Bulletin of the Seismological Society of America, Vol. 88, No. 1, pp. 160–169, February 1998

Seismicity and Stresses in the Brazilian Passive Margin

by Marcelo Assumpção

Abstract Seismicity and stresses along the Brazilian continental margin show different patterns between the northeastern and southeastern regions. In the northeastern margin, earthquakes tend to occur onshore under a strike-slip regime with horizontal compression parallel to the northern coast line. In the southeastern margin, higher seismicity is observed offshore, in areas where the continental crust was highly extended during the South Atlantic rifting in the Mesozoic. Three new events in the southeast continental shelf are shown to have reverse-faulting mechanisms, from waveform modeling of short-period teleseismic *P* waves. Multiple depth phases (reflections from the top and bottom of the water layer and double reflections in the water layer) could be identified and better constrained the hypocenters to middle and upper crustal depths. In South America, models of intraplate stresses caused by plate boundary forces and spreading effects due to the continental/oceanic crustal transition indicate higher compressional stresses in the SE offshore area, as compared with the continental area, in agreement with the observed higher seismicity and reverse-faulting mechanisms. The combination of regional stresses, local flexural effects from thick sedimentary loads, and a presumably weaker crust from Mesozoic thinning explains the main patterns of seismicity in the northeastern and southeastern Brazilian margins.

Introduction

Passive margin earthquakes are an important aspect of intraplate seismicity, not only because continental shelf and coastal earthquakes account for one-third of all seismicity in stable continental crust (Johnston, 1989) but also because of the increasing seismic risk in many oil-rich continental shelves.

Since Sykes (1978) suggested a relationship between intraplate earthquakes and crustal "zones of weakness," much progress has been made in understanding the seismicity of stable plate interiors. Johnston (1989) and Johnston and Kanter (1990) showed that 70% of all large intraplate earthquakes (magnitudes $>6\ M_s$) occurred in extended (and presumably weak) crust such as passive margins and Mesozoic rifts. Most intraplate earthquakes can probably be explained by rupture in pre-existing (hence weak) faults, such as demonstrated by Zoback (1992b) for North America by analyzing the focal mechanisms in relation to the crustal stresses.

Stein *et al.* (1989) emphasize that most mechanisms proposed to explain passive margin seismicity (stresses from continent/ocean density contrasts; flexure due to sediment loading; ridge-push stresses) should produce seismicity in all passive margins, but this does not seem to be the case. Some passive margins seem to be more active than others, specially those undergoing rebound from the unloading of the last ice age, but the evidence was not conclusive due to the lack of a more comprehensive survey and more complete catalogs (Stein *et al.*, 1989).

Different seismicity and stress patterns may be found for different passive margins. In the Atlantic margin of North America, compressional stresses tend to be oriented ENE–WSW, mostly parallel to the northeastern coastline, and are interpreted as due mainly to ridge-push forces (e.g., Seeber and Armbruster, 1988; Zoback, 1992a), although some local variations in stress directions are observed; offshore seismicity occurs mainly in the northeastern continental shelf, but not in the southeastern shelf. In northern Europe, despite some scatter in the stress observations in the continental shelf, a regional SH_{max} oriented NW–SE, roughly perpendicular to the Norwegian coast line, can be seen (Müller *et al.*, 1992; Gregersen, 1992) that has been interpreted as the result of both ridge-push forces from the North Atlantic and collisional forces with the African plate to the south. In the Indian subcontinent, seismicity seems to predominate onshore; the SH_{max} stress data near the margin is quite often oblique to the coastline, but a general NE–SW to N–S direction seems to be observed, most probably related to the Hymalaian collision to the north (Gowd *et al.*, 1992; Zoback, 1992a). Suleiman *et al.* (1993) showed that earthquakes in the Atlantic margin of west Africa tend to have strike-slip mechanisms, differently from the predominantly reverse faulting in the continental shelves of northeastern North

America and southeastern South America. No clear pattern for the SH_{max} orientation was found by Suleiman *et al.* (1993) for the Atlantic margin of Africa, with significant variations across the region.

Here we discuss the epicentral distribution and stresses along the Brazilian Atlantic coast to contribute to a better understanding of the seismicity of passive margins. The low seismicity rate in Brazil (as compared with other intraplate areas such as North America and Australia) generally precludes reliable statistical correlations between epicenters and geological features. However, the large extent of the Brazilian Atlantic coast makes the data presented in this article important for other global studies of passive margin seismicity. It is suggested that two main sources of stress (regional and local components) combine to produce the observed seismicity pattern. The regional component is a compressive stress oriented roughly E–W to ESE–WNW due to plate-wide forces such as ridge-push and asthenospheric drag. The local component is due to two processes: density contrast between continental/oceanic crusts ("spreading stresses") and lithospheric flexure due to sediment load in the continental shelf. In the upper crust, both spreading and flexural stresses produce extension onshore and compression offshore.

Epicentral Distribution

Figure 1a shows all epicenters of the Brazilian catalog down to magnitude 3.0 m_b. This catalog was based on the compilation of Berrocal *et al.* (1984) and the *Brazilian Seismic Bulletins* (1984 to 1996), published by the *Brazilian Journal of Geophysics* (*Revista Bras. de Geofisica*). Historical compilations and seismic bulletins tend to be biased: more events are located in areas of higher population density (such as along the Brazilian coast) and in regions with a better coverage of local seismic stations (such as southeastern and northeastern Brazil). To compare the seismic activity of the continental margin with the mid-continental region, a dataset with uniform coverage was extracted from the Brazilian catalog. For this purpose, magnitude thresholds depending on the year of occurrence (Table 1) were used to select events from the "whole" catalog (Fig. 1a) and form a "uniform dataset" (Fig. 1b).

Since about 1950, the ISS and ISC international catalogs should be complete for central and eastern South America for magnitudes above 6.0 m_b. The installation of the WWSS and the Canadian networks in the early 1960s allowed events down to about 5.0 m_b to be detected by the ISC catalog. In the mid 1960s, the installation of a WWSS station in Natal (NE Brazil) and a highly sensitive array in central Brazil (SAAS; Berrocal, 1974) improved the detectability of the ISC catalogs for Brazilian events; also, Brazilian events started to be located by Brazilian institutions (Berrocal, 1974; Berrocal *et al.*, 1984). In the late 1970s, many seismographic stations were installed in Brazil (mainly to monitor reservoir-induced seismicity), and regional events down to 3.0 m_b (not included in the ISC bulletins) started to be located on a routine basis by the joint efforts of the universities of Brasilia, São Paulo, and Rio Grande do Norte (UnB, USP, and UFRN, respectively). The thresholds in Table 1 were based partly on expert opinion (experience of university staff with earthquake locations in Brazil) and partly on frequency-magnitude plots (Berrocal *et al.*, 1996). In SE Brazil, for example, where station density is higher, events down to magnitude 3.2 are thought to be complete since 1980 (Berrocal *et al.*, 1996; Assumpção *et al.*, 1997). Selecting earthquakes with the thresholds of Table 1 produces 32 events above magnitude 4.5 in the period 1968 to 1996, that is, about 1 event/year; and 146 events above 3.5 from 1980 to 1996, or about 9 events/year. This shows that the thresholds in Table 1 are self-consistent, and large errors are not expected.

Figure 1b shows the earthquakes of this uniform dataset, selected from the Brazilian catalog with the thresholds of Table 1. Clearly, more stringent threshold criteria could be used, but the number of selected events would be too small to allow meaningful seismotectonic interpretations. The "uniform catalog" (Fig. 1b) should be complete enough to allow comparisons of the seismic activity between different regions. Some interesting patterns are apparent in Figure 1b:

1. The Brazilian passive margins (continental shelf + coastal areas) do not seem to be significantly more active than the average continental interior.
2. In the continental region, earthquakes tend to occur in areas of low topography. The plateau areas in Eastern Brazil (altitudes higher than 600 m) seem to be less active than the rest of the continental mid-plate areas: The uniform catalog has no events larger than magnitude 4.2 in the plateau area (Fig. 1b), and they are fewer in number there.
3. In the northern and northeastern margins (north of 10° S), earthquakes tend to occur in the continent with almost no activity offshore. South of 15° S the activity tends to be concentrated in the continental shelf, with lower levels of activity onshore.

Chang *et al.* (1992) mapped the limit of the continental crust that was extended during rifting of the South Atlantic (Fig. 1b, dashed line). It is remarkable that this limit of extended crust also seems to be the limit of the seismicity in the SE continental margin. This is consistent with the interpretation of Mesozoic extended crust being a "zone of weakness" (e.g., Sykes, 1978; Johnston, 1989). Additionally, Figures 1a and 1b seem to show a concentration of activity along the continental slope, i.e., between the 200- and 2000-m bathymetry, approximately along the axis of maximum sedimentary thickness (Fig. 5).

The estimated epicentral errors for offshore events are small enough to ensure the above conclusions. In the southeastern margin, the majority of the offshore events, in the magnitude range 3.5 to 4.0, have been located by stations in

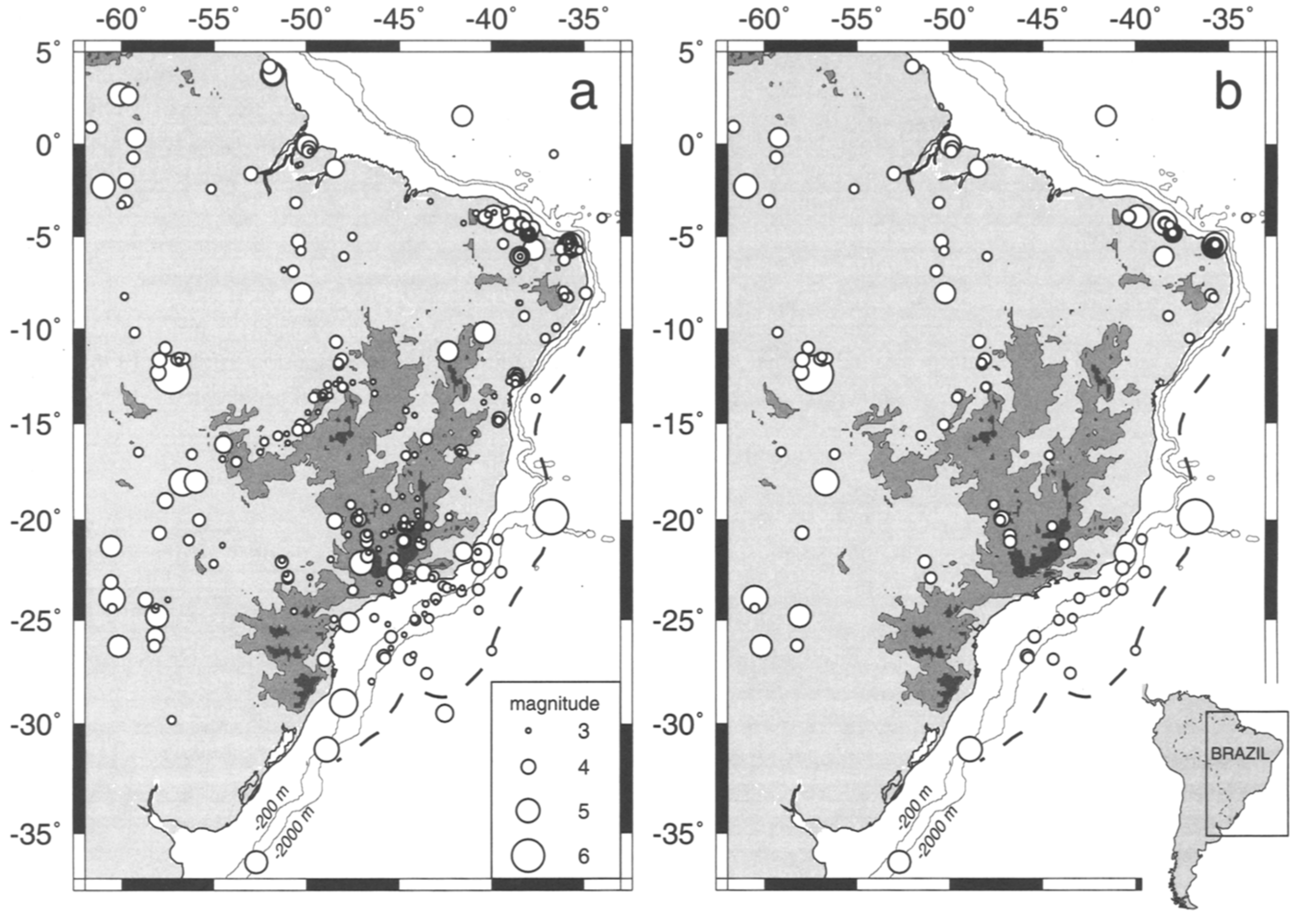

Figure 1. Epicentral distribution in eastern Brazil. (a) "Whole catalog" including all historical and instrumental data. (b) "Uniform catalog" is a selection from the whole dataset using the thresholds of Table 1 to make the coverage geographically uniform. Darker areas in the continent are topographies higher than 600 m (gray) and 1000 m (black). Lines offshore are the 200- and 2000-m bathimetry. The dashed line offshore is the limit of extended continental crust (Chang *et al.*, 1992).

the distance range 400 to 1500 km, with 10 to 20 readings (*P* and *S*). The standard errors in the epicenters range from about 10 to 50 km. The events have been located using a velocity model appropriate for Brazil (Kwitko and Assumpção, 1990) with upper mantle velocities higher than Herrin's model. Using other velocity models, such as Herrin's, does not shift the epicenters by more than ± 30 km. So, the overall epicenter accuracy is probably better than 100 km for most of the events in Figure 1b, which is roughly about the "width" of the continental slope (using the 200- and 2000-m bathimetry as a measure of that width in the southeastern margin). Because continental shelf earthquakes are often well recorded by some stations located more than 1000-km inland, events in the oceanic crust outside the ocean–continent boundary (dashed line in Fig. 1) should also be detected by stations near the southeastern coast. Finally, the four largest offshore events in Figure 1b ($m_b > 4.7$), located teleseismically, are all within the limits of extended crust. So, despite the small number of earthquakes in the "uniform" catalog, there is strong evidence for higher seismicity in the extended crust beneath the southeastern continental shelf, as compared both with the onshore continental margin and the old oceanic crust.

Table 1
Magnitude Thresholds Used in Figure 1b

Year	Magnitude	Comments
1950	6.0	International Seismological Summary catalog
1962	5.0	WWSS world network
1968	4.5	start of Brasília array (SAAS) and NAT station
1980	3.5	regional networks of UnB, USP, and UFRN

The opposite pattern of seismicity (onshore/offshore) between northern and eastern margins may be related to the different processes of continental rifting. In the South Atlantic rifting, E–W extension predominated causing large extensional deformation (and subsidence) of the original continental crust with stretching factor $\beta > 3$ for hundreds of kilometers (Chang *et al.*, 1992). In the Equatorial Atlantic, the rifting process was characterized by predominantly transcurrent motion associated with long fracture zones of the central mid-Atlantic ridge. Possibly, little extended crust remains in the continental shelf of northern Brazil.

Focal Mechanisms and Stresses in the Continental Margin

Earthquake focal mechanisms along the continental margin (Fig. 2) show that the southeast (offshore) and northeast (onshore) seismic areas have different stress regimes: thrust faulting characterizes the activity of the southeast continental shelf, whereas strike-slip stresses predominate in the northeast onshore coastal area. The available focal mechanisms in the south are few and far apart, as compared to the data in the northeast. However, although local variations in the stress regime are possible, the consistent nature of the five mechanisms in the south indicate that compressional stresses should predominate along the southeastern continental shelf.

Focal Mechanisms in the Northeastern Continental Margin (North of 10° S)

In northeastern Brazil, the earthquakes tend to occur around the onshore border of the Potiguar Mesozoic marginal basin with strike-slip focal mechanisms (Fig. 2) at upper crustal depths (Assumpção, 1992; Ferreira *et al.,* 1995, 1998). Except for one strike-slip mechanism, determined by regional polarities and teleseismic *P*-wave modeling (Assumpção *et al.,* 1985), all other focal mechanism solutions were determined with *P*-wave polarities at local stations in detailed studies of earthquake swarms or aftershocks. In 7 of the 10 local studies, the main fault plane could be identified by the hypocenter distribution and was consistent with the independently determined nodal plane solution (Ferreira *et al.,* 1987, 1995, 1998). In Figure 2, the larger beach balls in the northeast denote focal mechanisms with identified fault planes, and smaller beach balls denote less reliable solutions.

Average SH_{max} directions (represented by white bars in Fig. 2) were determined at several clusters of three or more points, within 100 to 150 km, including breakout measurements (Lima *et al.,* 1997) or focal mechanisms (Ferreira *et al.,* 1998). Both focal mechanism and breakout data show that the stress regime in the continental border of northeastern Brazil is strike slip with compression parallel to the coastline and extension perpendicular to it.

Finite-element modeling of intraplate stresses in the South American plate (Meijer, 1995; Coblentz and Richardson, 1996), caused mainly by ridge-push and plate contact resistance, produces compressive stresses oriented roughly WNW–ESE to E–W in northeastern Brazil. When the effect of density contrasts between continental/oceanic crusts ("spreading stresses" of Bott and Dean, 1972) is included in the modelings, the resulting stress regime onshore becomes strike slip and better fits the observed earthquake fo-

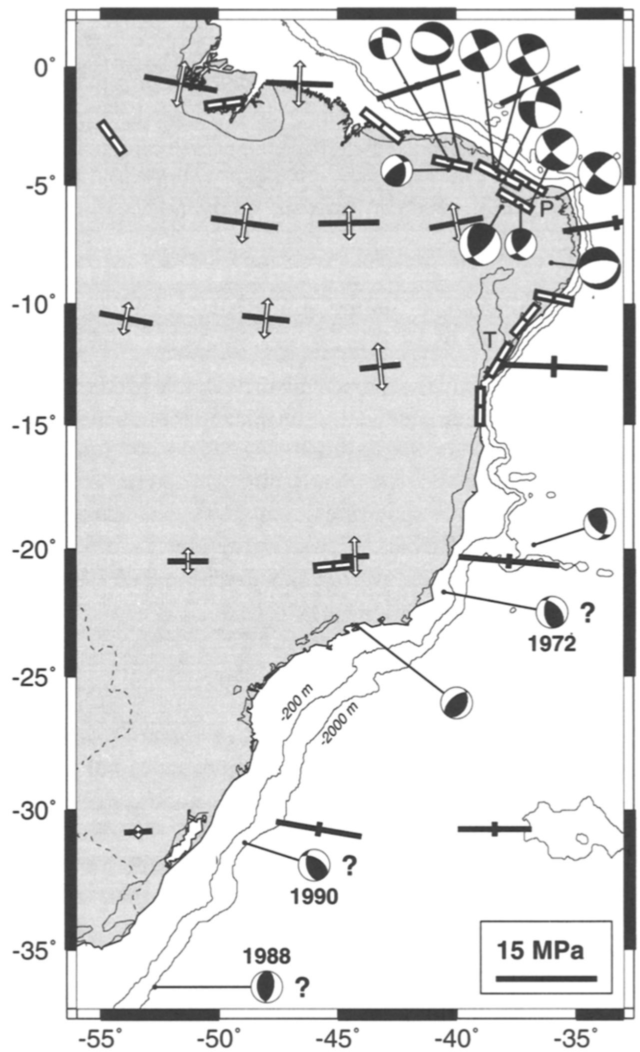

Figure 2. Intraplate stresses and focal mechanisms along the continental margin (lower hemisphere; compressional quadrants shaded). Larger beach balls are more reliable solutions with identified fault plane. Solid bars and open arrows are compressional and extensional principal horizontal stresses, respectively, modeled by Coblentz and Richardson (1996) for a 100-km-thick elastic plate subject to plate boundary forces and continental/ocean spreading forces. Short open bars indicate the observed average SH_{max} orientations (for clusters with three or more focal mechanisms or breakout measurements) from Lima *et al.* (1997). The "?" beside the nodal plane solutions 1972, 1988, and 1990 indicate that the strikes of the nodal planes are uncertain, although the nature of the mechanism (reverse faulting) is reliable. Note the predominance of strike-slip faulting in NE onshore and reverse faulting in SE offshore. Thin lines offshore are the 200- and 2000-m bathimetry; thin lines in the continent denote Mesozoic marginal basins such as Tucano (T) and Potiguar (P); dashed line onshore is Brazil political border.

cal mechanisms (Meijer, 1995; Coblentz and Richardson, 1996). Thus, for the onshore region of northeastern Brazil, both observations and theoretical modelings confirm the superposition of a regional compression, oriented roughly E–W (related to ridge-push and other plate contact forces), with local extensions oriented perpendicular to the coastline (related to the "spreading stresses" and also to flexural bending, as discussed below).

Focal Mechanisms in the Southeastern Continental Shelf (South of 15° S)

So far, the only focal mechanism available for the southeastern continental shelf is a fault-plane solution for a 6.1 m_b earthquake that occurred in 1955 at about 20° S near the Victoria–Trindade Island chain (Fig. 2), determined with teleseismic *P*-wave polarities and *S*-wave polarizations (Mendiguren and Richter, 1978). A composite reverse-faulting mechanism for a swarm of microearthquakes that occurred right at the coast (Berrocal *et al.,* 1993) is also included in Figure 2 (mechanism near 23° S).

Here we discuss the focal mechanisms of three additional events (Table 2, Fig. 3). The mechanisms are all clearly reverse faulting, although the orientations of the *P* axes cannot be constrained. The focal mechanism diagrams shown in Figure 3 represent one possible solution and were plotted in Figure 2 (beside a "?" mark) to help emphasize the different nature of the faulting regimes in the southeastern continental shelf as compared with the northeastern onshore margin.

Because no *P*-wave first motions could be clearly identified at regional distances, the focal mechanisms were calculated by modeling short-period *P*-wave seismograms at teleseismic distances. Two methods were tried: matching the relative amplitudes of the depth phases *pP* and *sP* by a grid search method (Assumpção and Suárez, 1988), and waveform inversion of the *P*-wave train (Nabelek, 1984). No teleseismic *S* waves could be identified to help constrain the mechanisms as all three events had M_s magnitudes lower than 4.5.

In all three events, the *pP* phases are very conspicuous in all stations, having amplitudes comparable to the direct *P* wave, but with opposite polarity. These characteristics, together with the clear compressional first motion at teleseismic stations, ensures that the events had reverse-faulting mechanisms. The nodal planes had dips in the range 30° to 60° but had unconstrained strikes. Although the direction of the *P* axes could not be determined, the teleseismic *P*-wave modeling allowed the hypocentral depths to be well determined, which is important for seismotectonic interpretations. Table 2 shows the hypocentral data for these three earthquakes and the layer thicknesses in the upper crust used for the *P*-wave modeling. Examples of a possible mechanism for each event, with comparison of synthetic and observed seismograms, are shown in Figure 3 and are discussed as follows.

Rio de Janeiro, 24 October 1972. This event occurred in shallow waters of the continental shelf of the Campos sedimentary basin. Mendiguren and Richter (1978) had already shown the reverse-faulting nature of this earthquake based on *P*-wave polarities. Crustal structure in the epicentral area is fairly well known from deep seismic reflection and gravity modeling (Mohriak and Dewey, 1987). Sedimentary layers with a total thickness of 6.4 km overly an extended or thinned crust with Moho depth near 20 km (Mohriak and Dewey, 1987). The hypocentral depth of the 1972 event, well constrained by the clear *pP* phases (Fig. 3a) and known crustal velocities, was found to be between 8 and 9 km, that is, about 3 km below the sediment/upper crust interface.

Uruguay, 26 June 1988. This event occurred in the continental slope, close to the 2000-m bathymetry, in an area with about 5 km of sediments. No detailed crustal model is available for this region, but information on sedimentary layers and upper crustal velocities from nearby refraction surveys (compiled by DNPM, 1984) was used to calculate the synthetic seismograms shown in Figure 3b. The large depth phase (*pP*) is the reflection from the water surface. The reflection from the sediments–water interface is a small precursor to the surface reflection (named *bP*), having small amplitudes due to the low-velocity contrast between water and unconsolidated sediments at the top of the sedimentary pack. The multiple reflection in the water layer (named *wwP*) can also be identified in some seismograms. Figure 4 shows

Table 2
Reverse-faulting earthquakes and hypocentral depths in southeastern continental shelf (events modeled in Fig. 2). Values of water depths and sediment thicknesses were those used in the waveform modeling. All depths and thicknesses are in kilometers.

Date	24 October 1972	26 June 1988	12 February 1990
Location	Rio de Janeiro	Uruguay	Rio Grande do Sul
Latitude, longitude (°)	−21.72, −40.53	−36.27, −52.73	−31.19, −48.92
m_b magnitude	4.8	5.2	5.5
Water depth	0.1	1.8	2.2
Sediment thickness	6.4	5.3	6.8
Hypocentral depth	8.5	17.7	12.8
Depth into basement	2.0	10.6	3.8

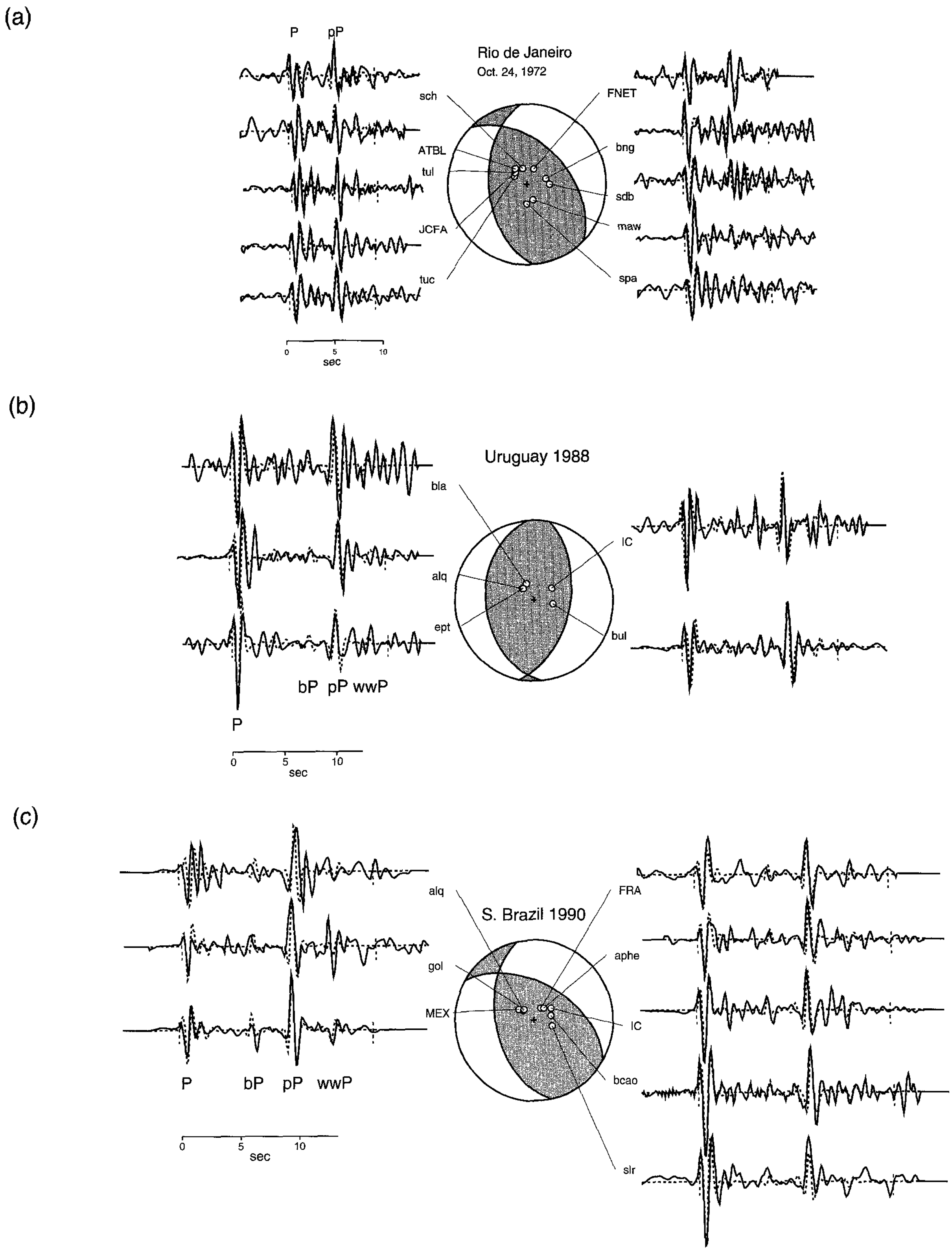

Figure 3. Examples of focal mechanism solutions for the three events in SE continental shelf (lower hemisphere; compressional quadrants shaded): (a) Rio de Janeiro 1972, (b) Uruguay 1988, and (c) southern Brazil 1990. Solid and dashed lines are the observed and calculated seismograms, respectively. Single-station seismograms are identified by lower-case letters; stacked seismograms, by capital letters. The strikes of the nodal planes are not well constrained, but the reverse-faulting nature of the mechanism is well determined by the relative amplitudes and polarities of *pP*. The other depth phases (*bP* and *wwP*) are defined in Figure 4.

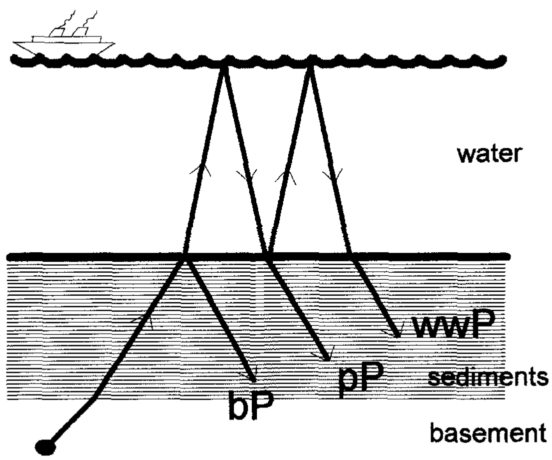

Figure 4. Diagram to illustrate names of the depth phases used in Figures 3b and 3c. All paths are *P*-wave type.

a diagram of the three depth phases. The 17.7-km hypocentral depth places this event at about 10 km below the base of the sediments. No detailed information on Moho depth is available, but a nearby seismic refraction profile (compiled by DNPM, 1984) and extrapolation of the interpretations of Chang *et al.* (1992) suggest that this event occurred in the middle to lower part of an extended continental crust.

Rio Grande do Sul (Southern Brazil), 12 February 1990. This event occurred in the continental slope of the Pelotas sedimentary basin. In the epicentral area, 2.3 km of water overly about 7 km of sediments. Figure 3c shows the conspicuous *pP* phase (reflection from the sea surface) together with the precursor *bP* (reflection from the water–sediment interface). In the seismograms, the first multiple reflection in the water layer (*wwP*, Fig. 4) can also be seen. The hypocentral depth was calculated at 13 km, that is, about 4 km below the sediment–upper crustal interface.

The alignment between observed and synthetic signals for the three depth phases is not perfect because of the large lateral variation in water and sediment thicknesses in the epicentral area. However, the main characteristics of the depth phases (amplitudes and polarities) are well reproduced. The opposite polarity of the multiple reflection *wwP*, as compared to *bP* and *pP*, is clearly seen in most seismograms. This confirms the interpretation of the depth phases and assures the reliability of the hypocentral depth.

Berrocal *et al.* (1996) suggested that the 1990 Rio Grande do Sul event could have been a landslide, considering that large mass movements have been observed in the continental slope in recent geological times. This hypothesis has important implications for seismic risk estimates along the continental shelf. Large slumps in the continental slope can be associated with large earthquakes, such as the magnitude 7.2 Grand Banks event of 1929 off the eastern coast of Canada. The Grand Banks seismic event has been interpreted as the slump itself (Hasegawa and Kanamori, 1987; Hasegawa and Herrmann, 1989) or as a deep crustal complex double-couple earthquake causing the sediment slumping (Bent, 1995). The 1990 Rio Grande do Sul event, in southern Brazil, clearly is not a slump: Both its reverse-faulting nature and its well constrained depth of 4 km into the upper crustal basement rule out the slump hypothesis.

Discussion

Several mechanisms have been proposed to explain passive margin seismicity such as ridge-push compression, density contrasts between continental and oceanic crusts ("spreading stresses"), and flexural stresses due to sediment loading (see review by Stein *et al.*, 1989). Any single mechanism cannot explain all features of passive margin seismicity, as pointed out by Stein *et al.* (1989). However, it is proposed here that a combination of those various sources of stress together with geometrical constraints can explain most features of the seismicity observed thus far in the Brazilian margin.

Ridge push has long been considered an important source of intraplate stress (e.g., Mendiguren and Richter, 1978; Stein *et al.*, 1989; Zoback *et al.*, 1992a) and is thought to play a significant role in passive margin seismicity. Stresses in the South American plate have been calculated by 2D finite elements (Stefanick and Jurdy, 1992; Meijer, 1995; Coblentz and Richardson, 1996) with ridge push modeled as a distributed body force in the oceanic part of the plate, balanced by asthenospheric drag or boundary forces from the neighboring plates. In the Brazilian continental margin, all models show about the same SH_{max} direction, roughly WNW–ESE to E–W. In Figure 2, we show the principal horizontal directions (SH_{max} and Sh_{min}) of Coblentz and Richardson's (1996) "model 3," calculated with ridge push balanced by collisional forces with the Nazca and other minor plates and a small positive basal shear stress. In this model, the effect of the "spreading stresses" in the continent to ocean transition has been included, which causes the stress regime in the continent to be strike slip, and it also makes the E–W compressional stresses larger (15 to 20 MPa) in the oceanic part and smaller (5 to 10 MPa) in the continental area, as shown in Figure 2. These horizontal stress magnitudes refer to a uniform 100-km-thick elastic plate.

In the southeastern Brazilian shield (about 20° S, 45° W, Fig. 2), away from the perturbing effects of flexural stresses at the continental margin, the observed stress regime obtained with inversion of four focal mechanisms (Assumpção, 1998) was strike slip with E–W SH_{max}, in very good agreement with the theoretical predictions of Coblentz and Richardson (1996), as seen in Figure 2.

In the southeast continental shelf, the predominance of reverse faulting is also consistent with the theoretical compressional stress regime predicted for the oceanic part (Fig. 2). In addition, the finite-element modelings indicate higher stress levels in the oceanic part as compared with the continental area, which may help explain the higher seismicity of the continental shelf (Fig. 1b).

The finite-element modelings, however, fail to predict the coast parallel SH_{max} near the Tucano marginal basin (about 12° S in Fig. 2). Lima *et al.* (1997) proposed that flexural stresses due to (1) locally uncompensated sediments of the onshore Tucano basin and (2) sediment load in the nearby offshore continental shelf could combine to produce extensional stresses perpendicular to the coast with magnitudes large enough to overcome the regional stresses. In fact, local flexural stresses are not included in the force models of Meijer (1995) or Coblentz and Richardson (1996) but can potentially produce local stresses higher than the regional stresses from plate-wide forces (Cloetingh *et al.,* 1984, Stein *et al.,* 1989; Zoback and Richardson, 1996).

Figure 5 shows the main sedimentary packs in the continental shelf. Only the 6- and 4-km total isopach is shown for each depocenter. Note that in the SE continental shelf, sedimentary layers tend to be very thick with the axis of the sedimentary load roughly along the continental slope, coincident with the epicentral trend. In the NE continental shelf (north of 10° S) sedimentary layers are much thinner, and no extensive basins deeper than 4 km are found. Only in the northern continental shelf, in the Amazon Fan (around 3° N, Fig. 5), are deep sediments present again. The current flexural stresses should not be caused by the total sedimentary thickness, as probably most of the load has already been released by inelastic deformation and faulting during the evolution of the margin. However, it is reasonable to expect that flexural effects should be more important in the SE continental shelf than in the NE margin because of the greater amount of sedimentary load present there.

Flexural bending of the lithosphere caused by sediment load in the continental shelf produce compressional stresses in the upper crust right under the load and extensional stresses about 100 to 200 km away from the load axis toward the peripheral bulge (e.g., Turcotte and Schubert, 1982; Cloetingh *et al.,* 1984). This additional flexural effect should enhance the compressional stresses in the SE continental shelf (compression under the load) and also enhance the strike-slip stresses in the onshore part of the NE margin (extension perpendicular to the coast toward the peripheral bulge). In the northeastern margin, hypocentral depths are usually less than 10 km (Ferreira *et al.,* 1997); the three earthquakes in the SE continental shelf analyzed here also have depths less than about 10 km into the upper crustal basement (Table 2). The shallow nature of the earthquakes would place them above the neutral plane of a bending plate, consistent with flexural compression under the load (SE continental shelf) or flexural extension toward the peripheral bulge (NE onshore margin).

Another interesting mechanism causing extension perpendicular to the coast line, proposed by Kemp and Stevenson (1996), is the thermal subsidence of old oceanic lithosphere which is flexurally resisted by the adjacent continent. This occurs because the lithosphere created in a new ocean basin, at a depth comparable to mid-ocean ridges, establishes mechanical continuity with the adjacent continent. Later subsidence of the oceanic lithosphere to abyssal depths will be flexurally inhibited by the continent causing tensile stresses perpendicular to the coast line. These tensile stresses can potentially reach magnitudes high enough to overcome the compressional stresses from ridge push (Kemp and Stevenson, 1996). This mechanism will be more important in areas where the continent–ocean crustal transition is sharp

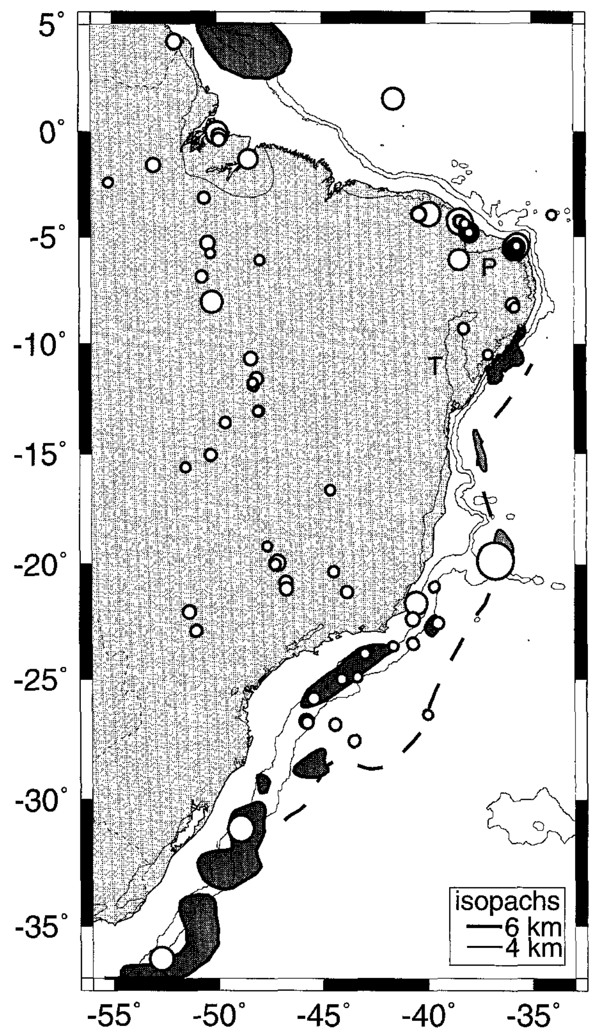

Figure 5. Location of the thickest sedimentary packs along the continental shelf indicated by solid-shaded areas. Only isopachs of 6 or 4 km total thickness are indicated for each depocenter. In the southeast continental shelf, thicknesses larger than 6 km are common, whereas in the northeast continental shelf, no extensive sedimentary packs thicker than 4 km are found. Further to the north, in the Amazon Fan (about 3° N), thick sedimentary layers can be found again. Isopach data from DNPM (1984) complemented with Ludwig *et al.* (1978) and Lucchesi (Petrobrás, written comm., 1991). Epicenters from the "uniform catalog" as in Figure 1b. Lines offshore as in Figure 1. In the continent, the thin lines indicate marginal basins, such as Potiguar (P) and Tucano (T).

rather than gradual, that is, in areas with a large difference between the isostatic depths of the continent and the ocean bottom. So, this coast-perpendicular flexural extensional stress, due to oceanic subsidence, should be more important in northeastern Brazil (say, north of about 15° S) and less important in the southeastern margin. In fact, extension perpendicular to the coast (coast parallel SH_{max}, Fig. 2) is clearly observed in most of the northeastern margin.

In the northeastern margin, earthquakes tend to occur along the WNW–ESE-trending coast, with little activity along the N–S-trending coast (Fig. 1). Flexural extension perpendicular to the coastline would enhance the strike-slip stresses along the WNW–ESE-trending coast. Along the N–S-trending margin, the local flexural extension has the same E–W orientation of the regional compression (SH_{max}). This would reduce the resultant differential stresses, which may explain the lower seismicity (Assumpção, 1992; Ferreira *et al.*, 1998). The geometry of the northeastern margin can therefore affect the superposition of local and regional stresses.

It is interesting to observe that in the northeastern margin, seismicity occurs preferentially in basement faults around the Potiguar Mesozoic rift basin (Ferreira *et al.*, 1998; see basin "P" in Fig. 5), similar to other Mesozoic rift basins in the northeastern margin of North America (Kafka and Miller, 1996). However, other Mesozoic rift basins such as the Tucano (near 10° S, Fig. 5) do not seem to correlate with any seismicity concentration. This sporadic correlation between seismicity and rift border faults is also seen in eastern North America (G. Bollinger, personal comm.).

Conclusions

The northeastern and southeastern Brazilian margins have different stress and seismicity patterns, which can be explained by the contribution of several factors:

1. *Onshore northeastern margin:* The strike-slip stresses in the upper crust result from the combination of roughly E–W compression (due to ridge-push and plate-margin forces) with coast-perpendicular extensional stresses (due to both "spreading stresses" and flexural bending). The geometry of the northeastern coast controls the resulting total stress field: Lower magnitudes of the differential stress are expected in areas of lower seismicity rate.
2. *Offshore southeastern continental shelf:* Large compressional stresses in the upper crust, probably oriented E–W to WNW–ESE, are the result of constructive superposition of several sources of stress: (a) ridge-push and other plate boundary forces, (b) compression due to lateral density contrasts between oceanic/continental crusts, and (c) compression from flexural bending beneath the load axis of the thick sedimentary pack in the continental shelf. In addition to the expected large stresses, the seismogenic crust had been severely extended by the rifting process of the South Atlantic and should be a zone of weakness more susceptible to failure.
3. *Southeastern continental plateau:* In the continental plateau (altitudes higher than 600 m) of southeastern Brazil (Fig. 1b), the lower seismicity could be related to low intraplate stress levels that arise from the spreading effect of the continents as modeled by Meijer (1995) and Coblentz and Richardson (1996, Fig. 2). These models only took into account the effect of the continent–ocean transition along the whole coast of South America. The spreading effect of the continental plateau itself, while of minor importance, is an additional effect contributing to the small magnitude of the resultant stress field.

Acknowledgments

This work was supported by Brazilian Grants CNPq 30.6227/79, CNPq 20.0756/91-5, and FAPESP 90/3741-5. Part of this work was carried out during a sabbatical visit to Lamont-Doherty Earth Observatory with computing support from Grant NCEER91-1021. I thank Won-Young Kim for help with computing problems and Cláudio Lima for discussions on intraplate neotectonics. Gil Bollinger and Allison Bent made valuable suggestions to improve the article.

References

Assumpção, M. (1992). The regional intraplate stress field in South America, *J. Geophys. Res.* **97,** 11889–11903.

Assumpção, M. (1998). Focal mechanisms of small events in SE Brazilian shield: a test of stress models of the South American plate, *Geophys. J. Int.,* in press.

Assumpção, M. and G. Suárez (1988). Source mechanisms of moderate-size earthquakes and stress orientation in mid-plate South America, *Geophys. J.* **92,** 253–267.

Assumpção, M., G. Suárez, and J. A. V. Veloso (1985). Fault plane solutions of intraplate earthquakes in Brazil: some constraints on the regional stress field, *Tectonophysics* **113,** 283–293.

Assumpção, M., J. R. Barbosa, J. Berrocal, A. M. Bassini, J. A. V. Veloso, V. I. Mârza, M. Huelsen, and L. C. Ribotta (1997). Seismicity patterns and focal mechanisms in southeastern Brazil, *Rev. Bras. Geofis.,* in press.

Bent, A. L. (1995). A complex double-couple source mechanism for the M_s 7.2 1929 Grand Banks earthquake, *Bull. Seism. Soc. Am.* **85,** 1003–1020.

Berrocal, J. (1974). South American seismotectonics from SAAS data. *Ph.D. Thesis,* Edinburgh University, UK, 240 pp.

Berrocal, J., M. Assumpção, R. Antezana, C. M. Dias Neto, R. Ortega, H. França, and J. Veloso (1984). *Sismicidade do Brasil,* Instituto Astronômico e Geofisico, Universidade de São Paulo, Brazil, 320 pp.

Berrocal, J., C. Fernandes, A. Bueno, N. Seixas, and A. Bassini (1993). Seismic activity in Monsuaba-RJ, Brazil, between December/88 and February/89, *Geophys. J. Int.* **113,** 73–82.

Berrocal, J., C. Fernandes, A. Bassini, and J. R. Barbosa (1996). Earthquake hazard assessment in southeastern Brazil, *Geofis. Int.* **35,** 257–272.

Bott, M. H. P. and D. S. Dean (1972). Stress systems at young continental margins, *Nature Phys. Sci.* **235,** 23–25.

Cloetingh, S. A. P. L., M. J. R. Wortel, and N. J. Vlaar (1984). Passive margin evolution, initiation of subduction and the Wilson cycle, *Tectonophysics* **109,** 147–163.

Chang, H. K., R. Kowsman, A. M. F. Figueiredo, and A. A. Bender (1992). Tectonics and stratigraphy of the east Brazil rift system: an overview, *Tectonophysics* **213,** 97–138.

Coblentz, D. D. and R. M. Richardson (1996). Analysis of the South American intraplate stress field, *J. Geophys. Res.* **101,** 8643–8657.

DNPM (1984). Mapa geológico do Brasil e da área oceânica adjacente, incluindo depósitos minerais. Escala 1:2.500.000. Departamento Nacional da Produção Mineral, Brasília.

Ferreira, J. M., M. Takeya, J. M. Costa, J. A. Moreira, M. Assumpção, J. A. V. Veloso, and R. G. Pearce (1987). A continuing intraplate earthquake sequence near João Câmara, Northeastern Brazil, preliminary results, *Geophys. Res. Lett.* **14,** 1042–1045.

Ferreira, J. M., R. T. Oliveira, M. Takeya, and M. Assumpção (1995). Superposition of local and regional stresses in NE Brazil: evidence from mechanisms around the Potiguar marginal basin, *Proc. IV Intern. Congr. Braz. Geophys. Soc.* **2,** 1025–1028.

Ferreira, J. M., R. T. Oliveira, M. Takeya, and M. Assumpção (1998). Superposition of local and regional stresses in NE Brazil: evidence from mechanisms around the Potiguar marginal basin, *Geophys. J. Int.,* in press.

Gowd, T. N., S. V. S. Rao, and V. K. Gaur (1992). Tectonic stress field in the Indian subcontinent, *J. Geophys. Res.* **97,** 11879–11888.

Gregersen, S. (1992). Crustal stress regime in Fennoscandia from focal mechanisms, *Geophys. Res.* **97,** 11821–11827.

Hasegawa, H. S. and R. B. Herrmann (1989). A comparison of the source mechanism of the 1975 Laurentian Channel earthquake and the tsunamigenic 1929 Grand Banks event, in *Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound,* S. Gregersen and P. W. Basham (Editors), Kluwer Academic, Boston, 547–562.

Hasegawa, H. S. and H. Kanamori (1987). Source mechanism of the magnitude 7.2 Grand Banks earthquake of November 1929: double couple or submarine landslide? *Bull. Seism. Soc. Am.* **77,** 1984–2004.

Johnston, A. C. (1989). The seismicity of stable continental interiors, in *Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound,* S. Gregersen and P. W. Basham (Editors), Kluwer Academic, Boston, 299–327.

Johnston, A. C. and L. R. Kanter (1990). Earthquakes in stable continental crust, *Sci. Am.* **262,** 68–75.

Kafka, A. L. and P. E. Miller (1996). Seismicity in the area surrounding two Mesozoic rift basins in the northeastern United States, *Seism. Res. Lett.* **67** (3), 69–86.

Kemp, D. V. and D. J. Stevenson (1996). A tensile, flexural model for the initiation of subduction, *Geophys. J. Int.* **125,** 73–94.

Kwitko, R. and M. Assumpção (1990). An upper mantle velocity model for Brazil and determination of regional epicenters, *36th Braz. Geol. Congr., Natal, Proc.* **5,** 2464–2469 (in Portuguese).

Lima, C., E Nascimento, and M. Assumpção (1997). Stress orientations in Brazilian sedimentary basins from breakout analysis—implications for force models in the South American plate, *Geophys. J. Int.* **130,** 112–124.

Ludwig, W. J., G. Carpenter, R. E. Houtz, A. G. Lonardi, and F. F. Rios (1978). Sediment isopach map: Argentine continental margin and adjacent areas, *AAPG.*

Meijer, P. T. (1995). Dynamics of active continental margins: the Andes and the Aegean region. *Ph.D. Thesis,* Utrecht Univ., The Netherlands, 218 pp.

Mendiguren, J. A. and F. M. Richter (1978). On the origin of compressional intraplate stresses in South America, *Phys. Earth Planet. Interiors* **16,** 318–326.

Mohriak, W. U. and J. F. Dewey (1987). Deep seismic reflectors in the Campos basin, offshore Brazil, *Geophys. J. R. Astr. Soc.* **89,** 133–140.

Müller, B., M. L. Zoback, K. Fuchs, L. Mastin, S. Gregersen, N. Pavoni, O. Stephansson, and C. Ljunggren (1992). Regional patterns of tectonic stress in Europe, *J. Geophys. Res.* **97,** 11783–11803.

Nábelek, J. (1984). Determination of earthquake source parameters from inversion of body waves, *Ph.D. Thesis,* Massachusetts Inst. of Technology, Cambridge, Massachusetts.

Seeber, L. and J. G. Armbruster (1988). Seismicity along the Atlantic seabord of the U.S.; intraplate neotectonics and earthquake hazard, in *The Geology of North America, The Atlantic Continental Margin: U.S.,* Vols. 1–2, R. E. Sheridan and J. A. Grow (Editors), Geology Society of America, Boulder, Colorado, 565–582.

Stefanick, M. and D. M. Jurdy (1992). Stress observations and driving force models for the South American plate, *J. Geophys. Res.* **97,** 11905–11913.

Stein, S., S. Cloetingh, N. H. Sleep, and R. Wortel (1989). Passive margin earthquakes, stresses and rheology, in *Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound,* S. Gregersen and P. W. Basham (Editors), Kluwer Academic, Boston, 231–259.

Suleiman, A. S., D. I. Doser, and D. R. Yarwood (1993). Source parameters of earthquakes along the coastal margin of West Africa and comparisons with earthquakes in other coastal settings, *Tectonophysics* **222,** 79–91.

Sykes, L. (1978). Intraplate seismicity, reactivation of pre-existing zones of weakness, alkaline magmatism, and other tectonism postdating continental fragmentation, *Rev. Geophys. Space Phys.* **16,** 621–688.

Turcotte, D. L. and G. Schubert (1982). *Geodynamics Applications of Continuum Physics to Geological Problems,* Wiley, New York, 450 pp.

Zoback, M. L. (1992a). First- and second-order patterns of stress in the lithosphere: the World Stress Map Project, *J. Geophys. Res.* **97,** 11703–11728.

Zoback, M. L. (1992b). Stress field constraints on intraplate seismicity in Eastern North America, *J. Geophys. Res.* **97,** 11761–11782.

Zoback, M. L. and R. M. Richardson (1996). Stress perturbation associated with the Amazonas and other ancient continental rifts, *J. Geophys. Res.* **101,** 5459–5475.

Institute of Astronomy and Geophysics
University of São Paulo
P.O. Box 3386
São Paulo, SP, 01065, Brazil

Manuscript received 3 April 1997.