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Geodetic observations of interseismic strain segmentation at the Sumatra subduction zone

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Abstract. Deformation above the Sumatra subduction zone, revealed by Global Positioning System (GPS) geodetic surveys, shows nearly complete coupling of the forearc to the subducting plate south of 0.5°S and half as much to the north. The abrupt change in plate coupling coincides with the boundary between the rupture zones of the 1833 and 1861 (Mw>8) thrust earthquakes. The rupture boundary appears as an abrupt change in strain accumulation well into the interseismic cycle, suggesting that seismic segmentation is controlled by properties of the plate interface that persist through more than one earthquake cycle. Structural evidence indicates that differences in basal shear stress may be related to elevated pore pressure in the north.

Introduction

Subduction zones vary dramatically in their ability to store elastic strain energy. Such variation has been explained by differences in convergence rates and subducting plate ages[*Kanamori*, 1983], presence of subducting sediments and seamounts[*Ruff*, 1989; *Cloos*, 1992], upper plate deformation[*McCaffrey*, 1993], motion of the subducted slab through the mantle[*Scholz and Campos*, 1995], and temperature of the plate interface[*Hyndman and Wang*, 1993]. These proposed models all involve large scale properties of subduction zones. Earthquake histories, used to discriminate among these models, are too short compared to the strain accumulation time to resolve small spatial variations on individual subduction zones. Consequently, some conceptual models of earthquake recurrence, such as seismic gaps, time and slip predictability, and the Gutenberg-Richter magnitude-frequency relationship[*Rundle*, 1989] assume uniform long-term slip with no permanent lateral barriers. Clearly, these models do not hold if lateral variations in plate interface properties cause long-term seismic slip rate to vary along a subduction zone.

Geodetic GPS surveys reveal an abrupt change in the velocity field between adjacent sections of the Sumatra subduction zone well into the interseismic cycle (Plate 1) that we interpret as showing variations in plate coupling. Rupture of two 19th century Mw>8 quakes appear to abut at the boundary between these sections. The change occurs where the Investigator Fracture Zone (IFZ) is subducting, but the coupling difference persists far from it, suggesting that the difference is due to properties of the interface rather than a mechanical barrier, as one might expect from a subducted ridge or seamount [*Kelleher and McCann*, 1976]. We suggest that high fluid pore pressures due to sediment subduction cause weaker coupling in the north.

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GPS Data

A total of 60 sites on Sumatra and the forearc islands were occupied in 1989, 1991, and 1993 (Plate 1). The GPS data comprise dual-frequency (L1 and L2 band) carrier phase and pseudorange measurements. Each site was surveyed for 12 to 22 hours per day over an average of 4 consecutive days. Site coordinates, satellite state vectors, tropospheric zenith delay parameters, and phase ambiguities were estimated for each day by weighted least squares[*King and Bock*, 1995]. We used precise satellite orbits computed at the Scripps Orbit and Permanent Array Center at SIO[*Fang and Bock*, 1996] from data collected by the permanent tracking stations of the International GPS Service for Geodynamics[*Beutler et al.*, 1994]. The daily solutions were then passed to a Kalman filter[*Feigl et al.*, 1993] to estimate site coordinates and velocities.

Interseismic Deformation

The velocity field obtained from the GPS surveys (Plate 1), shows abrupt rotation of the forearc vector azimuths at about 0.5°S, near the Batu islands. South of 0.5°S the forearc vectors are roughly parallel to the convergence direction of the Indian Ocean and Eurasia. In the north, the vectors are more parallel to the Sumatra Fault (SF). This pattern suggests strong coupling of the forearc to the subducted plate in the south and weaker coupling in the north.

To quantify the change in coupling, we calculate the interseismic surface deformation using *Okada's* [1985] formulation, in which locked faults are modeled as dislocations in a half-space. We describe the subduction zone by 3 elastic blocks (Eurasia, forearc, and Australia) separated by two faults: the SF and the thrust fault. The rigid-body motion of each block is specified by a pole of rotation and fault surfaces separating the blocks are specified by nodes. The slip rate v at each node is the difference in the local velocities of the blocks that are in contact across the fault. To model strain accumulation, the slip deficit (locking rate) imposed at each node is $v\chi$[*Savage*, 1983], where χ is defined as the fraction of seismic slip on the fault. Hence, the aseismic slip rate across a node is $v(1\text{-}\chi)$. We model variation in coupling by specifying χ for each node and integrating the slip deficit over the fault surfaces. The SF is modeled as a vertical strike-slip fault, locked to a depth of 15 km, with a dextral, deep slip rate of 30 mm/yr (a geologic estimate for slip rate on the SF at 2°N is 28 mm/yr[*Sieh et al.*, 1991]). The thrust fault comprises three connected surfaces, with dip angles increasing

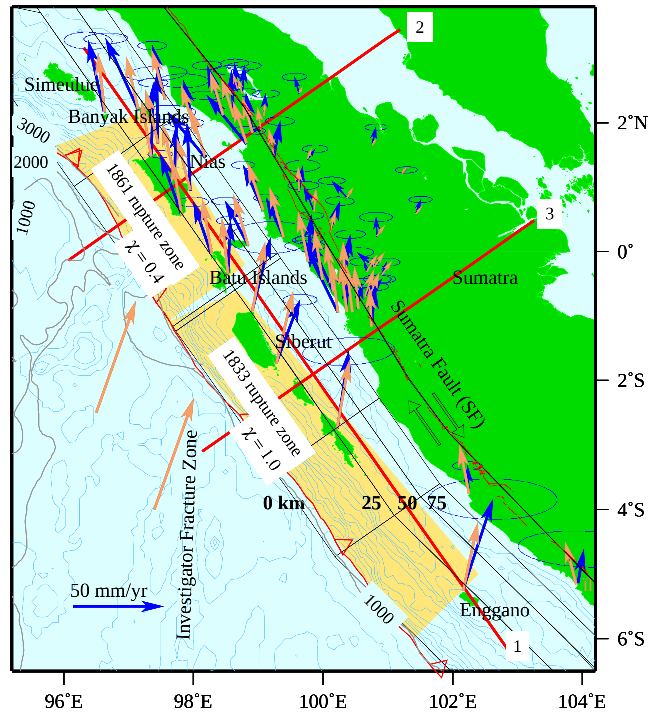

Plate 1. Map of bathymetric contours (500m), and sediment isopachs (gray contour lines, meters) [*McDonald,* 1977], velocity vectors from GPS (blue arrows) relative to Eurasia [*DeMets et al.,* 1994], and from our model (orange arrows). Yellow shaded areas show rupture zones for great events. χ is geodetic coupling coefficient discussed in the text. Black lines outline modeled fault surfaces, with contours at 0, 25, 50, and 75 km depths. Open arrows show motion on SF. Arrows in Indian Ocean show plate motion relative to Eurasia. Red lines indicate location of profile lines discussed in Figure 1. Ellipses indicate 95% confidence level assuming a white noise model for the daily GPS positions. Time series analysis of continuous GPS data by *Zhang* [1996] indicates that velocity uncertainties may need to be scaled by a factor of 2-3 to account for colored noise in the daily positions

with depth (Plate 1 and Figure 1G). The dip angle (12°) of the shallowest segment is estimated from seismic refraction profiles[*Kiekhefer*, 1980] and deeper slab geometry is inferred from earthquake hypocenters[*Fauzi et al.*, 1996]. Different values of χ are used for the north and south sections of the thrust fault above 50 km depth. Between 50 and 75 km, χ decreases linearly to zero on both sections.

Agreement between modeled and observed velocities is best revealed by examining their arc-parallel and arc-normal components separately (Figure 1). Because convergence of the Indian Ocean with the forearc is nearly normal to the trench (as shown by earthquake slip vectors,[*McCaffrey*, 1991]), coupling on the thrust fault influences only the arc-normal components. The arc-parallel components, indicating slip on the SF are, on average, consistent with about 30 mm/yr strike-slip and 15 km locking depth (Figures 1D and 1F). The positive gradient in the arc-parallel vectors north of 0.5°S indicates arc-parallel stretching at a strain rate of 3x10^{-8}/yr (20 mm/yr change over 600 km; Figure 1B). Similar stretching is not observed in the southern forearc, where our model indicates stronger coupling. Along-strike, the arc-normal components show abrupt change north of the Batu Islands (Figure 1A) where they suggest that χ is 0.4 or less (Figure 1C). In the southern forearc, large arc-normal components require nearly complete coupling to the subducting slab (Figure 1E). The coupling transition is seen in the surface displacements over a few tens of km along strike (Figure 1A) and the transition on the plate interface must be at least this narrow (we arbitrarily use a 10 km wide transition zone). The modeled strain accumulation on the thrust fault is also consistent with 5-10 mm/yr subsidence rates estimated from coral heads on islands west of Nias[*Zachariasen et al.*, 1995]. Such subsidence is attributed solely to elastic strain because there is no evidence there for large permanent subsidence.

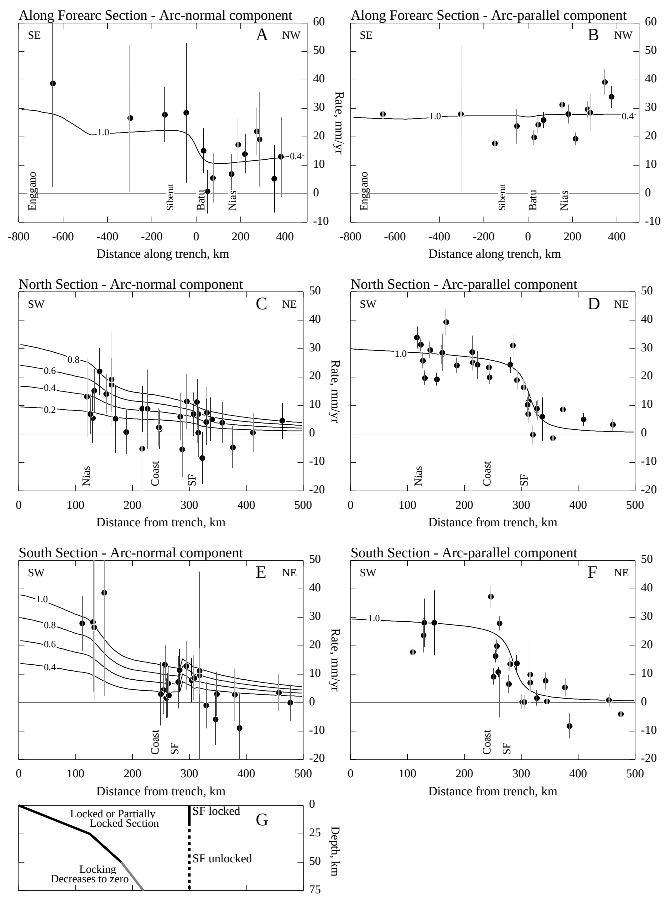

Figure 1. Along-strike variation in arc-normal (A) and arc-parallel (B) components of velocity vectors located within 200 km of trench. Curves show predicted values on profile line 1 (see Plate 1 for location of profile lines). (C) through (F) show across-arc variations in arc-parallel and arc-normal vector components in north and south sections. Curves are predicted values calculated along profile lines 2 (C,D), and 3 (E,F), for various coupling coefficient (χ) values, as labeled. (G) Profile of subducting plate model. We assume χ for each thrust fault segment is uniform down to 50 km, then decreases linearly to zero at 75 km depth. The SF is fully locked to 15 km depth and slips freely below that.

Discussion

The coupling change near 0.5°S coincides with the boundary between ruptures of the 1833 (Mw≈8.7) and 1861 (Mw≈8.4) great earthquakes (Plate 1). The 1833 event ruptured about 600 km of the plate margin from Enggano to the Batu islands. The 1861 quake ruptured between the Banyak and Batu islands, a 300 km segment northwest of and abutting the 1833 rupture[*Newcomb and McCann*, 1987]. These rupture lengths were estimated from the extent of reported maximum intensities. Comparing these estimates with rupture lengths of great events on similar subduction zones[*Acharya*, 1979], and taking into account the uncertainty in the historical records, we ascribe an uncertainty of at least 50 km to the rupture boundary location. Smaller (Mw≈7) events in 1907, 1908, 1914, and 1921 also appeared not to cross the 0.5°S boundary[*Newcomb and McCann*, 1987]. The fact that the estimated rupture boundary of past thrust events coincides with the coupling boundary suggests that asperities on the Sumatra subduction zone may remain stationary through more than one earthquake cycle instead of behaving dynamically as rate-dependent features of the fault zone[*Cochard and Madariaga*, 1994].

A similar along-arc change in geodetically measured coupling occurs at the Alaska subduction zone from the unlocked Shumagin seismic gap to 400 km east where it is fully locked[*Lisowski et al.*, 1988]. The nature of decrease is unknown because of limited lateral geodetic coverage. Either the main thrust fault beneath the Shumagin Islands is presently unlocked or asthenospheric relaxation causes strain rates to be relatively small late in the seismic cycle[*Lisowski et al.*, 1988]. Relaxation cannot apply to Sumatra unless the repeat times for the earthquakes north and south are quite different, because the north side, with the more recent great event, now shows less coupling. The asthenospheric relaxation model may hold if the south side has a much longer cycle and is now earlier in that cycle than the north side.

The IFZ subducts near the coupling boundary, but we think that its elevated topography does not increase coupling because sites on the forearc near 98°E, directly above the IFZ, show uncoupled behavior, and the coupling difference persists far from the IFZ. Sea-floor magnetic anomalies reveal a 10 Ma age difference across the IFZ[*Royer and Sandwell*, 1989] but the younger, more buoyant oceanic crust in the NW corresponds with lower coupling. There is no resolvable variation in dip angle along the forearc[*McCaffrey*, 1994] and any dip angle difference would have to be very large to appreciably change the surface vectors. In the south, the forearc basin is deep and gravity is low suggesting that the upper and lower plates are in poor contact, yet coupling is high. In the north where coupling is low, the forearc basin is shallow and free-air gravity is high[*Sandwell and Smith*, 1994], which would normally suggest that the forearc is in strong contact with the subducted slab. Our present knowledge of forearc basin structure does not explain the abrupt change in coupling.

Our interpretation of the observed coupling change is that pore pressures along the thrust fault in the north are elevated by subduction of thicker sediments (Plate 1) from the Bengal fan[*Curray et al.*, 1980], bringing excess water into the fault zone. High pore pressure enhances aseismic slip by decreasing the effective normal stress on the fault[*Scholz*, 1990]. Support for this theory is seen in the accretionary wedge structure. *Davis et al.* [1983] showed that wedge topographic slope (α) and basal fault dip (β) may be related to basal shear stress and pore pressure as a fraction of lithostatic pressure. For a seismic profile south of Siberut[*Karig et al.*, 1980*a*], α=6°and β=5°, indicating that pore pressure is a small fraction of lithostatic pressure. A profile SW of Nias[*Moore and Curray*, 1980] shows α=6°and β=3°, suggesting that pore pressure is about 80% of lithostatic pressure. Furthermore, *Karig et al.* [1980b] cite oil company profiles as showing landward verging folds in the accretionary prism off Simeulue (about 2°N) indicative of low basal shear stress, probably due to high pore pressures[*Seely*, 1977]. These observations alone are insufficient to explain the contrast in coupling because they do not necessarily reflect shear stress on the seismogenic part of the plate boundary. However, if pore pressure below the accretionary prism is increased by subducting fluids then it may also be higher along deeper parts of the thrust fault. If the seismic character of subduction zones is controlled by a property as spatially variable as pore pressure, earthquake hazard assessment models that assume spatially uniform slip on entire subduction faults may be unrealistic.

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