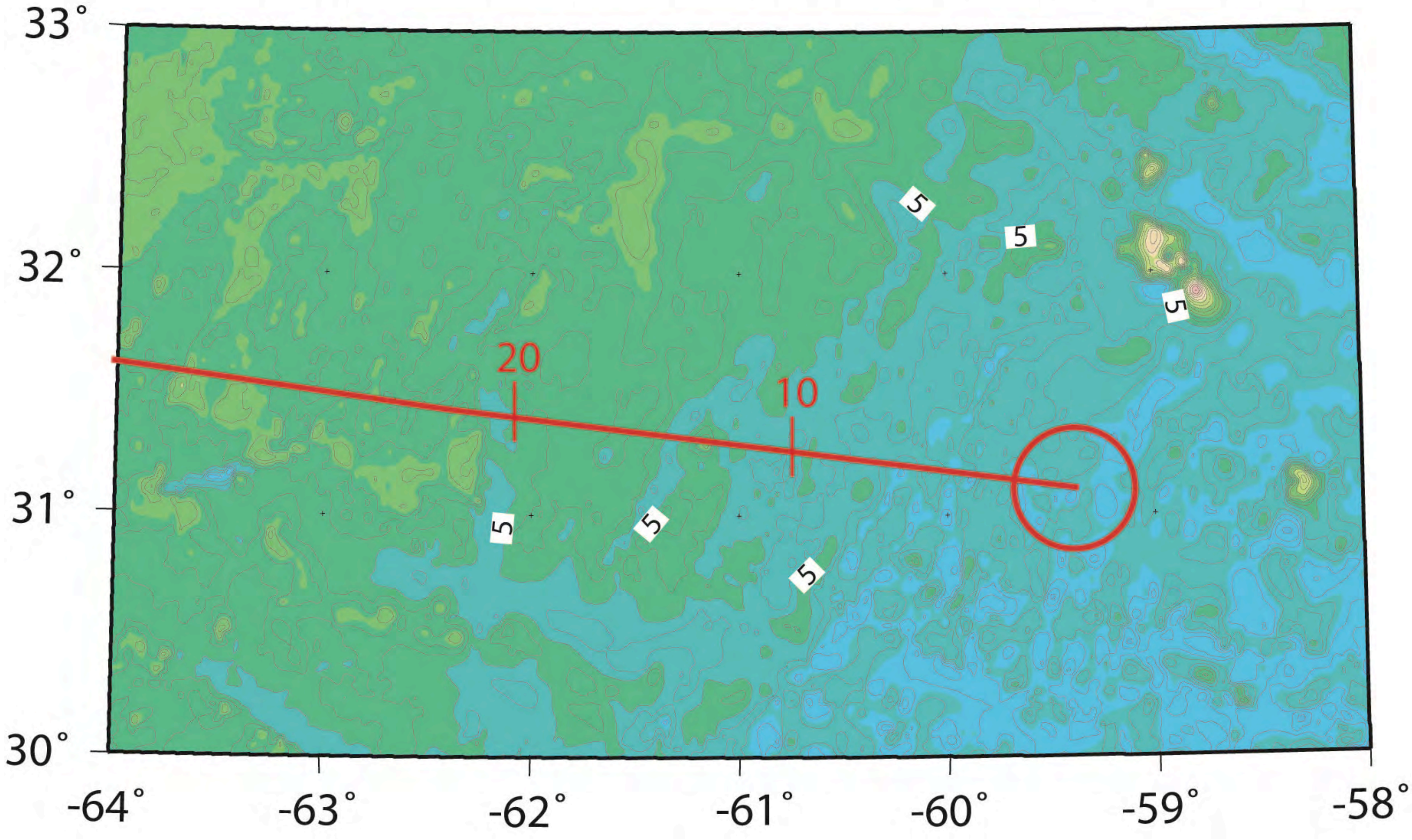


Figure 8

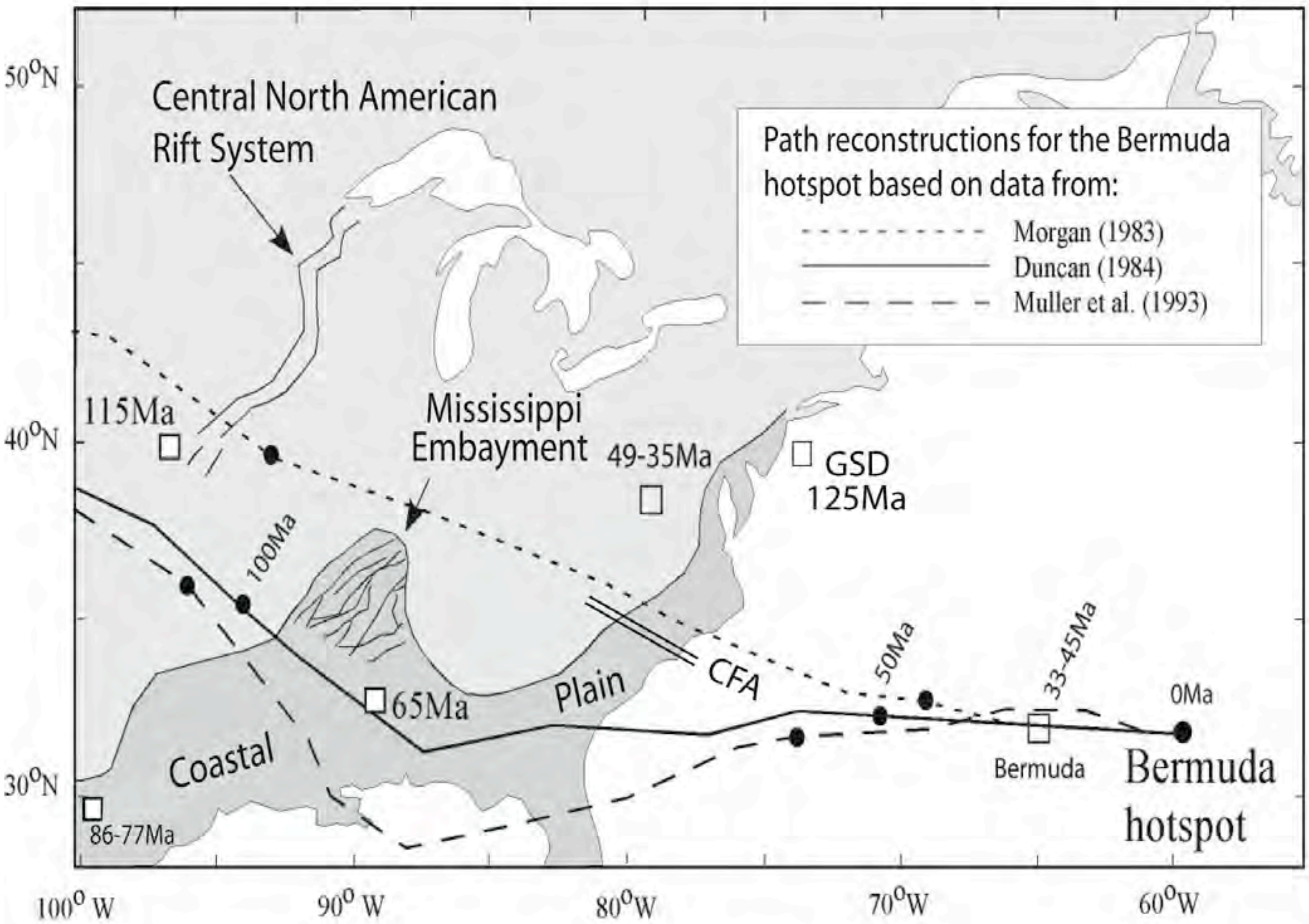

50°N
40°N
30°N
100° W
90° W
80°W
70°W
60°W
Central North American
Rift System
Path reconstructions for the Bermuda
hotspot based on data from:
Morgan (1983)
Duncan (1984)
Muller et al. (1993)
115Ma
Mississippi
Embayment
49-35Ma
GSD
125Ma
100Ma
65Ma
Plain
CFA
50Ma
33-45Ma
0Ma
Coastal
Bermuda
Bermuda
hotspot
86-77Ma

Figure 9

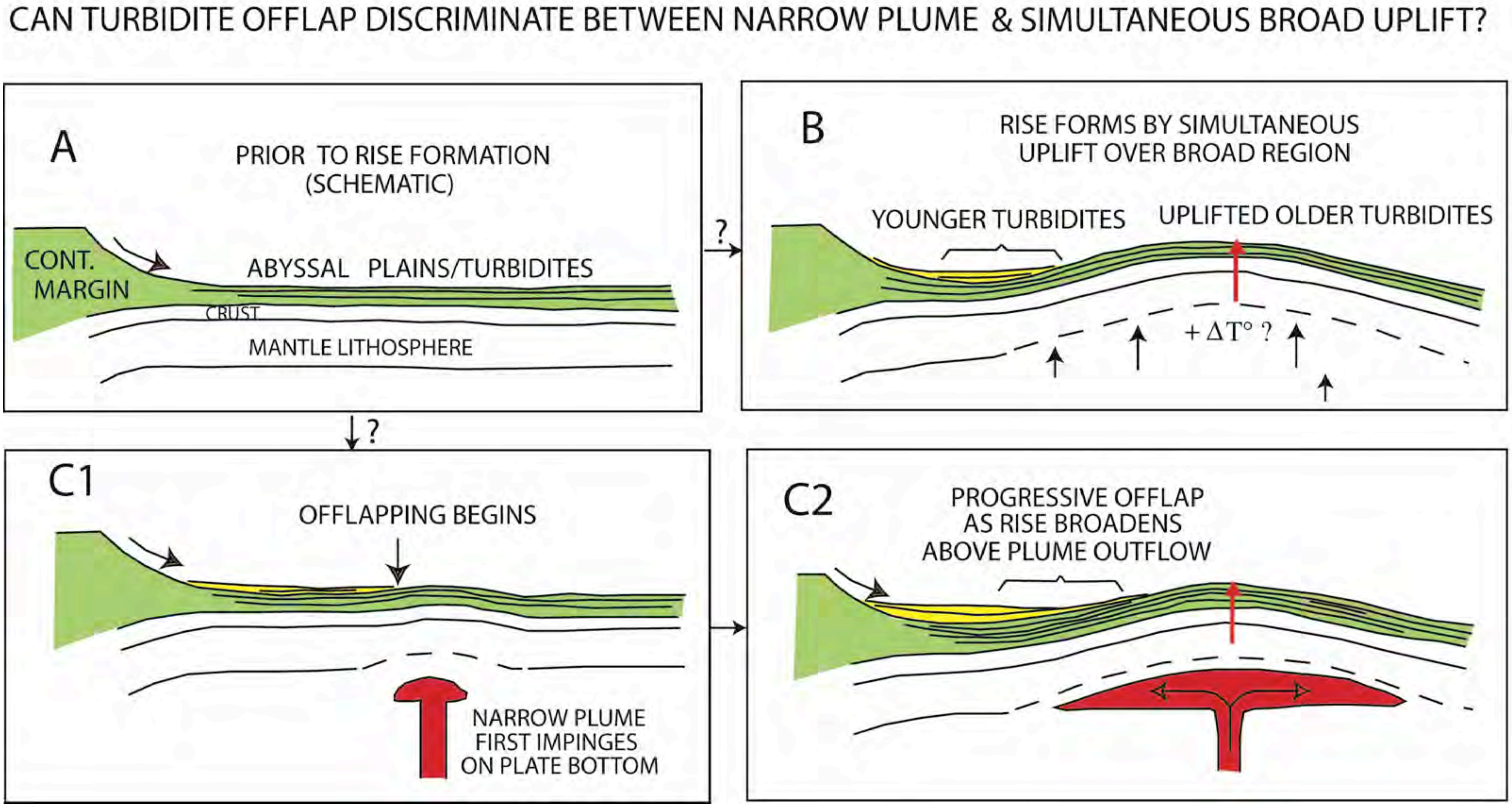
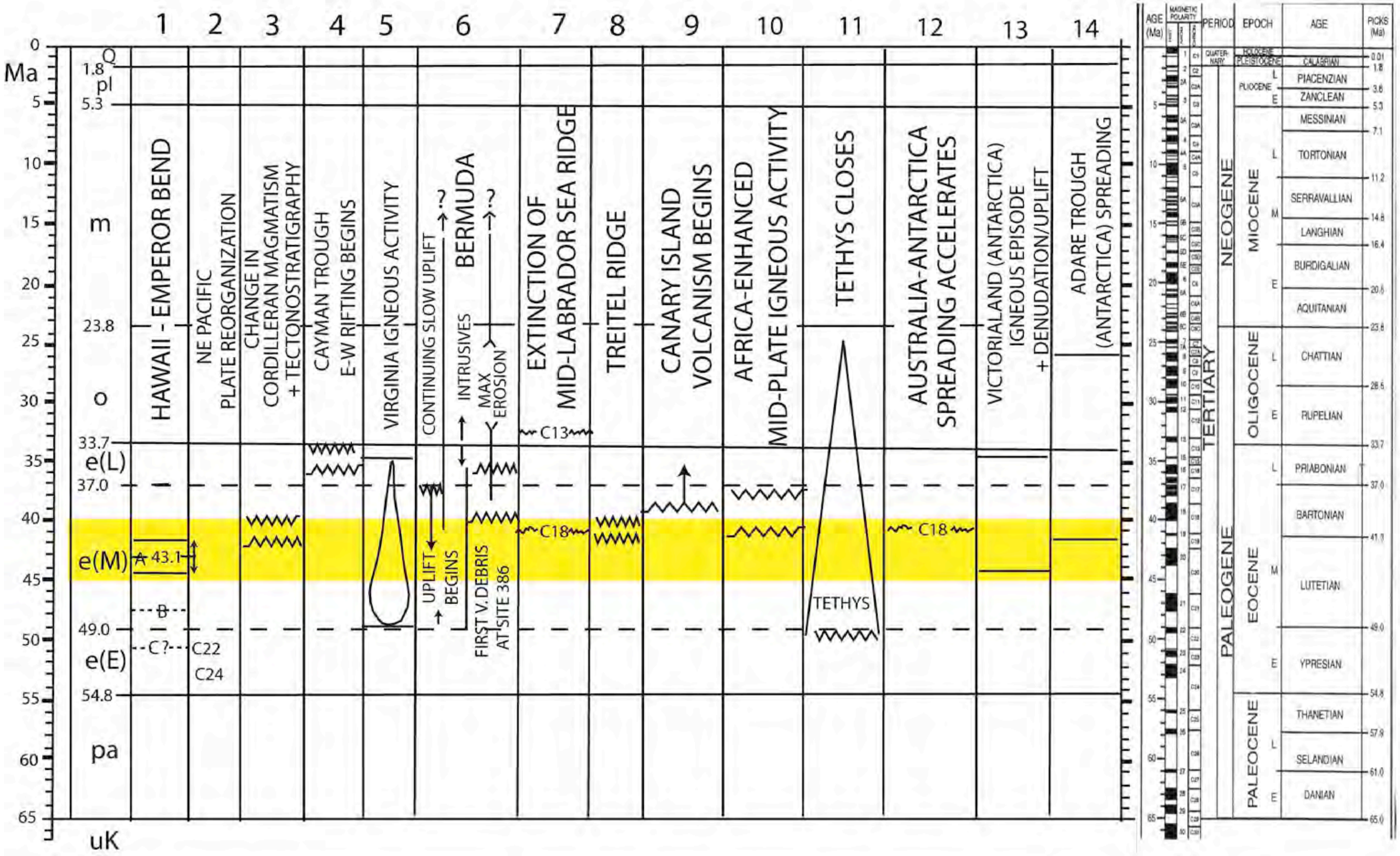

CAN TURBIDITE OFFLAP DISCRIMINATE BETWEEN NARROW PLUME & SIMULTANEOUS BROAD UPLIFT?
A
PRIOR TO RISE FORMATION
(SCHEMATIC)
CONT.
MARGIN
ABYSSAL PLAINS/TURBIDITES
CRUST
MANTLE LITHOSPHERE
?
B
RISE FORMS BY SIMULTANEOUS
UPLIFT OVER BROAD REGION
YOUNGER TURBIDITES
UPLIFTED OLDER TURBIDITES
+ ΔT° ?
?
C1
OFFLAPPING BEGINS
NARROW PLUME
FIRST IMPINGES
ON PLATE BOTTOM
C2
PROGRESSIVE OFFLAP
AS RISE BROADENS
ABOVE PLUME OUTFLOW

Figure 10



1
2
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6
7
8
9
10
11
12
13
14
Ma
Q
pl
m
o
e(L)
e(M)
e(E)
pa
uK
1.8
5.3
23.8
33.7
37.0
49.0
54.8
HAWAII - EMPEROR BEND
43.1
B
C?
C22
C24
NE PACIFIC
PLATE REORGANIZATION
CHANGE IN
CORDILLERAN MAGMATISM
+ TECTONOSTRATIGRAPHY
CAYMAN TROUGH
E-W RIFTING BEGINS
VIRGINIA IGNEOUS ACTIVITY
CONTINUING SLOW UPLIFT
UPLIFT BEGINS
BERMUDA
INTRUSIVES
MAX EROSION
FIRST V. DEBRIS AT SITE 386
EXTINCTION OF
MID-LABRADOR SEA RIDGE
C13
C18
TREITEL RIDGE
CANARY ISLAND
VOLCANISM BEGINS
AFRICA-ENHANCED
MID-PLATE IGNEOUS ACTIVITY
TETHYS CLOSES
TETHYS
AUSTRALIA-ANTARCTICA
SPREADING ACCELERATES
VICTORIALAND (ANTARCTICA)
IGNEOUS EPISODE
+ DENUDATION/UPLIFT
ADARE TROUGH
(ANTARCTICA) SPREADING
AGE (Ma)
MAGNETIC POLARITY
PERIOD
EPOCH
AGE
PICKS (Ma)
QUATERNARY
TERTIARY
NEOGENE
PALEOGENE
HOLOCENE
PLEISTOCENE
PLIOCENE
MIOCENE
OLIGOCENE
EOCENE
PALEOCENE
CALABRIAN
PIACENZIAN
ZANCLEAN
MESSINIAN
TORTONIAN
SERRAVALLIAN
LANGHIAN
BURDIGALIAN
AQUITANIAN
CHATTIAN
RUPELIAN
PRIABONIAN
BARTONIAN
LUTETIAN
YPRESIAN
THANETIAN
SELANDIAN
DANIAN