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Plate tectonic models for Indian Ocean "intraplate" deformation

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Abstract

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The equatorial region of the conventionally defined Indo-Australian plate has long been recognized as containing a type example of intense "intraplate" deformation. We trace the development of tectonic models for the area to illustrate techniques for the analysis of such deformation. The identification of anomalous seismicity near the Ninetyeast and Chagos–Laccadive Ridges demonstrated the existence of the deformation. Focal mechanisms from recent and historic earthquakes showed strike-slip motion occurring along the Ninetyeast Ridge; seismic moment data allowed the rate to be estimated. Similar studies showed north–south tension in the Chagos Bank region and north–south compression in the region between the Ninetyeast and Chagos ridges. Global plate motion studies indicated non-closure of the Indian Ocean triple junction, suggesting the conventional plate geometry was inadequate for a rigid plate description of the area. Gravity and marine geophysical data indicated intense north–south compressional deformation south of the Bay of Bengal. These observations are reconciled by a plate motion model in which Australia and India lie on distinct plates divided by a boundary that intersects the Central Indian Ridge near the equator. In this model Arabia, usually considered a separate plate, has negligible motion relative to India. The resulting Euler vector for Australia relative to Indo-Arabia lies just east of the Central Indian Ridge, and predicts approximately 0.5–1.5 cm/yr compression in the Central Indian Basin and 1.5–2 cm/yr strike-slip motion along the northern Ninetyeast Ridge, consistent with the seismological and geophysical data. In contrast to conventional oceanic plate boundaries, the boundary deformation is distributed over a wide zone. This diffuse nature may reflect either the boundary's recent inception or slow rate of motion. Analysis of seismicity and deformation in the boundary zone should offer insights into the mechanics of its development and its implications for the evolution of plate boundaries.

Introduction

As demonstrated by the contents of this symposium volume, deviations from the simplest model of plate tectonics, in which all deformation occurs at clearly defined boundaries separating rigid plates, can be dramatic and significant. The increasing sophistication of tectonic studies has improved our ability to identify and characterize anomalous regions for which the simple model is inadequate. Moreover, it is now possible to determine the point at which such regions should be treated not as deformation within a plate, but as a distinct but previously unresolvable plate boundary.

The most dramatic and best studied such region in the ocean basins is the area of the Indian Ocean between the Central Indian Ridge and the

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Sumatra Trench, which displays a level of "intraplate" seismicity and deformation unequaled elsewhere in the oceanic lithosphere. Seven magnitude seven earthquakes have occurred in this region since 1913, whereas elsewhere magnitude seven oceanic earthquakes not directly associated with plate boundaries are rare and apparently confined to passive continental margins or sites of intraplate volcanism such as Hawaii. Further evidence for the intensity of this deformation is provided by marine seismic profiling, which shows extensive faulting and folding of sediments and unusually high heat flow in the southern Bay of Bengal and the Central Indian Basin. Recent modeling of long-wavelength basement undulations and geoid anomalies indicate buckling of the entire lithosphere in response to north–south compressional stress.

In this paper we provide a broad overview of the seismicity and geophysical data and briefly discuss the development of tectonic models to describe this deformation. We then summarize the most recent model, which suggests a new plate geometry for the Indian Ocean in which the deforming zone is described as a diffuse boundary separating the Australian Plate from a combined Indo-Arabian plate. We conclude by discussing some of the possible implications of a diffuse oceanic plate boundary for Indian Ocean plate tectonics and plate boundary processes in general.

Seismicity

The unusual tectonics of the northern Indian Ocean are clearly demonstrated by the seismicity. Figure 1 shows earthquakes in the area between the Central Indian Ridge and the Sumatra Trench. This seismicity distribution differs from the usual pattern, identified early in the history of global seismology, in which earthquakes are generally concentrated along narrow zones we now know to be plate boundaries.

The locations and focal mechanisms of the earthquakes provide crucial data for characterizing the deformation. Recent seismological studies, including Stein and Okal (1978), Stein (1978), Wiens and Stein (1983; 1984), Bergman and Solomon (1984; 1985), and Wiens (1986) provide a detailed picture of northern Indian Ocean seismicitiy. Large earthquakes, including events of M 7.7 (1928) and M 7.2 (1939), occur along the northern Ninetyeast Ridge. Focal mechanisms determined for the 1939 event and for several recent events suggest left-lateral strike-slip faulting, with the orientation of one nodal plane coincident with the orientation of the Ninetyeast Ridge. Seismic moments estimated from surface wave spectra for the largest events, $4\text{–}8 \times 10^{27}$ dyne cm for the 1928 event and 2×10^{27} dyne cm for the 1939 event, suggest that the ridge has a level of seimicity comparable to transform faults elsewhere. The ridge itself, traditionally described as "aseismic" (Kennett, 1982), apparently represents a hotspot track formed approximately 80–40 Ma ago along a long offset transform fault (Morgan, 1972; Sclater and Fisher, 1974; Peirce, 1978). Thus, the locations of earthquakes suggest that the Ninetyeast Ridge represents a weak zone along which slip is concentrated. No direct evidence of recent slip is found in the morphology, although the blocky nature of the northern part of the ridge may be due, in part, to the recent deformation.

Several other events, including large earthquakes in 1949 (M 7.0) and 1955 (M 7.0), are located to the east of the Ninetyeast Ridge. Recent smaller events in this area show strike-slip focal mechanisms, apparently indicating left-lateral motion along north–south trending fault planes. This suggests that some of the left-lateral strike-slip motion found along the Ninetyeast Ridge may occur in a diffuse shear zone to the east.

Lower levels of seismicity also occur between the Ninetyeast and Chagos–Laccadive Ridges in the southern Bay of Bengal and the Central Indian Basin. Although large historical events occurred in 1913 (M 7.0) and 1918 (M 6.5), recent events have been smaller and generally show thrust faulting or strike-slip faulting with north–south compressional axes.

Seismicity in the Chagos–Laccadive region indicates a complex pattern of deformation, as focal mechanisms show thrust, normal, and strike-slip faulting. Chagos Bank is a region of particularly intense seismicity, with large earthquakes in 1912 (M 6.8), 1983 (M_S 7.6), and a swarm of moderate

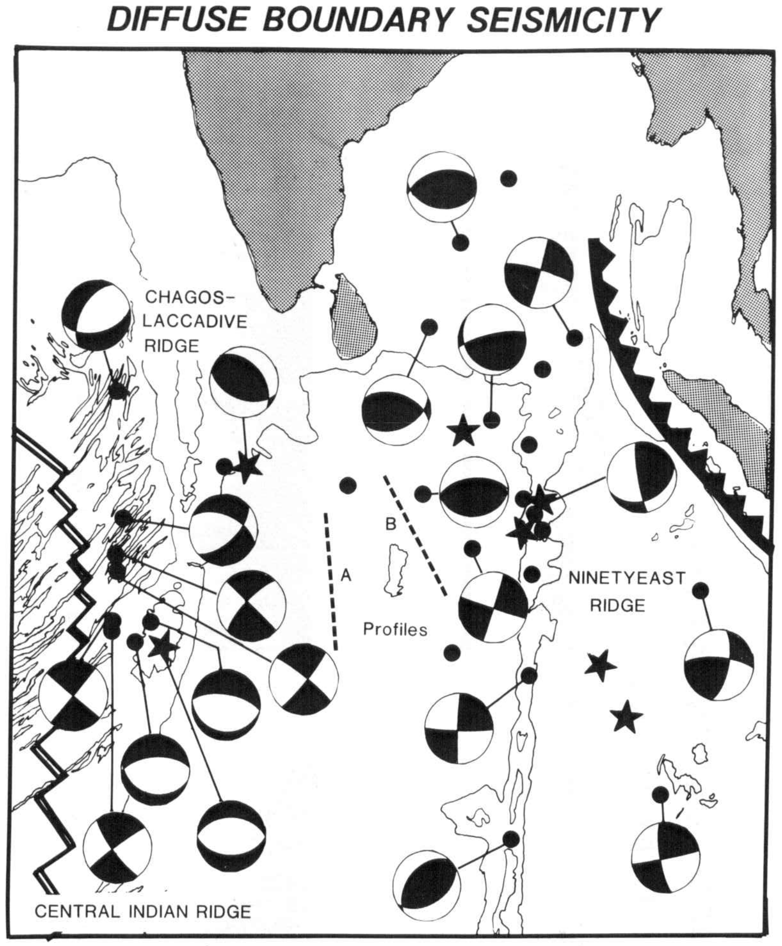

Fig. 1. Seismicity and focal mechanisms along the deforming zone between the Central Indian Ridge and the Sumatra Trench. Focal mechanisms are from Stein and Okal (1978), Stein (1978), Wiens and Stein (1983; 1984), Dziewonski et al. (1984), Bergman and Solomon (1984; 1985) and Wiens (1986). Earthquakes with magnitude greater than seven are denoted by stars; dots denote events with magnitude greater than six or events for which focal mechanisms have been determined. Only representative events are shown for the Chagos Bank sequences of 1967–1968 and 1983–1984.

sized earthquakes in 1965–1968. Focal mechanisms for most of the recent earthquakes show normal faulting with north–south oriented tensional axes. North and west of Chagos Bank, small earthquakes generally show strike-slip faulting with nodal planes coincident with fracture zones, but these events also show a north–south tensional axis orientation. In contrast, the 1944 (M 7.2) earthquake, 800 km northeast of Chagos Bank, shows thrust faulting with a north–south or northeast–southwest compressional axis orientation, suggesting a stress field more consistent with that found to the east.

Gravity and marine geophysical data

Geodetic and marine geophysical data provide further evidence of intense deformation in the region south of the Bay of Bengal. Marine seismic profiling (Eittreim and Ewing, 1972; Weissel et al., 1980; Geller et al., 1983) reveals widespread deformation of originally flat lying sediments (Fig. 2). The deformation includes reverse faults and unusual undulations of acoustic basement with wavelengths of approximately 200 km and relief of up to 3 km. The undulations and reverse faults both strike roughly east–west, suggesting they may re-

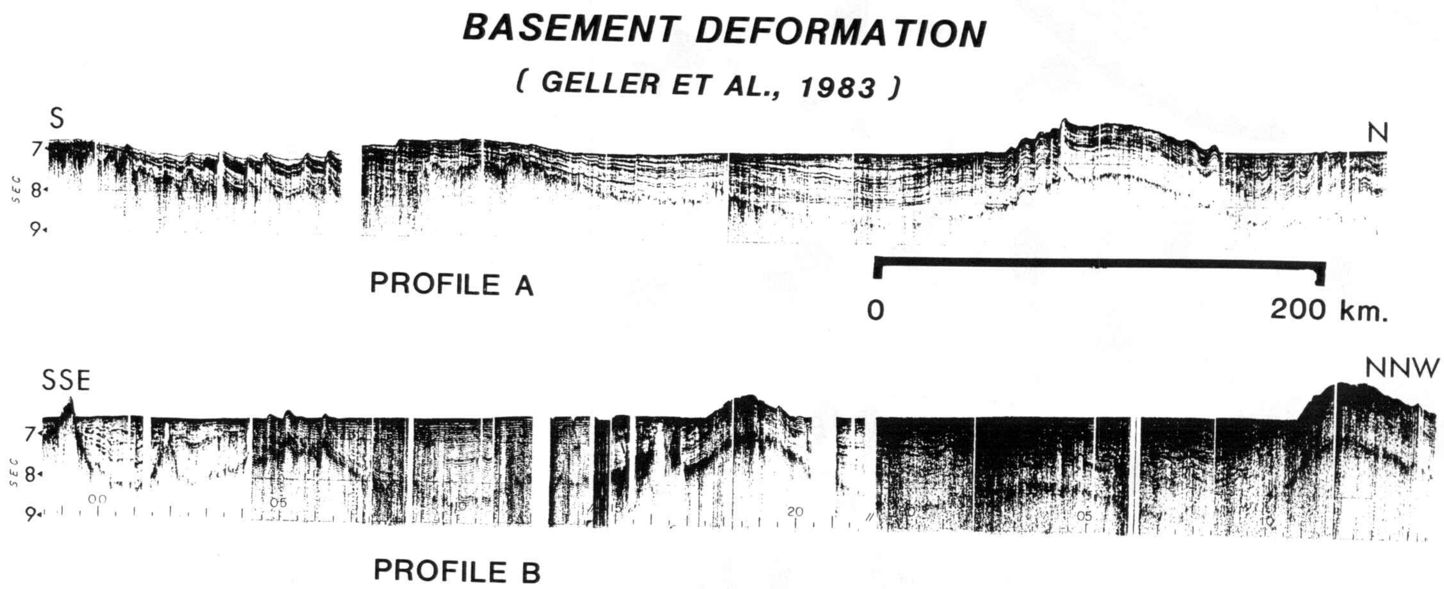

Fig. 2. Two long seismic reflection profiles from the southern Bengal Fan (see Fig. 1 for locations). Note the long wavelength undulations of basement and numerous reverse faults.

sult from north–south shortening. Deep-sea drilling and piston coring results indicate a Late Miocene age for a prominent unconformity separating deformed sediments from overlying deposits (Moore et al., 1974). The absence of synsedimentary deformation below the unconformity suggests this age provides an approximate date for the onset of deformation (Weissel et al., 1980). Substantial heat flow anomalies are also found, suggesting heat generation at shallow depths by some process possibly related to the deformation (Geller et al., 1983).

The basement undulations are accompanied by gravity anomalies, which give rise to linear east–west trending geoid anomalies in SEASAT data (Fig. 3). Although differing in specific details, lithospheric buckling models can successfully predict the wavelengths observed for the undulations (McAdoo and Sandwell, 1985; Zuber and Parmentier, 1985). These results generally suggest that the undulations arise from buckling of the entire lithosphere under a north–south horizontal compressive stress of several kilobars.

Tectonic models

As the data accumulated, a succession of models was offered to describe the deformation. Gutenberg and Richter (1954) discussed the "peculiarly isolated group of shocks near 2°S,

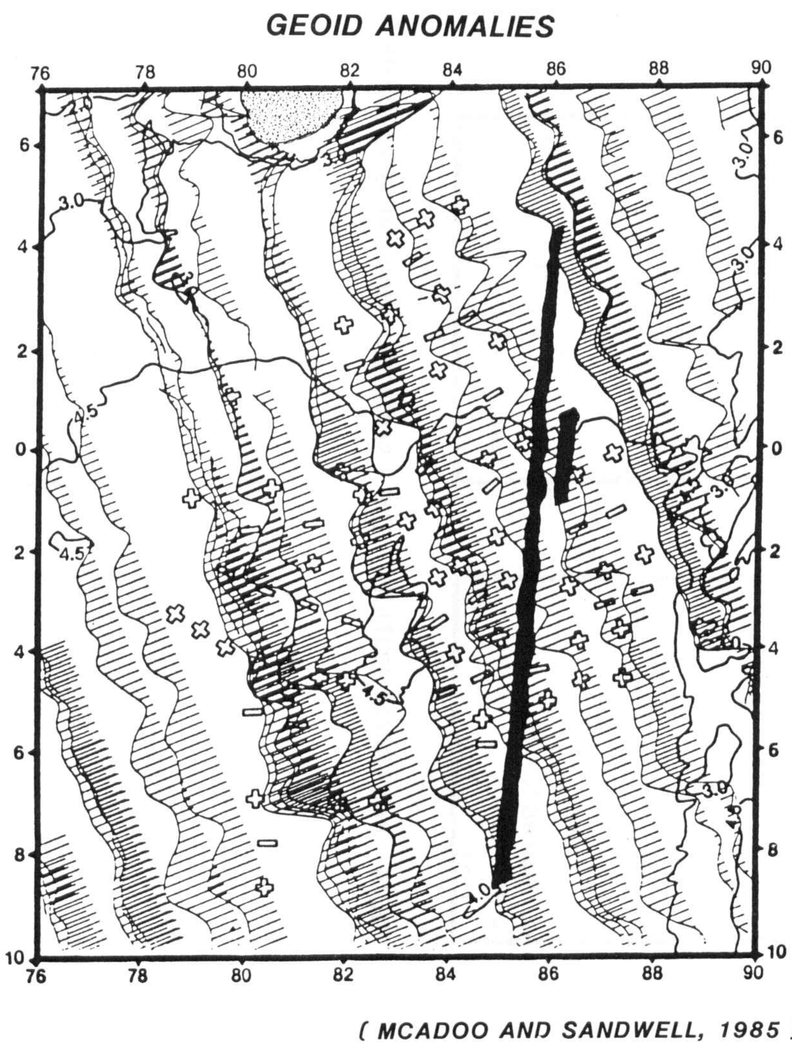

Fig. 3. Along track deflections of the vertical for SEASAT altimeter passes over the southern Bengal Fan. The geoid anomalies correlate with the linear long-wavelength undulations of basement, indicating buckling of the lithosphere.

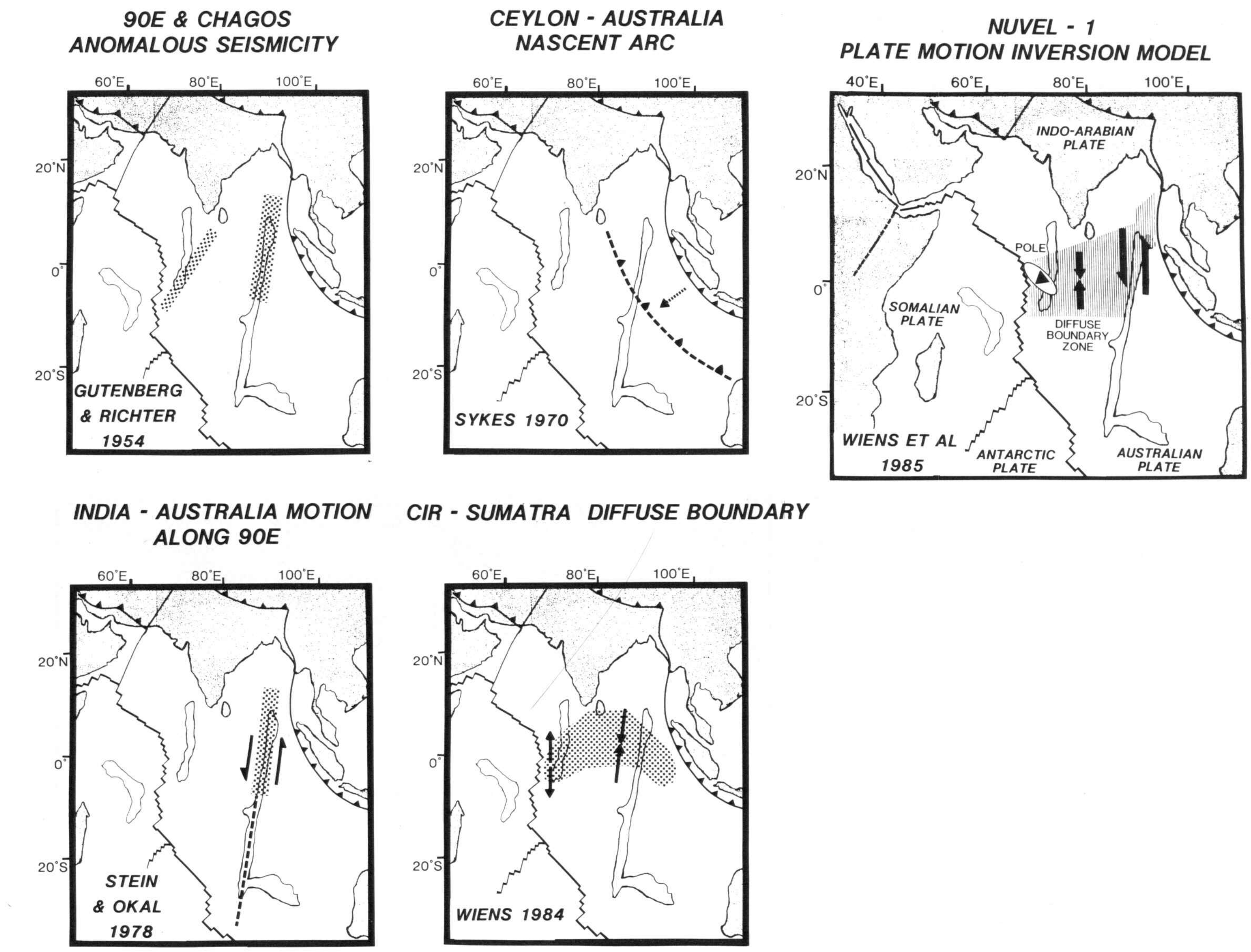

Fig. 4. Schematic diagrams illustrate the historical development of tectonic models for the northern Indian Ocean seismicity and deformation. For discussion see text.

89° E" and suggested the existence of a minor seismic belt along the northern Ninetyeast Ridge. They also suggested that a minor seismic belt trended southwest from Ceylon to the Central Indian Ridge (Fig. 4, top left). Though based on a limited dataset, this picture of the seismicity is remarkably similar to that accepted today. The earthquakes in the Ninetyeast Ridge region were also discussed by Stover (1966) and Rothe (1969).

In attempting to fit the earthquake observations into the newly developed theory of plate tectonics, Sykes (1970) suggested this seismicity reflected an incipient island arc stretching from Sri Lanka to Australia, perhaps resulting from the outward migration of the Indonesian Arc (Fig. 4, top right). However, later study by Stein and Okal (1978) and Bergman and Solomon (1985) showed that this model was incompatible with earthquake focal mechanisms, which generally show strike-slip faulting in the proposed region of incipient subduction.

Based on earthquake focal mechanisms and locations, Stein and Okal (1978) suggested that much of the seismicity reflects left lateral strike slip motion along the northern Ninetyeast Ridge (Fig. 4, middle left). They estimated the slip rate along the Ninetyeast Ridge as 2 cm/yr based on the summed moments of earthquakes. They further suggested this slip represents relative motion between the western half of the Indian Plate, which presumably encounters greater resistance along the Himalayan zone of continental collision, and the eastern half, which subducts normally beneath the Indonesian Arc.

Relative plate motion data provided further insight. Minster and Jordan (1978) noted that, assuming the conventional plate boundary geometry, Indian Ocean relative motion data were poorly fit. In particular, they found significant nonclosure around the Indian Ocean triple junction, which they attributed to internal deformation within the Indian Plate. They noted that splitting the Indian Plate along the Ninetyeast Ridge improved the fit to the data and predicted motion consistent with the seismological results for the northern part of the ridge. Stein and Gordon (1984) showed that the improvement of fit caused by splitting the Indian Plate was statistically significant.

However, a major difficulty was that the Ninetyeast seismicity and the possibly associated blocky morphology did not extend south of about 10°S. Thus the relationship of the Ninetyeast Ridge strike-slip motion to the conventional boundaries of the Indian plate was unclear. Wiens (1984; 1986), based on study of historical earthquakes, suggested that the Ninetyeast Ridge seismicity formed part of a continuous zone of deformation extending from the Central Indian Ridge to the Sumatra Trench (Fig. 4, middle right). Wiens further proposed that the previously enigmatic concentration of normal faulting near Chagos Bank and the compression to the east could both be explained by rotation of northern and southern segments of the Indian Plate about a pole located just east of Chagos Bank.

The most recent model (Fig. 4, bottom) was derived using relative plate motion data (spreading rates, transform azimuths, and earthquake slip vectors along the conventional boundaries) to test alternative plate boundary models independent of the "intraplate" seismicity. In this model, a diffuse plate boundary extends from the Central Indian Ridge to the Sumatra Trench, separating an Australian plate (AU) from the Indo-Arabian (IA) plate.

The Indian Ocean diffuse boundary is largely in accord with the previous proposal, but the incorporation of Arabia into part of the Indian plate may seem surprising at first. Past plate models, beginning with Wilson (1965) and Morgan (1968), have assumed relative motion between Arabian and Indian plates along the Owen Fracture Zone, which has been an active plate boundary in the past (McKenzie and Sclater, 1971). However, Laughton et al. (1970) pointed out that Carlsberg Ridge and Gulf of Aden plate motion data could be fit with a single rotation pole. Moreover, the seismicity along the Owen Fracture Zone is meager. As the largest earthquake is magnitude 5.6, the slip rate estimated from the summed moments of all historic earthquakes is about 0.25 mm/yr. Thus, on the basis of the seismic, gravimetric, and marine geophysical data, there is strong evidence of displacement between India and Australia, concentrated in the Central Indian Basin and along the Ninetyeast Ridge. On the other hand the low level of seismicity belies significant motion between Arabia and India along the Owen Fracture Zone. This hypothesis can be critically tested with plate kinematic data.

Figure 5 (top) shows the critical test using relative plate-motion data from the Gulf of Aden and the Carlsberg and Central Indian Ridges. The region south and west of these spreading centers was treated as a single Somalian (SO) plate, and the region to the north and east, which includes areas conventionally regarded as portions of the Indian and Arabian plates, as two plates divided by a boundary which is to be located. This plate geometry excludes possible complications due to poorly constrained Nubia–Somalia relative motion (Chase, 1978; Minster and Jordan, 1978; Stein and Gordon, 1984). The data for this three plate system were inverted with different locations assumed for the boundary between the plate to the southeast (Australian) and the northwest (Arabian). The minimum in squared error occurs for a boundary between 4°N and the equator. (The location could not be resolved better because of sparse data.) In contrast, a boundary location at the Owen Fracture Zone yields a significantly higher error and does not even represent a local minimum in the squared error.

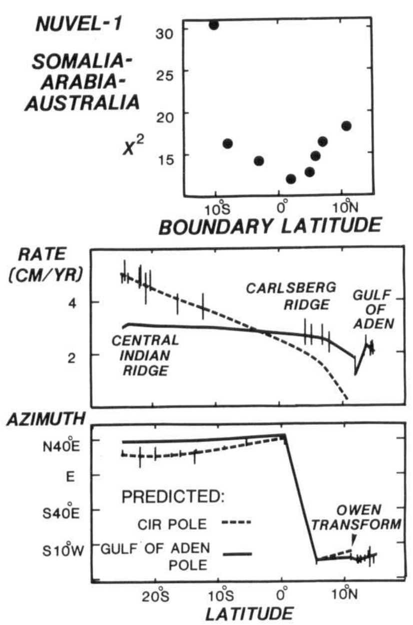

Fig. 5. Plate kinematic test for the best boundary location. (Top). Misfit to relative motion data as a function of the assumed location of the boundary between Arabian and Australian plates. The minimum misfit occurs for a boundary near the equator. (Bottom), Rate and azimuth data along the boundary separating Somalia from plates to the east. The Carlsberg Ridge spreading rates and the azimuth of the Owen Transform are better fit by the predictions of an Arabia–Somalia pole derived from Gulf of Aden data (solid line) than by an Australia–Somalia pole derived from Central Indian Ridge data (dashed line), suggesting that the Carlsberg Ridge is an Arabia–Somalia boundary.

This conclusion is quite robust; although it was first derived from the datasets of Chase (1978) and Minster and Jordan (1978), it is equally compatable with the NUVEL-1 dataset which contains many data published since 1978 (DeMets et al., 1985). Results similar to these from the three plate inversion are also obtained in tests inverting the entire global plate motion dataset.

Figure 5 (bottom) shows why the new model is preferred. The dotted lines represent spreading rates and transform azimuths predicted by an Euler vector derived only from Central Indian Ridge (Somalia–Australia) data. Similarly, the solid lines represent the predictions of an Euler vector derived from Gulf of Aden (Somalia–Arabia) data. Spreading rates along the Carlsberg Ridge and the orientation of the Owen Transform (the active segment of the Owen Fracture Zone, traditionally treated as a Somalia–India boundary) are better fit by the Euler vector derived from Gulf of Aden data than by the Euler vector derived from Central Indian Ridge data. This suggests that the Carlsberg Ridge data reflect motion between the same pair of plates that are separating in the Gulf of Aden.

Critical to this analysis are the four spreading rates from the Carlsberg Ridge. The four rates we use were chosen by Minster et al (1974) from sixteen profiles across the Carlsberg Ridge (McKenzie and Sclater, 1971, figs. 8–10]. All sixteen profiles were modeled by McKenzie and Sclater (1971) with spreading rates of 24 or 26 mm/yr., the lower rate applying to profiles at 7–10°N and the higher rate applying to profiles at 3–6°N. Minster et al. (1974) estimated rates of 24 to 28 mm/yr. for these profiles. All of these estimates significantly exceed the rates of approximately 15 mm/yr. predicted by the Central Indian Ridge Euler vector, but are in excellent agreement with the 25 mm/yr. rate predicted by the Gulf of Aden Euler vector.

The F-ratio test of additional boundaries (Stein and Gordon, 1984) was used to test whether the Owen Fracture Zone is an active plate boundary in addition to the newly identified boundary. The value of F for the Owen Fracture Zone boundary was well below the threshold for a statistically significant improvement, indicating an additional plate boundary was unjustified on the basis of the plate kinematic data. This test does not exclude motion as slow as a few millimeters per year along the Owen Fracture Zone, but only shows that any such motion is too small to be resolved by the current plate motion data. A low level of seismicity is present along the Owen Fracture Zone, including magnitude 5 earthquakes (Quittmeyer and Kafka, 1984), and a recent (April 7, 1985) M_s 4.7 strike-slip event (Dziewonski et al., 1986). However, this level of seismicity is more comparable to that on "inactive" fracture zones (Wiens and Stein, 1984) or intraplate bathymetric features (Stein, 1979) than to active plate boundaries. Thus both the seismicity and plate motion data suggest that motion along the Owen Fracture Zone is negligible.

When incorporated into a global plate motion model (DeMets et al., 1985), the new plate geometry helped resolve or reduce several long-standing problems for plate motions in the Indian Ocean basin. The new model reduced the non-closure of the Indian Ocean triple junction (Minster and Jordan, 1978), and reduced the total error in global plate inversions without including any additional free parameters. Furthermore, the new model produced a better fit to data along the Southeast Indian Ridge, which were poorly fit in past models.

Concentrations of intraplate earthquakes along the Southeast Indian Ridge (Bergman et al., 1984; Wiens and Stein, 1984) suggest some additional non-rigid behavior in the Indian Plate not explained by the diffuse boundary. However, the largest earthquake in that area is M_s 6.6, indicating the deformation is much less intense than that found along the diffuse boundary. As the stress orientations shown by the earthquake focal mechanisms are in agreement with the predictions of finite-element stress modeling for the Indian Plate, this zone may represent a secondary region of internal deformation within the Australian Plate (Wortel et al., 1985). This suggests that the new Indian Ocean plate geometry should be regarded not as a complete description of the Indian Ocean deformation, but rather as the simplest description of a complexly deforming area in terms of idealized, internally rigid plates.

The diffuse boundary

The plate motion data serve to locate the best position of the IA–SO–AU triple junction along the ridge system, but provide no information about the boundary location elsewhere. The previously discussed seismicity (Fig. 1) and other evidence of deformation suggest the boundary trends eastward from the Central Indian Ridge to the Ninetyeast Ridge region. Strike-slip seismicity along the Ninetyeast Ridge indicates that the boundary then trends northward to the Sumatra Trench in the Andaman Sea region (Fig. 4 bottom).

Inversion of the global NUVEL-1 dataset (DeMets et al., 1985) yields an AU–IA pole just east of the Central Indian Ridge. The motions and rates predicted by this Euler vector are consistent with those observed from earthquake focal mechanisms in the boundary region. The inversion results suggest left-lateral strike-slip motion along the Ninetyeast Ridge at a rate consistent with the cumulative seismic moment along that feature and predict convergence in the Central Indian Basin at a rate of about 0.5 to 1.5 cm/yr. Focal mechanisms in the Chagos–Laccadive region are harder to interpret as thrust, strike-slip, and normal faulting are all observed. This complexity can be ascribed in general to the close proximity of the rotation pole. In this region different poles within the confidence ellipse predict differing senses of motion; some of these poles are consistent with both normal faulting at Chagos Bank and the thrust faulting observed northwest of Chagos Bank.

The seismicity pattern suggests that the boundary is diffuse, with deformation extending over a wide zone which is difficult to define precisely. Near the Central Indian Ridge the seismicity zone is approximately 800 km wide, apparently extending from the equator to about 7° S. The deformation becomes more diffuse in the region between the Chagos–Laccadive and Ninetyeast Ridges, with the seismic zone extending at least as far north as Sri Lanka (7° N). The large width of the seismic zone in this region is consistent with the distribution of the long-wavelength lithospheric undulations observed in marine seismic profiles and the geoid, which indicate a very broad region of north-south compression. Strike-slip motion along the eastern portion of the diffuse boundary, while largely concentrated along the Ninetyeast Ridge, also appears diffuse, as left-lateral strike-slip earthquakes occur east of the ridge along other fracture zones. Further study of earthquakes in that region, including the larger historical (pre-1963) events, is needed for a definitive interpretation.

Implications

Diffuse plate boundaries have long been recognized in continental regions (e.g., the Great Basin). Recently such boundaries have also been recognized in oceanic regions; in addition to the boundary described here, other possibly diffuse boundaries include the North America–South America boundary (Weinstein et al., 1985), the Azores–Gibraltar boundary (Grimison and Chen, 1986), and boundaries in the Scotia plate (Forsyth, 1975; Farmer et al., 1982) and Caroline Plate (Weissel and Anderson, 1978; Hegarty et al., 1983) regions. All of these boundaries have several common characteristics, notably slow rates of relative motion or fairly recent inception or both. The Indian Ocean diffuse boundary also seems to display these characteristics.

Because of uncertainties in the time-history of deformation, the plate motion results can only give order of magnitude estimates for the total amount of shortening in the convergent region. Sedimentological data suggest a Late Miocene date for the onset of convergence in the Central Indian Basin (Weissel et al., 1980). If the present convergence rate has persisted over the past 5 m.y., the cumulative convergence would be about 50 km. As this convergence is taken up over a zone of lithosphere 1000–1500 km wide, the shortening would be approximately 3–5 percent. If we assume the plate motion convergence rate is applicable only to the past 3 m.y. (the averaging interval of the spreading rates in the inversion), a smaller shortening estimate of 2–3 percent results. These figures can be compared to results from numerical modeling of the topographic and geoid undulations in the Central Indian Ocean. Modeling the undulations as elastic buckling predicts only insignificant shortening, but more realistic modeling assuming a viscous rheology (Zuber and Parmentier, 1985) are consistent with larger estimates of the shortening.

The Indian Ocean diffuse plate boundary is the only oceanic convergent boundary lacking a morphologic trench or deep seismicity. This absence may be due to the recent development of the boundary and its slow rate of convergence. Although the mechanics of subduction initiation are unknown, perhaps subduction zones develop from such diffuse compressional tectonics if convergence is maintained over a long period of time or if the rate increases. Simple calculations show that a subduction zone may form if the convergence rate and stress exceed 1.3 cm/yr and 800 bars, respectively (McKenzie, 1977). Both conditions seem to be met in the Indian Ocean region. Thus the current deformation may represent the pre-subduction phase of an evolving convergent boundary.

Past study has shown that the development of a new plate boundary is generally contemporaneous with changes in the velocities of the surrounding plates (e.g. Gordon et al., 1978). Thus we may be observing a regional reorganization of plate boundaries and velocities in the Indian Ocean. Although the exact timing and sequence are unknown, a number of important tectonic events has occurred in the past 5–10 m.y. in the Indian Ocean. Seafloor spreading began in the Gulf of Aden about 10 Ma and propagated west of 45°E at 4–5 Ma (Laughton et al., 1970; Stein and Cochran, 1985). Thus the onset of convergence in the Central Indian Ocean in Late Miocene time may be part of a process of regional plate boundary reorganization contemporaneous with the separation of Arabia from Somalia and the cessation of motion on the Owen Fracture Zone.

Recent modeling results also shed light on the dynamical causes for the initiation of the new boundary. Stein and Okal (1978) suggested the deformation was the result of contrast between collisional resistance in the Himalayan region and normal subduction in the Sumatra arc. Finite-element modeling (Cloetingh and Wortel, 1985 and this issue) confirms this suggestion, and predicts stress levels of several kilobars near the northern end of the Ninetyeast Ridge where stresses are "focused". This suggests that plate boundary forces can cause intense deformation in the interior of plates, which may eventually result in the breakup of the plate and the initiation of a new plate boundary configuration.

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