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Buckle-controlled seismogenic faulting in peninsular India

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Abstract

As intraplate earthquakes are often not associated with major known faults their location as well as their timing is unpredictable. In peninsular India the larger ($M \geqslant 5.0$) events occur mainly on reverse faults in a series of belts ~400 km apart which are aligned roughly normal to the azimuth of convergence between the Indian and Eurasian plates. The location of the belts is controlled largely by the buckling wavelength of the lithosphere, and the seismogenic faults do not generate folding and sometimes result from it. There is consequently no need to postulate the creation of regularly spaced normal faults in an antecedent extensional phase, and the deformation is consistent with a plate-driving force such as gravity glide which is unlikely to reverse its polarity and which creates structures that are influenced by plate geometry at the leading edge. The thesis is potentially of value to seismic hazard mitigation as it identifies the zones that are most at risk.

1. Introduction

The Indo-Australian plate (Fig. 1) is separated by broad deforming zones from the Eurasian plate to the north and east. To the west it borders the Arabian plate along the Owen transform and to the SW the Somalia portion of the African plate along the Carlsberg Ridge. Until recently, the plate was viewed as a single geodynamic entity (e.g. Coblentz et al., 1998) but there is growing evidence that it consists of three component units—the Indian plate, the Australian plate and the newly identified Capricorn plate—which are separated by diffuse zones of deformation dominated either by convergence or by divergence (Royer and Gordon, 1997).

Seismicity is concentrated along the NW and NE plate margins and the Indonesian trench, with a weaker line of mainly strike-slip activity demarcating the Carlsberg Ridge. In peninsular India, the largest historical event is the 1819 Kutch (Khachch) earthquake, with an estimated M_w 7.8. During the twentieth century the 'stable continental region' witnessed several moderate-sized earthquakes ($M \geqslant 5.3$), some of which were apparently located within palaeorifts (Rajendran, 2000) and which were attended by variable levels of aftershock activity. In addition a number of zones characterised by numerous minor earthquakes have been identified (e.g. Sreedhar Murthy, 2002).

Various authors, notably Subrahmanya (1996) and Bilham et al. (2003), have linked crustal deformation and associated seismicity within the Indian plate to its collision with Eurasia. The present paper develops the

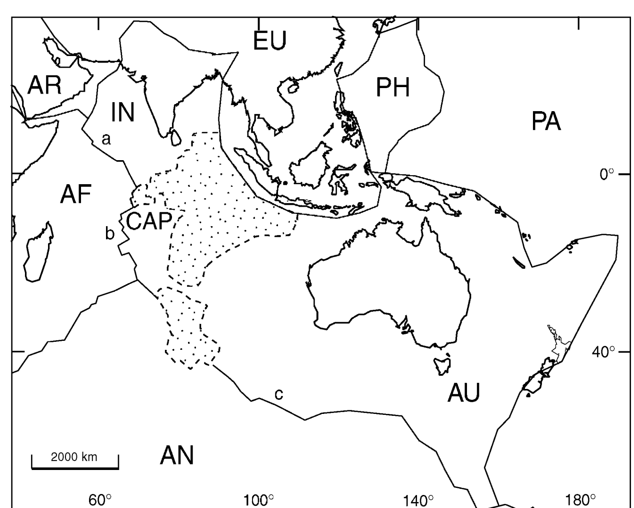

Fig. 1. Location of Indo-Australian plate. (a) Carlsberg Ridge, (b) Central Indian Ridge, (c) Southeast Indian Ridge. Stippled area shows diffuse boundaries between the Indian, Australian and Capricorn (CAP) plates, after Royer and Gordon (1997).

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Table 1
Earthquakes $M \geqslant 5$ in peninsular India 1900–2001

No.	Name	Lat, long (deg)	Date	Depth (km)	M	Source
1	Palghat	10.8, 76.2	25 Sep 2001	10	m_b 5.5	Brückner (1989)
2	Coimbatore	11.0, 77.0	28 Feb 1900		$M \geqslant 5.5$	Agrawal and Guzder (1972)
3	Ongole	15.6, 80.1	27 Mar 1967	13	m_b 5.4	Bendick and Bilham (1999)
4	Koyna-Warna	17.1, 73.6	8 Dec 1993	25	M_w 5.1	Bilham and Gaur (2000) and [8]Brückner (1989)
		17.1, 73.7	12 Mar 2000	*	M_w 5.0	Bilham and Gaur (2000) and [8]Brückner (1989)
		17.2, 73.7	2 Sep 1980	*	M_s 5.5	Brückner (1989)
		17.2, 73.5	1 Feb 1994	10	m_b 5.0	Bilham and Gaur (2000) and [8]Brückner (1989)
		17.3, 73.6	20 Sep 1980	*	m_b 5.3	Brückner (1989)
		17.4, 73.7	17 Oct 1973		M 5.1	Bilham and Gaur (2000)
		17.4, 73.9	5 Sep 2000	10	m_b 5.4	Brückner (1989)
		17.7, 73.9	10 Dec 1967	5	M_s 6.5	Bilham and Gaur (2000) and [8]Brückner (1989)
5		17.4, 77.5	29 Oct 1993	10	m_b 5.0	Brückner (1989)
6	Bhadrachalam	17.9, 80.6	13 Apr 1969	14	m_b 5.3	Bendick and Bilham (1999)
7	Killari/Latur	18.1, 76.4	29 Sep 1993	5	M_w 6.1	Bazant and Cedolin (1991) and [5]Bilham and Gaur (2000)
8		20.6, 71.4	24 Aug 1993	24	m_b 5.0	Brückner (1989)
9	Satpura	21.1, 75.8	14 Mar 1938	40	M_w 6.3	Bilham and Gaur (2000)
10	Broach (Bharuch)	21.7, 73.0	23 Mar 1970	11	M_w 5.4	Bilham et al. (2003) and [5]Bilham and Gaur (2000)
11	Midnapore	21.7, 88.0	15 April 1964	26	m_b 5.5	Bendick and Bilham (1999)
12	Balaghat	22.0, 80.0	25 Aug 1957		M 5.5	Bilham and Gaur (2000)
13	Jabalpur	23.1, 80.1	21 May 1997	36	M_w 5.8	Bilham and Gaur (2000)
14	Bhuj	23.3, 70.2	26 Jan 2001	10	M_w 7.7	Biswas and Majumdar, 1997 and [8]Brückner (1989)
15	Anjar	23.3, 70.0	21 Jul 1956	15	M_w 6.0	Bilham, 1998
16	Son valley	24.0, 82.0	2 Jun 1927	~35	M_w 6.4	Bilham et al. (2003) and [6]Bilham and Gaur (2000)
17		24.3, 69.9	7 Apr 1985	8	m_b 5.0	Brückner (1989)
18	Mt Abu	24.6, 72.4	24 Oct 1969	15	m_b 5.3	Bilham (1998) and [9]Cazenave et al. (1987)

Magnitudes as reported; M_w is given whenever available. Published default depths of 33 km are marked by * ; blanks signify no data available. Sources: (1) Rajendran and Rajendran (1996); (2) Rajendran and Rajendran (1999b); (3) Biswas and Majumdar (1997); (4) Chung and Gao (1995); (5) Rajendran and Rajendran (1999a); (6) Mandal et al. (2000); (7) Thakur and Wesnousky (2002); (8) USGS National Earthquake Information Center website (21 November 2003); and (9) Cloetingh and Wortel (1986).

proposal (Vita-Finzi, 2002) that large earthquakes in peninsular India tend to occur on one of a series of elastic buckles resulting from plate convergence. It complements the seismic evidence with neotectonic data and it suggests that some of the reverse faults on which the earthquakes nucleate are a consequence of buckling.

2. Intraplate seismicity

In peninsular India, over 20 earthquakes with $M \geqslant 5$ have been recorded instrumentally or have been described during the 20th century. They are listed in Table 1. Magnitudes are given as published, with preference given to M_w values where available. Similarly, depths are reported on the understanding that few of them are sanctioned by synthetic wave modelling. Most of the earthquakes are identified in the table by familiar names and all of them by latitude/longitude values for the epicentre of the main event. Two events which slightly postdate the 20th century are included because they add useful detail to the review. The 2001 Bhuj earthquake (#14, Table 1; Fig. 2), which was at the time of writing the largest recorded in the peninsula since 1900, occurred close to the Anjar earthquake of 1956 and the Kutch 1819 event; the 25 September 2001

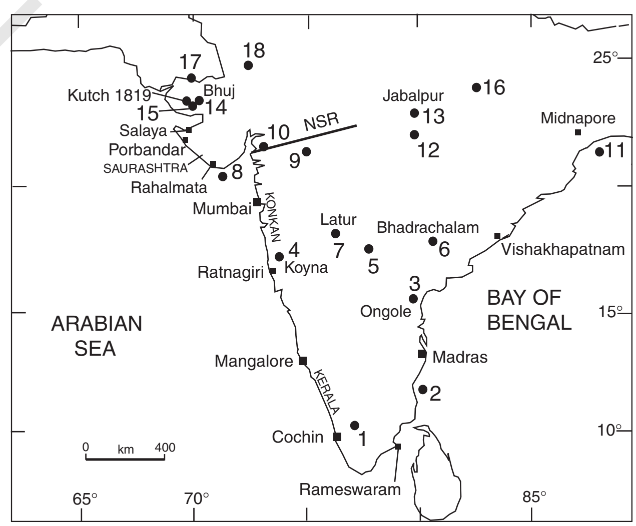

Fig. 2. Location of earthquakes (see Table 1) and places discussed in text. NSR: Narmada–Son rift.

event (#2) supplements the scanty record of the Palghat Gap.

The Palghat Gap (Rajendran and Rajendran, 1996) itself, which strikes roughly E–W, is represented in Table 1 by the Coimbatore earthquake of 1900 (#1). The

Palghat Gap has also witnessed the Waddakancheri events of 2 December 1994 (M_L 4.3), 25–26 February 1993 (M_L 3.6), and 15 March 1989 (M_L 3.0), the first of which, to judge from field data, was located on an E–W fault (Rajendran and Rajendran, 1996). The 1967 earthquake at Ongole (#3) is associated with a reverse fault striking 098°. The 1969 Bhadrachalam event (#6) was on a reverse fault striking 037° (Biswas and Majumdar, 1997; Talwani and Rajendran, 1991).

Koyna is known to seismologists for the possible influence of its dam on the timing and magnitude of the local seismicity. Several $M > 5$ events have been recorded there in the 20th century. The 1967 (M_s 6.5) earthquake ruptured the northern part of a NNE–SSW trending fault, and it was followed by six $M \geqslant 5.0$ aftershocks. An M_w 5.1 event on 8 December 1993 activated the southern part of the same fault and ruptured another fault striking NNE–SSW and about 20 km south of Koyna. In the light of these and earlier events, Mandal et al. (2000) conclude that, although there is some normal faulting on structures striking NW–SE, seismicity in the Koyna area is concentrated at a depth of about 10 km on two left-lateral strike slip faults aligned NNE–SSW.

The 1993 Killari (Latur) earthquake (M_w 6.2) had a seismic moment M_o of 1.8×10^{18} Nm. Slip measured 0.8–2.1 m on a reverse fault striking 135°; drill hole evidence suggests that the event was the latest of at least six within a pre-existing shear zone (Rajendran and Rajendran, 1999b). Trenching shows evidence for SW–NE thrusting during the 1993 and earlier events; historical seismicity, including an event at Ter, 40 km NW of Killari, about 1500 y BP and several earthquakes of MM = III–IV, are taken by Rajendran and Rajendran (1999b) to indicate a NW fault alignment.

The Broach earthquake has been ascribed to inversion under compression within the Narmada–Son rift (Chung, 1993), with thrust-dominated strike–slip movement on an E-W fault (Mandal et al., 2000). The focal-plane solution for the Jabalpur earthquake of 21 May 1997 showed reverse faulting with a strike–slip component on a structure striking 080°; the few aftershocks also indicated a steep fault with a SE dip. Both the Satpura (M_w 6.3) and the Jabalpur events appear to have had an unusually deep focus. The Son Valley 1927 earthquake is thought to have nucleated on the same south (Narmada) fault as the Jabalpur event, namely a thrust striking 61°; the Midnapore earthquake of 1964 was on a reverse fault striking 21° (Biswas and Majumdar, 1997; Mandal et al., 2000).

The Kutch (Kachchh) region is only 250 km inland of the western margin of the India plate (Thakur and Wesnousky, 2002) but its structural grain is predominantly E–W and is thought to embody Mesozoic normal faults which have been reactivated as reverse faults (Rajendran, 2000). The 2001 event occurred on a reverse fault striking 112°, part of a system which had displayed surface folding driven by blind reverse faulting in 1819 (Bilham, 1998; Ellis et al., 2001). The Mt Abu event of 24 October 1969 had a reverse mechanism with a left lateral component on a nodal plane striking 68°, as did the Anjar event of 21 July 1956 on a fault striking 55° (Chung and Gao, 1995).

In short, the majority of the large events of peninsular India during the 20th century have occurred on reverse faults indicative of shortening on azimuths between NW–SE and NE–SW or on strike–slip faults oriented NNE.

3. Neotectonic belts

Over the years several attempts have been made to identify areas or belts of heightened seismicity which would shed light on peninsular geodynamics. Murthy (2002), for example, compared earthquake distribution with topography, gravity and tectonic lineaments and noted that events of M 4–6 were concentrated on the west coast, the eastern Ghats, the Narmada–Son lineament and a western region comprising Saurashtra and the Kutch peninsula. Mahdevan (1995) paid special emphasis to what he termed deep continental structures; Rajendran (2000) likewise noted that most earthquakes in the stable continental part of India are clustered around pre-existing structural features, including the Kutch and Narmada rifts, although he also recognised mid-cratonic events, such as the Killari 1993 earthquake, which occur in areas where no rifting has developed since the Precambrian.

Other workers have complemented seismic records with gravity and neotectonic data in order to identify zones of active deformation. Ramasamy (1989) drew attention to coastal convexities at the ends of a lineament running east from Cochin and another linking Mangalore with Madras and, having shown that these features were associated with seismicity and river displacement, he interpreted them as tectonic upwarps which were at least partly Quaternary in age. The second of these lineaments corresponds broadly with a line of active buckling which has been identified by Subrahmanya (1996) close to latitude 13° N on the basis of displaced and incised river channels, coastal uplift and progradation, positive gravity, a thin crust and active microseismicity.

Further evidence of localised deformation was obtained by Bendick and Bilham (1999), who used tide-gauge and levelling data to infer uplift of Mangalore relative to Cochin (Table 2) and who recorded a series of Quaternary synclinal structures trending ENE on the West Indian coast between about 8° N and 20°. A study of 12 tide-gauge records for India by Emery and Aubrey (1989) identified five dependable sequences. Setting eustatic and local depositional factors aside, the data for Mangalore (1953–1976) and Cochin (1939–1982) confirmed that the former had been uplifted relative to the latter (1.3 mm/yr vs–2.1 mm/yr). Relative subsidence was recorded at Madras (1916–1982: −0.4 mm/yr), Mumbai (Bombay) (1878–1982: −0.9 mm/yr) and Vishakhapatnam (1937–1982: −0.7 mm/yr).

The search for additional evidence for neotectonic deformation inland is hampered by volcanicity, erosion and deposition. Nevertheless the Narmada valley has yielded geomorphological and stratigraphic information on fault movement within the Narmada–Son rift which indicate Holocene inversion in response to N–S compression (Chamyal et al., 2002). On the coast, palaeoshorelines offer scope for extending the skimpy tidal record, although many of the published age/height values present serious problems of interpretation, and postglacial hydroisostatic adjustment could account for the Holocene emergence by ~1 m reported at various locations by Brückner (1989).

In Saurashtra, fossil shorelines point to sustained uplift for the last 125,000 years (Chamyal et al., 2003); Brückner (1989) reports post-Pliocene marine deposits up to 10 m above sea level near Porbandar, where a series of ^{14}C ages (Table 3) indicates emergence of 3.5 m between 6400 and 7000 yr ago at an average rate of ~6 mm/yr. Even if probably exaggerated by the quirks of beach deposition, the results suggest that there was net uplift at this location. Corals dating from 6200–7100 yr BP at Salaya (22° 21′N) confirm these estimates by showing that emergence since they accumulated has exceeded 4.2 m (Gupta and Amin, 1974; Somayajulu et al., 1985). ^{14}C dating also points to emergence at Porbandar relative to locations north and south by up to 5 m in 6500 yr. that is about 1 mm/yr: for instance, the ~6760 yr BP waterline is at 5–6.4 m at Porbandar and 2.8 m about 54′ of latitude to the south. Radiocarbon dating of beachrock, generally a good indicator of intertidal waterlines, gives a similar trend. Table 4 shows a difference in height above high water of coeval deposits at Mumbai (~19°N) and at Ratnagiri (~17°N) of about 4.5 m in ~2500 yr (1.8 mm/yr) over a distance of 1°54′. There is some evidence of Holocene subsidence off southern coastal Kerala, on the Konkan coast and on Rameswaram island (Brückner, 1989).

Evidence for Holocene uplift offshore is reported by numerous authors, but unambiguous intertidal deposits are difficult to identify in the published lists. Table 5 lists coeval age pairs from the compilation by Hashimi et al. (1995). It will be seen that, granted sample quality remains problematic, all the pairs apart from that for ~10,400 yr BP indicate a significant northward dip which echoes the Holocene trend. That the difference has a major component resulting from vertical movements of the land rather than differential compaction or subsidence offshore is implied by widespread tidal flats and mudflat deposits north of 18°N compared with emergent stacks and sea caves to the south (Manjunatha and Shankar, 1992).

Table 2
Tidal records at 5 stations expressed as change in mm/yr. (a) after Emery and Aubrey (1989), (b) after Bendick and Bilham (1999)

	Lat (N)	Period *a*		Period *b*	
Mumbai	18°56′	1878–1982	−0.9	1878–1990	−0.19
Mangalore	12°53′	1953–1976	1.3	1953–1990	3.85
Cochin	9°55′	1939–1982	−2.1	1939–1990	−1.29
Madras	13°06′	1916–1982	−0.4	1916–1990	1.15
Vishakhapatnam	17°42′	1937–1982	−0.7	n.d.	

Table 3
Holocene shoreline ^{14}C dates for Saurashtra after Gupta (1977) and Agrawal and Guzder (1972)

Location	Lat. (N)	Elev.[a]	Material[b]	^{14}C age (yr BP)	Calib (yr bp)	Sample no
Porbandar	21°39′	6.2	Shell	6100±280	6510±690	TF-1058
		6.4	Shell	6300±250	6740±575	TF-1059
		6.5	Shell	6550±225	7046±600	5/1972
		5.5	Shell	6445±180	6920±425	8/1972
		5.0	Shell	6325±230	6770±535	6/1972
		4.2	Shell	6005±200	6405±470	9/1972
		3.5	Shell	6270±200	6715±236	7/1972
		3.0	Shell	5985±210	6395±480	10/1972
Rahalmata	20°45′	2.8	Shell	6320±270	6755±640	TF-1045

Calibrated for this study after program by Stuiver and Reimer, 1993 (version 4.1.2); ages rounded to nearest 5 yr; error given is larger of two limits at 2 s.d.
[a] Above high tide level.
[b] Species not given; calcite <1% XRD); isotopic normalisation not possible in the absence of ^{13}C values.

Table 4
Elevation differences between Ratnagiri and Mumbai for ~2800 and ~2200 yr BP

Age (yr BP)	Elevation (m)	
	Mumbai (18° 58′N)	Ratnagiri (17° 04′N)
2800±110	1.55	5.9
2100–2300	0.55	6.00

Data from Agrawal and Guzder (1972).

Table 5
^{14}C-dated pairs off west coast of India

Latitude	Material	Age (yr BP)	Depth (m)
15°15′	Limestone	11,040±135	95
19°05′	Algal pellet lst	11,150±130	150
17°00′	Algal bryozoan lst	10,415±250	180
20°24′	Ooid concentrate	10,400±300	85
14°25′	Carbonised wood	9630±120	32
20°10′	Sediment	9830±180	73
14°43′	Peat	8910±160	29
19°30′	Oolitic lst	8960±200	82
14°40′	Carbonised wood	8620±300	26.8
16°40′	Limestone	8395±145	68

Data from Hashimi et al. (1995).

The offshore, palaeoshore and tidal data on the west coast thus concur in indicating a synform at ~19°N, the latitude of Mumbai, between a zone of uplift at about 21°N and the antiform at ~13°N previously identified onshore.

4. Lithospheric buckling

Analysis of the Mangalore–Madras belt led Subrahmanya (1996) to suggest that it represented active deformation of the Indian plate in response to the regional stress field. The buckling reported by Bendick and Bilham (1999) on the SW coast of India had a wavelength of about 150–200 km. Vita-Finzi (2002) suggested that the larger earthquakes of peninsular India could be accommodated by five such zones of buckling between latitudes 10° and 25°N and aligned roughly SSW–ENE. Bilham et al. (2003) subsequently proposed that India's collision with Tibet had resulted in flexure with a wavelength of ~670 km, and argued that compressional stresses within a trough south of the bulge could account for the Bhuj, Latur and Koyna reverse earthquakes.

The five belts of buckling of Vita-Finzi (2002), slightly modified in the light of additional data, are shown in Fig. 3. Moving from south to north, the first (I) corresponds broadly with the Palghat Gap and may include the zone of uplift east of Cochin recognised by

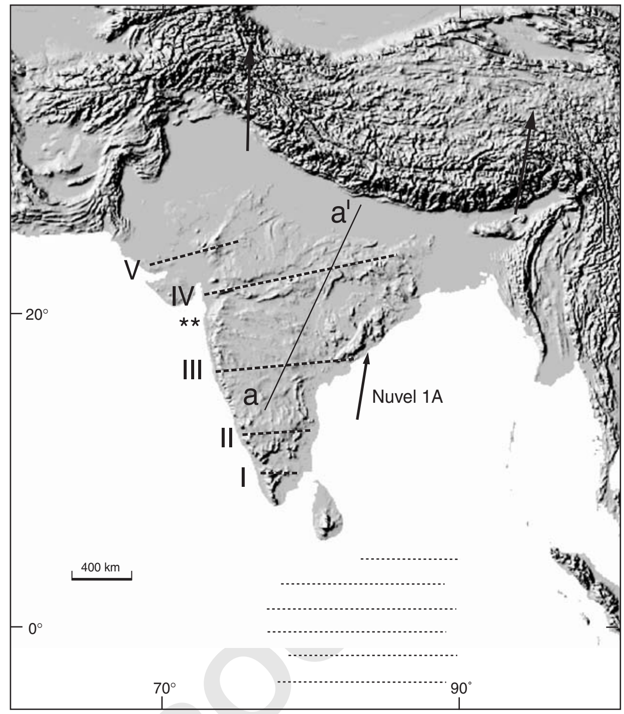

Fig. 3. Proposed belts of deformation corresponding to lithospheric buckles. (I) Based on geomorphology after Ramasamy (1989); (II) geomorphology and gravity data mainly after Subrahmanya (1996); (IV) extension of Narmada–Son rift; (III and V): belts proposed in this paper. Asterisks mark zone of Holocene subsidence indicated by palaeoshorelines. Line a–a′: geodetic contraction at 3±2 mm/yr in 1990–2000; bold arrows indicate approximate plate convergence vectors (after Bilham and Gaur, 2000). Dotted lines mark gravity lineations after Stein et al. (1990).

Ramasamy (1989), the second (II) to the Mangalore lineament of Subrahmanya (1996), the third (III) to the Koyna/Killari group of earthquakes, the fourth (IV) to the ENE–WSW Narmada–Son rift and the Saurashtra peninsula, and the fifth (V) to a Kutch group.

The spacing between the proposed buckles varies along strike and range approximately from 400 to 800 km. Relatively shallow compressional structures predominate, and there is geomorphological evidence for Holocene folding along lineaments I and II and more general uplift in the western part of lineament IV. The counterclockwise northward change in strike azimuth shown in Fig. 3 is consistent with the observation by Bilham and Gaur (2000) that, setting aside the complexities of plate interaction and the issue of rates, the plate rotation of the Nuvel-1A model (DeMets et al., 1994) predicts that convergence between India and Eurasia on an azimuth of 022° may be manifested as convergence at 017° in NE India and at 004° in the NW. The form of the Himalayan boundary doubtless also favours focusing of the stress field towards the NW.

In their discussion of buckling in peninsular India Subrahmanya (1996) and Bendick and Bilham (1999) referred to evidence for deformation of the floor of the Central Indian Ocean. Gravity and geoid data for the Indo-Australian plate derived from the Geos 3 and Seasat satellites show that the Central Indian Basin is characterised by a series of undulations which are oriented 040° and have a wavelength of $\sim$200 km (Cazenave et al., 1987). They have been detected on lithosphere ranging in age from $\sim$10 Ma to 35–40 Ma. At least 17 earthquakes of $M \geqslant 6$ were recorded in the basin during the 20th century (Deplus, 2001). A finite element study by Coblentz et al. (1998) has shown that the major contributor to the first-order intraplate stress field is ridge push, although variations in its magnitude and orientation owe a good deal to the stress focusing mentioned earlier.

Haxby and Weissel (1986) also favoured intraplate compression to explain the undulations in the central Indian Ocean east of the Southeast India Ridge even though they had proposed a convective origin for analogous features in the south Pacific. Louden (1995) has since demonstrated that for at least one of the ridges the deformation is by buckling. The problem then arises of how to overcome the large elastic thickness of the lithosphere by realistic end loading. One solution is to substitute an elastic–plastic model for a purely elastic configuration. McAdoo and Sandwell (1985) likewise showed that, using an elastic–plastic model, oceanic lithosphere with an age similar to that of the NE Indian Ocean (40–70 Ma) had a net compressive strength of about 12% of the elastic buckling stress.

Estimates for the elastic thickness of the Indian plate vary widely. Gravity measurements combined with the inferred flexure of the plate where it descends below the Indo-Gangetic plain point to a value in the range 70–120 km, but the free-air gravity field indicates a thickness of about 37 km (Bilham, pers. comm.), a figure which tallies with the apparent lack of earthquakes deeper than $\sim$40 km noted here. In the Bay of Bengal the maximum depth of earthquake nucleation is 36 km (Biswas and Majumdar, 1997), and here too seismic profiling and marine geodesy indicate undulations in acoustic basement with wavelengths of about 200 km (Biswas and Majumdar, 1997). For the period 1963–1992 the diffuse seismicity between the Indian peninsula and the Andaman-Nicobar subduction zone included 11 events of m_b 5.0–5.8, and the focal plane solutions for nine of them indicate pure thrusting. As the majority nucleated at depths of 20–36 km, Biswas and Majumdar (1997) suggest that brittle deformation occurs in the basement and folding in the upper sedimentary layers.

A lithosphere which was overlain by weak sediments at the start of intraplate deformation may have a compressive strength three times lower than it would in the absence of a sediment cover (Zuber, 1987). Indeed, it can be shown that the force responsible for lithospheric folding in the Central Indian Basin, where the lows are filled with sediment, is an order of magnitude smaller than where there are no sediments (Martinod and Molnar, 1995).

The resulting pattern may be distorted by pre-existing lithospheric deformation, notably that resulting from seamount loading (Karner and Weissel, 1990). In addition, the folds die out in the southern part of the Central Indian Basin, which has been explained either by the absence of sediments there or by a hypothetical southward reduction in the applied force. Such a reduction is compatible with the 'gravity glide' model of plate movement (Price et al., 1988), which predicts an across-strike gradient (in this instance a southward reduction) in the age and amplitude of any resulting buckles.

How far is the buckling displayed by the Central Indian Ocean pertinent to events further north? Whereas the eastern and western boundaries of the Indian plate between 10°N and 10°S are clearly defined respectively by a ridge and by a trench, its southern limits—if one accepts the validity of a Capricorn plate—are set by an area of diffuse seismicity and nondescript bathymetry. The likelihood is that a clear division between marine and continental crust cannot at present be made. Indeed, a measure of rheological continuity is apparent. Using elastic thickness as the touchstone, Manglik and Singh (1992) have estimated the thickness of the Indian shield to range between 65 and 79 km for a strain rate of $10^{-14}\,\mathrm{s}^{-1}$, with a corresponding strength value for the lithosphere of about $10^{13}\,\mathrm{N\,m}^{-1}$. This is close to the $2.56 \times 10^{13}\,\mathrm{N\,m}^{-1}$ thought to be required to initiate buckling in the central Indian Ocean and the $4 \times 10^{12}\,\mathrm{N\,m}^{-1}$–$10^{13}\,\mathrm{N\,m}^{-1}$ needed to sustain the Tibetan plateau (Gerbaud, 2000). Secondly, modelling on the basis of an elastic thickness of 35 km yields buckles with a wavelength of 150–200 km after 7 Ma (Gerbaud, 2000). In other words, different elastic thicknesses can be reconciled by a proportional change in wavelength.

5. Discussion

Mention of spacing raises the issue of the reactivation of normal faults in the basement as an alternative means by which seismogenic reverse faulting can develop transversely to the shortening direction. Existing structures will doubtless be exploited, as with the Narmada–Son graben. It is possible that they will already display some regularity in their spacing if the extension they represent was subject to the strain shadow effect (Bazant and Cedolin, 1991; Harris, 2000), which postulates that a new fracture will relieve the horizontal surface stress close to it to create a zone where further crack initiation is inhibited. But not all the larger peninsular earthquakes are associated with rifts (Rajendran, 2000), and the uniform regional extension required for pervasive normal faulting is difficult to visualise in an area such as the Indian shield which has been subject to long-term compression.

An economical alternative is to reverse the usual order of events by making some of the reverse faulting the consequence rather than the cause of fold localisation. Models are reported in which buckling is manifested after 1% of shortening and faults stabilise the position of the inflexion points after 5% (Gerbaud, 2000; see also Lambeck, 1983). A stress shadow effect will then operate under compression as well as under tension and help to regularise buckle spacing. There are some parallels with the scheme proposed by Montési and Zuber (2003) for the Central Indian Basin in which the long (~200 km) wavelength undulations in the basement result from the interaction between buckling and shear zones whose spacing is governed by what they term 'localisation instability'. A better analogy is perhaps the process by which the development of regularly spaced wrinkle ridges on a variety of planetary surfaces is followed by reverse faulting (Watters, 1991).

The trivial rate of present-day shortening of the peninsula revealed by geodetic surveys of India, which amounts to 3 ± 2 mm/yr on an azimuth of NNE–SSW for 1990–2000 (Bilham and Gaur, 2000), would seem inconsistent with buckling and related coseismic faulting, especially when we note from Table 1 that substantial events occurred in different parts of the peninsula in the same year. Two plausible explanations are that the decade of measurement was geodetically quiet and that the century of seismic record was unusually active. But there is a third possibility that some shortening leads to thickening of the ductile layer—though more pervasively than by bondinage—and is manifested fully at the surface only by coseismic deformation which is discontinuous in its distribution as well as its timing.

For these suggestions to be tested the seismic evidence must include sufficient well-instrumented events to establish any relationship between earthquake mechanism and fold geometry. There is also room for seeking analogous structures in other parts of the Indo-Australian composite plate: in Australia, as in India, many earthquakes are not associated with known major faults (Denham et al., 1979) and the in situ stress field strongly reflects the influence of plate boundary forces (Hillis and Reynolds, 2000).

Such tests are justified by the potential benefits of the buckle model for the mitigation of seismic hazard. For, rather than unrealistically claim to predict when or precisely where earthquakes will occur, it identifies a small number of relatively narrow belts within which earthquake-resistant structures and the advance provision of emergency relief can be concentrated.

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