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Slab Pull and the Seismotectonics of Subducting Lithosphere

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This synthesis links many seismic and tectonic processes at subduction zones, including great subduction earthquakes, to the sinking of subducted plate. Earthquake data and tectonic modeling for subduction zones indicate that the slab pull force is much larger than the ridge push force. Interactions between the forces that drive and resist plate motions cause spatially and temporally localized stresses that lead to characteristic earthquake activity, providing details on how subduction occurs. Compression is localized across a locked interface thrust zone, because both the ridge push and the slab pull forces are resisted there. The slab pull force increases with increasing plate age; thus because the slab pull force tends to bend subducted plate downward and decrease the force acting normal to the interface thrust zone, the characteristic maximum earthquake at a given interface thrust zone is inversely related to the age of the subducted plate. The 1960 Chile earthquake (M_W 9.5), the largest earthquake to occur in historic times, began its rupture at an interface bounding oceanic plate <30 m.y. old. However, this rupture initiation was associated with the locally oldest subducting lithosphere (weakest coupling); the rupture propagated southward along an interface bounding progressively younger oceanic lithosphere, terminating near the subducting Chile Rise. Prior to a great subduction earthquake, the sinking subducted slab will cause increased tension at depths of 50–200 km, with greatest tension near the shallow zone resisting plate subduction. Plate sinking not only leads to compressional stresses at a locked interface thrust zone but may load compressional stresses at plate depths of 260–350 km, provided that the shallow sinking occurs faster than the relaxation time of the deeper mantle. This explains K. Mogi's observations of $M \geq 7$ thrust earthquakes at depths of 260–350 km, immediately downdip and within 3 years prior to five great, shallow earthquakes of northern Japan. The slab pull model explains the lower layer of double seismic zones as due to tension from the deeper, sinking plate and the upper layer as due to localized in-plate compression, as plate motion is resisted by the bounding mantle. Just downdip of the interface thrust zone, there occurs an aseismic 20°–50° dip increase of subducted plate. This slab bend reflects the summed slab pull force of deeper plate and probably is at the crustal basalt to eclogite phase change. Resistance to subduction provided by a continually developing slab bend may be an important factor in the size of slab pull force delivered to an interface thrust zone.

INTRODUCTION

Although the slab pull force is recognized as dominant in moving tectonic plates, mechanisms by which this force is transmitted through locked subduction zones have received little attention. This is because many investigators believe that most of the slab pull force is balanced by friction between the plate and the surrounding mantle. Reasons for the persistence of this view include (1) the perception that great, interface thrust earthquakes at subduction zones primarily are caused by compression from the oceanic side, (2) results from glacial rebound studies indicating that mantle viscosity is high enough to inhibit deep plate motions from influencing shallow tectonics, and (3) calculations of vertical forces acting at trenches, based on observed trench topographies, indicating much smaller forces than are predicted from thermal modeling of subducted plate. In this paper a synthesis of earthquake and tectonic data weakens the main objections to accepting the dominance of the slab pull force in driving the earthquake cycle. Also, this synthesis provides the conceptual framework for reinterpretation of several important dynamic properties of subduction zones.

FORCES THAT MOVE LITHOSPHERIC PLATES

Plate-driving forces largely are due to within-plate density contrasts. These forces largely determine how plates move at subduction zones and consequently determine how stresses accumulate there.

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Ridge Push and Slab Pull

Horizontal density contrasts, resulting from cooling and thickening oceanic lithosphere, produce the ridge push (sliding plate) force which contributes significantly to plate motions [*Hales*, 1969; *Anderson et al.*, 1976; *Lister*, 1975; *Hager*, 1978; *Solomon et al.*, 1980; *Hager and O'Connell*, 1981]. The negative buoyancy of subducted oceanic lithosphere (the slab pull force) is established as the major driving force for plate motions [*Elsasser*, 1969; *McKenzie*, 1969; *Isacks and Molnar*, 1971; *Smith and Toksöz*, 1972; *Forsyth and Uyeda*, 1975; *Schubert et al.*, 1975; *Solomon et al.*, 1975; *Vlaar and Wortel*, 1976; *Richter*, 1977; *Chapple and Tullis*, 1977; *Carlson*, 1981, 1983].

Both a trench suction force and back arc spreading have been suggested as mechanisms for the trenchward movement of the leading edges of overriding plates [*Forsyth and Uyeda*, 1975; *Carlson*, 1983; *Chase*, 1978; *Karig*, 1974], but these mechanisms are shown to be related to the sinking and seaward propagation of subducted plate [*Garfunkel et al.*, 1986]. Hot spot push also may contribute to driving plate motions [*Morgan*, 1972; *Chase*, 1978].

Forsyth and Uyeda [1975] determined that the slab pull force, F_{SP}, is 10 times more important than the ridge push force, F_{RP}, in moving oceanic lithosphere but also suggested that most of the slab pull force is balanced by viscous dissipation in the mantle rather than a large component being balanced near shallow subduction zones. A linear inversion of plate velocities relative to specific subduction zone geometries by *Carlson* [1983] indicates that the slab pull force is about 3 times more significant than the ridge push force in determining plate velocities.

The slab pull force cannot initiate subduction. However, as

TABLE 1. Great Normal-Faulting Earthquakes, 1930 to the Present

Date	Location	M_0, dyn cm	Depth	Reference
Jan. 15, 1931	Oaxaca	3.5×10^{27}	shallow	*Singh et al.* 1985]
March 2, 1933	Sanriku	4.3×10^{28}	shallow	*Kanamori* [1971]
Nov. 4, 1963	Banda Sea	3.1×10^{28}	100 km	*Osada and Abe* [1981]
May 26, 1964	South Sandwich Islands	6.2×10^{27}	120 km	*Abe* [1972*a*]
May 31, 1970	Central Peru	1.0×10^{28}	60 km	*Abe* [1972*b*]
June 22, 1977	Tonga	2.3×10^{28}	65 km	*Silver and Jordan* [1983]
Aug. 19, 1977	East Sunda arc	$2.4–4.0 \times 10^{28}$	shallow	*Spence* [1986]

initially subducted lithosphere enters the upper mantle, it is apparent that slab sinking, due to the slab pull force, becomes increasingly dominant in subduction processes. The motions of plates and subducted lithospheres are important components of global patterns of mantle flow [*Hager et al.*, 1983; *Garfunkel et al.*, 1986].

Earthquake Evidence for Relative Size of Slab Pull and Ridge Push Forces

Although modeling of plate motions implies that the slab pull force is the dominant force at subduction zones, earthquake data have not been directly used to confirm this conclusion. The episodic nature of subduction is implicit in the widely accepted seismic gap hypothesis [*Fedotov*, 1965; *Mogi*, 1968; *Kelleher et al.*, 1973; *McCann et al.*, 1979; *Sykes and Quittmeyer*, 1981]. In the time interval between repeating great earthquakes at a seismic gap, plate motion there is largely blocked, and stresses accumulate. An examination of the sizes of earthquakes downdip and updip of interface thrust zones, outside of the times of great interface thrust earthquakes, should indicate the plate strains there and the relative size of the slab pull and ridge push forces acting at shallow subduction zones.

The four great normal-faulting earthquakes at depths greater than 50 km in Table 1 occurred downdip of the corresponding interface thrust zones. Also, the great Oaxaca earthquake probably is below the primary interface thrust zone. Although shallow-dipping lithosphere exists downdip of the central Peru and Oaxaca events (Figure 1), such normal-faulting earthquakes require extensional tectonics. The downdip tension axes for these five great earthquakes are consistent with their being caused by slab pull forces.

The great, normal-faulting Sanriku and East Sunda arc earthquakes (Table 1) occurred near oceanic trenches. Old and deeply subducted lithosphere exists downdip of these earthquakes, and the corresponding large slab pull forces are interpreted to have pulled the oceanic plate away from the overriding plate, partially decoupling the interface thrust zone. These two earthquakes are interpreted as due to slab pull stresses that have been transmitted through partially decoupled interface thrust zones to add to bending stresses at the trench zones [*Spence*, 1986].

While earthquakes due to compression sometimes occur in the oceanic lithosphere at the trench vicinity and seaward [*Christensen and Ruff*, 1983; *Hanks*, 1971; *Mendiguren*, 1971; *McAdoo et al.*, 1978], these earthquakes are unusual, and none have been observed with the size of the normal-faulting earthquakes of Table 1. The largest known such thrust event is the 1981 M_S 7.2 earthquake located seaward of the 1985 Valparaiso interface thrust earthquake [*Christensen and Ruff*, 1983, 1985; *Nishenko*, 1985]. The earthquake data of this section indicate that at shallow subduction zones the maximum tensional stresses of slab pull origin are significantly greater than the maximum compressional stresses of ridge push origin.

SLAB PULL AS A CAUSE OF SUBDUCTION ZONE EARTHQUAKES

In order to provide a framework for the development of this paper, the tectonic model advocated here is summarized. Subsequent sections provide observational support for the model or deal with model refinements.

How can tensional stresses arising from slab pull forces lead to thrust-faulting earthquakes at an interface thrust zone when such earthquakes imply that compression acts across this interface? Consider the analogy of pulling a massive, rigid box across a rough floor. While the body forces within the box would be tensional, a strong box would not deform significantly but would tend to translate, with compressional stresses localized at the interface between the box and floor. In the case of a subducting oceanic lithosphere, the slab pull force is resisted by the overriding plate, causing compression at the interface thrust zone.

Prior to a great subduction earthquake, the subducted plate slowly sinks, increasing extensional stresses at depths of 50 to 200 km, with the greatest stresses at shallow depths. Although slab pull stresses are guided updip, these stresses are diminished by forces resisting subduction, leaving 5–10% of the gross slab pull force to be supported at and near locked interface thrust zones, consistent with the views of *Shimazaki* [1974], *Davies* [1980], *Reyners and Coles* [1982], and *Spence* [1985]. Finally, stresses originating with the ridge push and slab pull forces exceed the strength of the locked interface thrust zone, and a great thrust earthquake occurs, with the maximum compressive stress, σ_1, oriented with the relative plate motion. Postseismically, because the slab pull force exceeds the ridge push force at interface thrust zones, some extension propagates seaward from the rupture zone, causing plate bending and occasional normal-faulting earthquakes near the trench. This regionally and temporally localized tension facilitates a trenchward diffusion of the oceanward horizontal plate, acting under ridge-push forces. Also postseismically, an independent strain pulse moves downdip from the rupture zone, returning the downdip plate to a less extended state. After the rupture of the great earthquake reheals, the process repeats.

MANTLE VISCOSITY AND SUBDUCTION

The slab pull force can act strongly at shallow depths, provided that the viscosity of the surrounding mantle is less than about $2–3 \times 10^{20}$ Pa s ($2–3 \times 10^{21}$ P) [*McKenzie*, 1969; *Hager and O'Connell*, 1981; *Hager et al.*, 1983]. Direct esti-

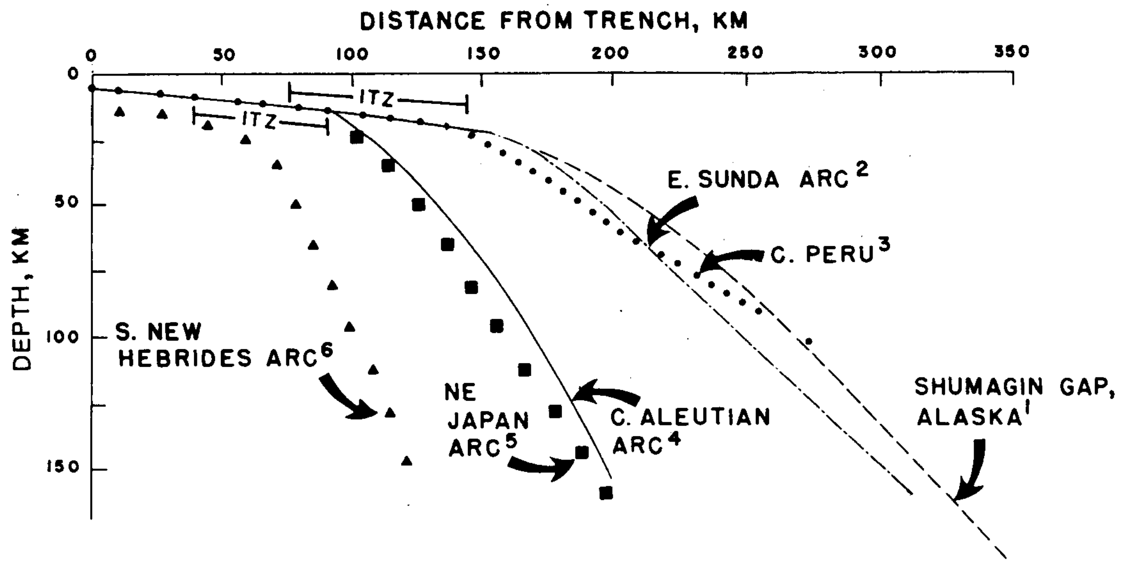

Fig. 1*a*. Dips of well-located subducting plates, at depths of 0–150 km. Sections are perpendicular to arcs. The dip of typical plate increases from about 10° at the interface thrust zone (ITZ) to 30°–70° in the upper mantle. This slab bend occurs along about 40 km of plate length, just downdip from the interface thrust zone. Sources for plate dips are (1) *Hauksson* [1985], (2) *Spence* [1986], (3) C. Langer and W. Spence (unpublished manuscript, 1984), (4) *Engdahl et al.* [1984], (5) *Hasegawa et al.* [1978], and (6) *Coudert et al.* [1981].

mates of mantle viscosity, based on lithospheric adjustment rates after great earthquakes, are 5×10^{18} Pa s [*Nur and Mavko*, 1974], 6×10^{18} Pa s [*Spence*, 1977], and 1×10^{19} Pa s [*Thatcher et al.*, 1980]. These estimates are determined within earthquake zones and within a small fraction of the earthquake cycle. These values are somewhat smaller than the top layer viscosities in the mantle models of *Cathles* [1975] and *Yokokura* [1981]. Cathles obtained a viscosity of 4×10^{19} Pa s between depths of 64 and 128 km and 10^{21} Pa s for the rest of the mantle, from consideration of glacial rebound data.

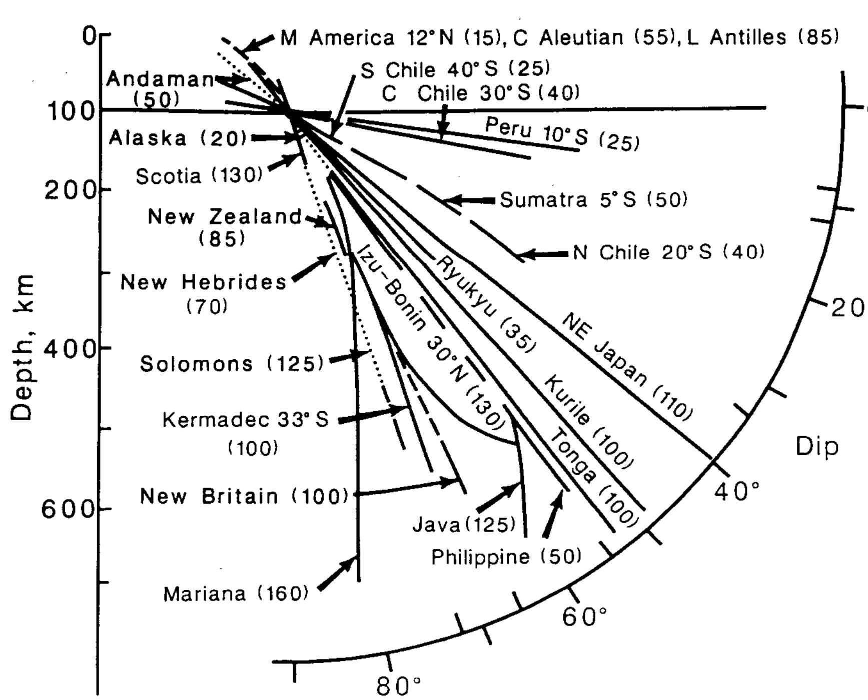

Fig. 1*b*. Dips of Benioff-Wadati zones below the slab bend zone; dip angle is referred to plate depth of 100 km (adapted from *Uyeda and Kanamori* [1979]). Dashed and dotted lines indicate zones with few or no earthquakes. Ages of plates at trenches are shown by numbers in parentheses [*Schlater et al.*, 1981]. Plate dips are positively correlated with age.

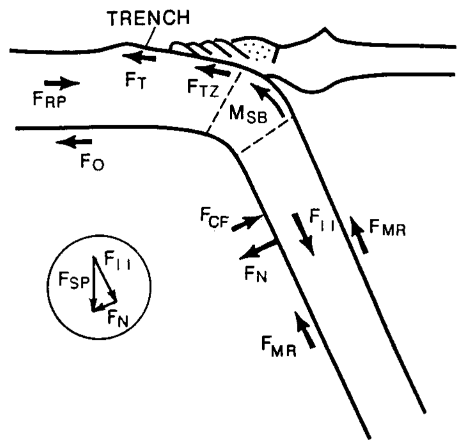

Fig. 2. Diagram of driving and resisting forces acting on subducting plate. The slab pull force, F_{SP}, shown by inset force diagram, acts throughout the entire length of subducted slab. Forces opposing plate motions include the viscous resistance to downdip plate motion provided by the mantle, F_{MR}, the corner flow that resists the plate falling to the vertical, F_{CF}, the moment leading to internal deformation and torquing at the slab bend, M_{SB}, the resistance to plate motion at the interface thrust zone, F_{TZ}, the forces leading to formation of the trench and outer rise, F_T, and the resistance of the attached oceanic plate to trenchward motion, F_0. Dashed lines indicate zone of slab bend.

Yokokura obtained viscosities of about 10^{19} Pa s for the top 100 km of the upper mantle, 10^{21} Pa s for the rest of the upper mantle, and about 10^{22} Pa s for below about 650 km, from consideration of plate dips and plate torques due to the negative buoyancy of subducted plates. The depths of most normal-faulting earthquakes in the upper mantle are within the low-viscosity, top layer of *Cathles* [1975] and *Yokokura* [1981].

The viscosity of 10^{21} Pa s beneath the top layer of the upper mantle seems to be too high to permit observed deep plate motions. During a subduction cycle, variations of pressure within the subducted plate and corresponding pressure variations in the immediately surrounding mantle may lead to temporarily lowered mantle viscosity [*Brady*, 1976] and more easily permit downdip movement of subducting plate.

Slab Pull Force and Force Balances

The slab pull force is proportional to the excess mass of the subducted oceanic lithosphere in relation to the mass of the warmer, displaced mantle. This excess mass is a function of the subducted plate's volume, and its positive density contrasts with the surrounding mantle.

McKenzie [1969] calculated the density contrast times volume of a deeply subducted plate, based only on temperature considerations, and found that the slab pull stress acting at shallow regions should be 100–200 MPa (1–2 kbars), if the mantle provided no resistance. *Schubert et al.* [1975] calculated this stress to be 65 MPa (0.65 kbar). Similar values for the slab pull stress are obtained by *Fitch* [1977], *Molnar and Gray* [1979], *Davies* [1980], and *Wortel* [1986]. Because subducted plates have widely different ages and penetration depths, there exist wide variations in the magnitude of the slab pull force. For example, *Davies* [1983] calculates the effective slab pull force acting at the Marianas trench to be nearly 15 times greater than at the Aleutian trench.

Schubert et al. [1975] and *Fitch* [1977] also calculated the additional, large body forces due to the phase change olivine to β spinel and a significant body force due to the deeper spinel to postspinel phase change. The depth of the olivine to β spinel transition may vary significantly between slabs, as this depth is a function of the internal temperature distribution in a slab, which in turn depends on slab age, slab thickness, and subduction rate [*Sung and Burns*, 1976*a*, *b*; *Liu*, 1983; *Rubie*, 1984]. Thus wide variations exist between slabs in the contribution to the slab pull force from the olivine to β spinel phase change.

The insets in Figures 2 and 3 show forces acting at a unit volume within the subducted slab that are due to that unit volume's excess mass. The primary vertical force (F_{SP}) is broken into two components, one parallel to the dip δ of the subducted plate ($F_{\parallel}$) and the other normal to the outer surfaces of the subducted plate (F_N). Similar forces pertain throughout the plate having excess mass. The downdip *T* or *P* stress axes of many mantle earthquakes [*Isacks and Molnar*, 1971] imply that the subducted slab is a stress guide at least down to the 650-km discontinuity. $F_{\parallel}$ sums throughout the length of subducted plate and promotes downdip plate motion. F_N acts to move the entire plate trenchward [*Tovish et al.*, 1978; *Yokokura*, 1981; *Carlson and Melia*, 1984; *Garfunkel et al.*, 1986].

The slab pull and ridge push forces act together to subduct oceanic lithosphere. Because subduction occurs, $F_{SP} + F_{RP}$ exceeds the sum of resisting forces. Corner flow mechanisms [*Jischke*, 1975; *Tovish et al.*, 1978; *McAdoo*, 1982] preserve plate dips below 100 km depth (Figure 1*b*). The seaward propagation of trenches [*Carlson and Melia*, 1984] combined with the fairly constant dip of deeply subducted plate implies that the entire subduction system slowly propagates seaward [*Garfunkel et al.*, 1986], i.e., $F_N > F_{CF}$ (Figure 2). The rate of seaward advance of plate is positively correlated with plate age [*Garfunkel et al.*, 1986], and related increases in F_{CF} and M_{SB} may explain the lack of strong correlation of subduction velocity with plate age [*Ruff and Kanamori*, 1980].

Slab Bend

Subducting oceanic lithosphere dips about 8°–15° beneath the accretionary wedge, until a depth of 20–40 km above the top of the subducted plate is reached [*Pennington*, 1983, 1984; *Davies and House*, 1979; *Ruff and Kanamori*, 1983*b*] (Figure 1*a*). Then the plate dip steepens to 30°–70° (Figure 1*b*), forming the slab bend. The six slab bend zones shown in Figure 1*a* are for subducting plates whose positions are well resolved from detailed seismicity studies; the slab bend zone is indicated in Figures 2 and 3. The slab bend occurs within about 40 km of plate length, which for an average convergence rate of 6 cm/yr occurs over a period of about 600,000 years. The degree of bending at slab end zones is much greater than at trench–outer rise systems. However, large earthquakes generally are absent at the slab bends, implying that the rheology throughout slab bends is largely anelastic.

Because of the beam strength of oceanic plate, the plate resists being bent downward and consequently remains coupled to the overriding plate at the interface thrust zone. A great interface thrust earthquake nucleates just updip of the slab bend and then propagates updip and laterally [*Kelleher and Savino*, 1975; *Dewey and Spence*, 1979; *Davies et al.*, 1981; *Lay et al.*, 1982; *Dmowska and Li*, 1982]. Downdip from the

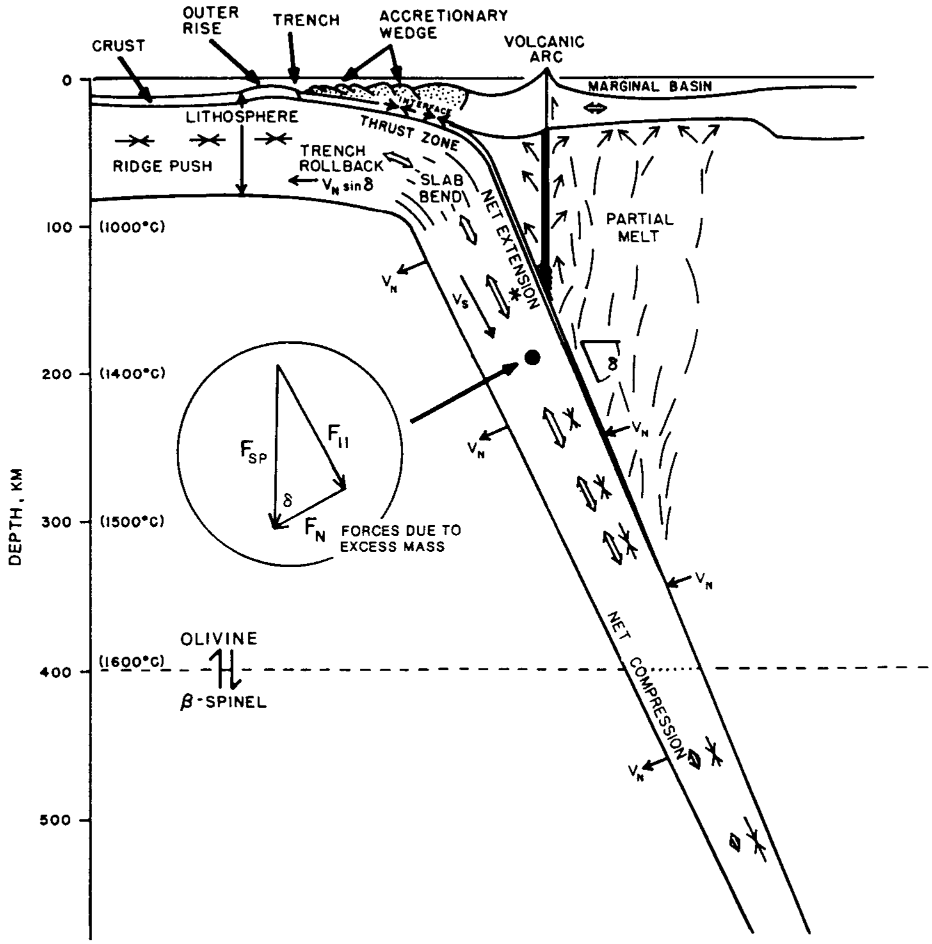

Fig. 3. Schematic diagram of the role of slab pull force in establishing steady state conditions at a subduction zone. The greatest plate bending is at the slab bend zone, just downdip of the interface thrust zone. Unlabeled arrows within the plate show extension (open arrows) and compression (arrows meeting) subducted; the body force at a given depth point is the sum of plate push from above, plate pull from beneath, and resistive forces (Figure 2). The lithospheric thickness is drawn for 60-m.y.-old lithosphere at the trench [*Watts et al.*, 1980]. Back arc basins tend to be above old subducting plate [*Molnar and Atwater*, 1978; *Ruff and Kanamori*, 1980; *England and Wortel*, 1980; *Carlson*, 1983; *Carlson and Melia*, 1984; *Garfunkel et al.*, 1986]. Temperatures of oceanic asthenosphere and the boundary of the olivine to β spinel phase change are from *Schubert et al.* [1975].

slab bend, earthquakes generally occur within the subducted plate.

Several investigators [*Rogers*, 1982; *Pennington*, 1982, 1984; *Ruff and Kanamori*, 1983*b*] suggest that the slab bend is associated with various dehydration reactions and the phase change in the subducted oceanic crust from "wet" gabbro to "dry" eclogite [*Ahrens and Schubert*, 1975]. Crustal hydration may be enhanced by the migration of water through normal faults that developed near oceanic trenches and that have moved downdip with the subducting plate.

The volume decrease associated with this crustal phase change in dry minerals is about 17%. Assuming total conversion within a 5- to 7-km-thick crust implies that the subducting crust would withdraw from the interface zone by about 1 km, decoupling the interface thrust zone [*Ruff and Kanamori*, 1983*b*]. The decoupling is aided by crustal thinning due to this bending. Forearc basins, on the inner trench slope, are positioned above the slab bend [*Rogers*, 1982], and thus the downdip edge of a forearc basin should correspond to the downdip edge of the interface thrust zone.

The dip angle of deeply subducted plate is positively correlated with lithospheric age [*Veith*, 1974; *Jarrard*, 1986] (Figure 1*b*), implying that the slab pull force is a major determiner of plate dip. In this paper the slab bend is interpreted to be a steady state pivot for subducted slab that is below the bend, reflecting the summed slab pull force of the deeper plate. The aseismic slab bend occurs at the zone of the crustal gabbro to eclogite phase change, because that rheology and plastic behavior in the remaining bending lithosphere [e.g., *McAdoo et al.*, 1978] permit the slab pull force to most easily bend the plate to its mantle equilibrium position.

Trench Depth Discrepancy

Trenches and outer rises are due to plate bending [*Hanks*, 1979; *Chapple and Forsyth*, 1979] and are dynamically sustained by downdip mass transfer beneath the trenches [*Davies*, 1983; *Chase and McNutt*, 1982]. Attempts to reconcile the calculated excess mass of a subducted plate with actual trench topography find that by taking reasonable values of F_{TZ} and F_{MR}, the calculated force due to slab pull is 3–10^+ times greater than that expressed in trench topography [*Chase and McNutt*, 1982; *Hager et al.*, 1983; *Davies*, 1983]. Because

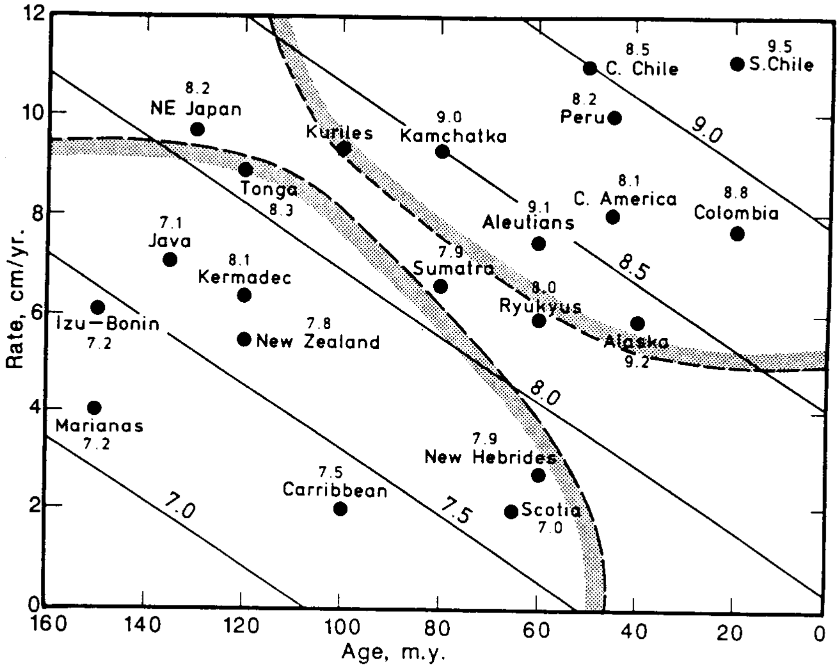

Fig. 4. Maximum characteristic earthquakes M_W, plotted as functions of subduction rate and age of subducted plate (adapted from *Ruff and Kanamori* [1983*b*]). Included in lower left-hand corner are oldest plates, which show weakest coupling and are associated with marginal basin formation. Included in upper right-hand corner are youngest plates, which show strongest coupling and are associated with seaward advance of overriding plate.

entire subduction systems tend to propagate seaward, it is implied that slab bend zones track this seaward propagation. The resistances to bending and seaward propagation of the slab bend may be an important resistance to the slab pull force acting at shallow subduction zones, and this neglected resistance may partly explain the trench depth discrepancy. Another factor which could help resolve the trench depth discrepancy is the episodic application of bending moment at the trench and outer rise, following decoupling earthquakes. As a ruptured interface thrust zone reheals, the slab pull force acting updip of the interface would be blocked, and thus the slab pull force would act at the trench and outer rise only for a small fraction of an earthquake cycle.

SEISMIC COUPLING

Ruff and Kanamori [1980] and *Jarrard* [1986] show that the force acting normal to an interface thrust zone is related to the age of the subducting lithosphere and by implication is related to the size of the ridge push and slab pull forces. Other factors that determine seismic coupling are the rate of seaward advance of plate in the back arc zone, the load of the accretionary wedge, the resistance of the plate to downward bending, the width and asperity content of the interface thrust zone, and possibly the quantity and composition of subducted sediments. As is implied in Figure 4, coupling due to the ridge push force and resistance to bending of oceanic lithosphere are more than offset by the decoupling due to the corresponding slab pull force.

Decoupling earthquakes with sizes greater than M_W 8.7 or M_0 10^{29} are shown in the upper right-hand corner of Figure 4, and these earthquakes indicate that factors that maximize seismic coupling. They occur where the back arc plate is advancing seaward and, except for the Kamchatka event, have occurred in zones where the age of the oceanic lithosphere is <60 m.y. B.P. These observations imply that seismic coupling is maximized by the buoyancy of young oceanic lithosphere and seaward advance of back arc plate. High seismic coupling leads to large asperities, whose ruptures cause great subduction earthquakes [*Lay et al.*, 1982; *Ruff and Kanamori*, 1983*a*].

The vertical force acting on very old (>80 m.y. B.P.) and dense subducted plate will minimize coupling at an interface thrust zone [*Ruff and Kanamori*, 1980; *Davies*, 1983; *Carlson et al.*, 1983; *Spence*, 1986], resulting in the absence of great interface thrust earthquakes and sometimes a fairly steady subduction process, as may be occurring at the Marianas. Low coupling can allow stresses of slab pull origin to migrate updip and directly add to bending stresses beneath an oceanic trench, causing great normal-faulting earthquakes such as those that occurred at Sanriku in 1933 and at the East Sunda arc in 1977.

The stresses that are loaded at interface thrust zones are obtainable from studies of great earthquakes. Stress drops of 1–6 MPa (10–60 bars) are representative for great interface thrust earthquakes [*Lay et al.*, 1982]; the total stresses (dynamic stresses) inferred from the rupturing of asperities are not significantly different from the static results of *Lay et al.* [1982] (G. Choy, personal communication, 1986), supporting the assumption of *Kanamori* [1977] of total stress drops for great earthquakes. The stress drop for the great 1977 Sumba earthquake, which faulted through the brittle oceanic lithosphere, was about 20 MPa [*Spence*, 1986]. The stresses greater than 60 MPa determined at the Shumagin gap [*House and Boatwright*, 1980] are atypical. Generally, stresses of subduction earthquakes are considerably less than those available from the total slab pull force of 50–200 MPa (0.5–2.0 kbars) and imply that much of the slab pull force is diminished by resisting forces that act more deeply than the thrust zone and trench.

The repeat times for great, interface thrust earthquakes vary from 35 to 150$^+$ years [*Sykes and Quittmeyer*, 1981]. Variations in strengths of the plate-driving and -resisting forces at a given seismic gap zone determine the magnitude and repeat time of earthquakes for that zone. Given enough time, even plate less than 30 m.y. old will sink sufficiently to strongly load a locked interface thrust zone.

THE GREAT 1960 CHILE EARTHQUAKE

The May 22, 1960, Chile earthquake is the greatest known earthquake in recorded history (M_W 9.5; M_0 2×10^{30} dyn cm [*Kanamori*, 1977]). The main shock rupture was about 900 km long, 60–200 km wide, and had an average displacement of 20–25 m along a low-angle thrust fault [*Plafker and Savage*, 1970; *Kanamori and Cipar*, 1974]. The area of rupture is indicated in Figure 5 by the primary earthquakes and vertical deformation associated with this earthquake series. The main shock produced a great, ocean-wide tsunami, resulting from uplift of the seafloor [*Plafker and Savage*, 1970].

In the 23 hours prior to the main shock, there were numerous foreshocks (including two of M $7\frac{1}{4}$–$7\frac{1}{2}$) at the northern end of the Mocha block. The main shock nucleated near these foreshocks and propagated southward [*Kanamori and Cipar*, 1974] to landward of the trench-rise-trench triple junction.

Figure 5 shows the largest known events ($M > 5.6$) of this earthquake series, for the years 1960–1962. Aftershocks at the interface thrust zone are concentrated at the Mocha block. South of the Valdivia fracture zone, large aftershocks are remarkably lacking at the interface thrust zone but are concentrated near the Chile trench. The trench activity began about 8 hours after the main shock and continued for at least $1\frac{1}{2}$ years; none of these have magnitude greater than 7.0. B. Isacks (cited by *Sykes* [1971]) found normal-faulting focal mechanisms for many of these earthquakes near the Chile trench.

The magnetic anomalies and implied ages of the oceanic plate shown in Figure 5 are from *Herron* [1981] and *Herron et al.* [1981]. The age of the Mocha block at the Chile trench is about 25 m.y. B.P. [*Herron*, 1981]. North of the Mocha fracture zone, the age of the subducting plate is about 32 m.y. B.P. South of the Valdivia fracture zone, the age of the Nazca plate at the trench is about 20 m.y. B.P., and the subducting plate becomes progressively younger to the south, approaching the intersection of the Chile rise with the Chile trench [*Herron*, 1981; *Herron et al.*, 1981].

Approximations of F_{SP} acting along the rupture zone of the 1960 Chile earthquake are given in Figure 6. These estimates assume that the subducting plate is an effective stress guide and do not consider forces resisting F_{SP}; thus the shape of the resulting curve is more significant than its absolute value. Figure 6 shows that the slab pull force is greatest at and just south of the Mocha block. As one looks south from the Mocha block, the progressively younger oceanic lithosphere would lead to both a smaller slab pull force and a larger coupling, until encountering extremely young oceanic lithosphere at the southern end of the 1960 Chile rupture, which may not be strong enough to sustain high stresses. Moreover, near the Chile rise the plate interface may be decoupled, because the low resistance to plate bending of the Chile rise will permit the slab pull force of locally subducted plate to easily bend this plate downward. This southward lessening of the slab pull force for the 1960 Chile zone is consistent with *Herron*'s [1981] observations that as one goes from the Mocha block to the Chile rise, trench depth shallows and the trench free-air gravity anomaly lessens; it is also consistent with the observation of *Kadinsky-Cade and Isacks* [1983] that the interface zone south of the Mocha block is more horizontal and has greater downdip extent (until decoupled at the slab bend) than that north of the Mocha block.

It is concluded that the 1960 foreshocks and main shock nucleated at the zone of strongest slab pull and weakest coupling, at the Mocha block. The main shock rupture then propagated southward into the zone of weaker slab pull force but stronger coupling. It is observed that rupture of the Mocha block independently has a much faster repeat time than the more southern zone [*Nishenko*, 1985]. Periodic ruptures of the entire 1960 zone may be triggered by a subset of Mocha block ruptures.

SLAB PULL AND TENSION IN THE UPPER MANTLE

As assumption in this paper is the dominance of tensional stress in subducted plate at depths of 50–200 km. However, there exist significant stress variations at intermediate depths in subducted plates [*Isacks and Molnar*, 1971; *Isacks and Barazangi*, 1977; *Vassiliou et al.*, 1984; *Fujita and Kanamori*, 1981]. *Fujita and Kanamori* [1981] find that as a global average, only 65% of intermediate-depth earthquakes have downdip stress axes. In the depth range of 50–200 km, deviations from downdip tensional focal mechanisms occur at the upper layer of double seismic zones and at certain geographic regions [*Fujita and Kanamori*, 1981], particularly for slabs whose seismicity is nearly continuous to over 600 km depth [*Isacks and Molnar*, 1971; *Vassiliou et al.*, 1984].

Plate Motion Oblique to Island Arcs

At the Aleutian arc, for example, a westward component of relative plate motion is due to the strong ridge push and slab pull forces arising from very old lithosphere on the west side of the Pacific plate [*Gordon et al.*, 1978]. Because of the physical integrity of plate subducted at the Aleutian arc, the oblique motion of the Pacific plate is transmitted to intermediate depths. This causes plate distortion, as is shown by earthquake focal mechanisms that indicate strike-slip deformation or lateral extension [*Stauder*, 1968, 1972].

Double Seismic Zones

The phenomenon of double seismic zones, where two layers of seismicity separated by about 35 km occur at the depth range of 60–190 km, has been critically reviewed by *Fujita and Kanamori* [1981]. They conclude that the zone near the slab's upper surface is dominated by downdip compression, that the zone deeper in the plate is dominated by downdip tension, and that double zones are defined best in old and "rapidly" subducting slabs. The popular explanation of double seismic zones is that they are due to unbending of plate that originally was bent at the top 50 km of plate, especially at the trench and outer rise [*Engdahl and Scholz*, 1977; *Isacks and Barazangi*, 1977; *Kawakatsu*, 1986]. *Sleep* [1979] and *Fujita and Kanamori* [1981] argue that plate unbending is an unlikely explanation for double seismic zones, particularly since such zones are not a general feature of subduction zones, all of which have zones of bending. Moreover, bending stresses

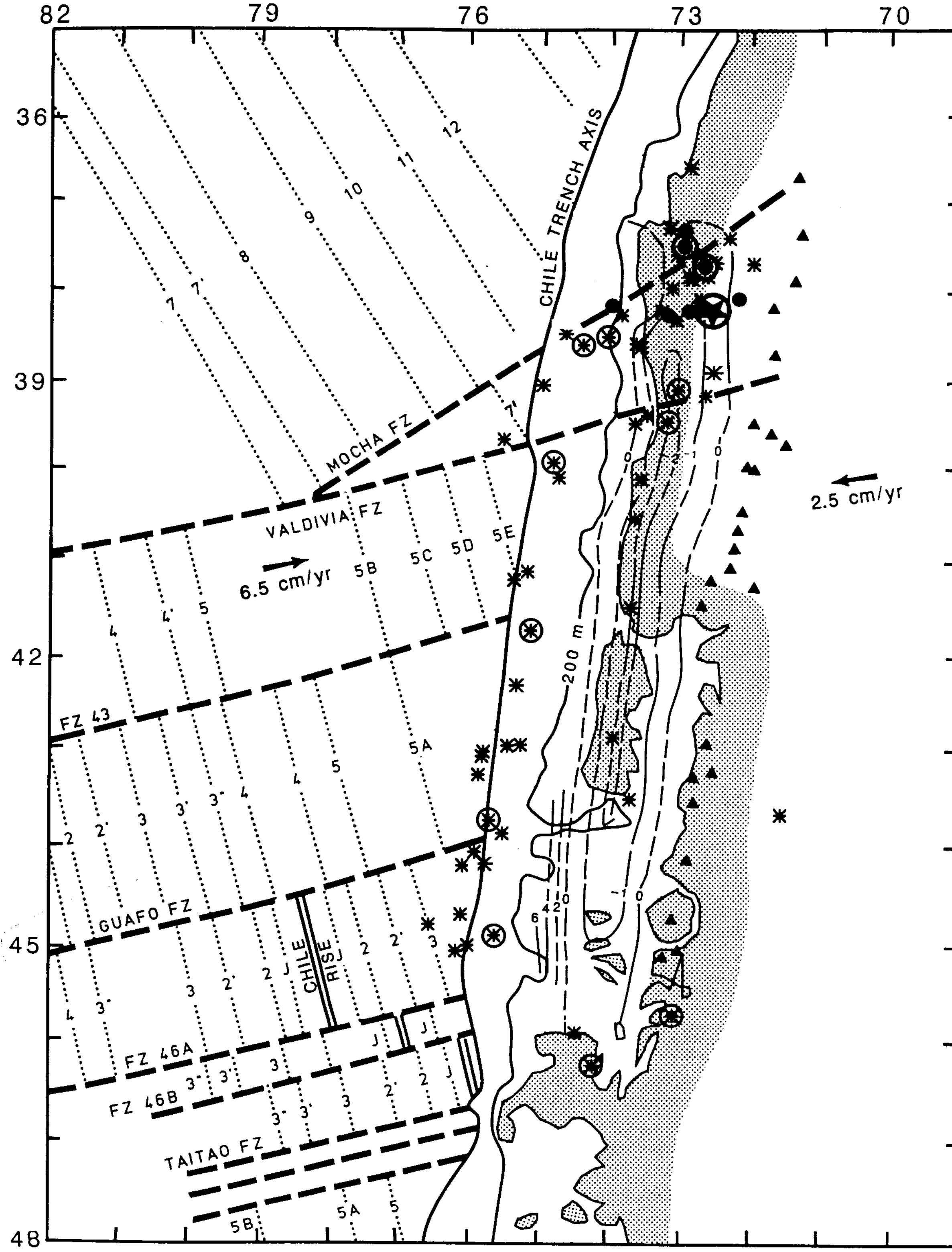

Fig. 5. Map showing foreshocks (solid circles), aftershocks (asterisks), the M_w 9.5 1960 Chile earthquake (circled star), associated vertical displacement in meters [*Plafker and Savage*, 1970], age of the oceanic lithosphere [*Herron*, 1981; *Herron et al.*, 1981], and volcanos (solid triangles). Earthquakes with $M > 6\frac{3}{4}$ are circled. Foreshocks, the main shock, and many aftershocks are concentrated near the Mocha block, defined by the zone between the landward extensions of the Mocha and Valdivia fracture zones. The rupture of the locked interface thrust zone allowed slab pull forces to migrate updip, producing high strains and normal-faulting earthquakes near the Chile trench. Earthquake locations were done on computer by the staff at the International Seismological Summary, and earthquake magnitudes are from the U.S. Coast and Geodetic Survey. Just north of the triple junction, earthquakes and volcanism occur downdip, implying the existence of plate to depths of at least 125 km. The apparent ease with which the Chile rise is subducted [*Herron et al.*, 1981; *Cande and Leslie*, 1986] suggests that rather than ridge push forces of the rise causing its own subduction, this subduction primarily is due to slab pull forces. Absolute plate velocities are from *Minster et al.* [1974].

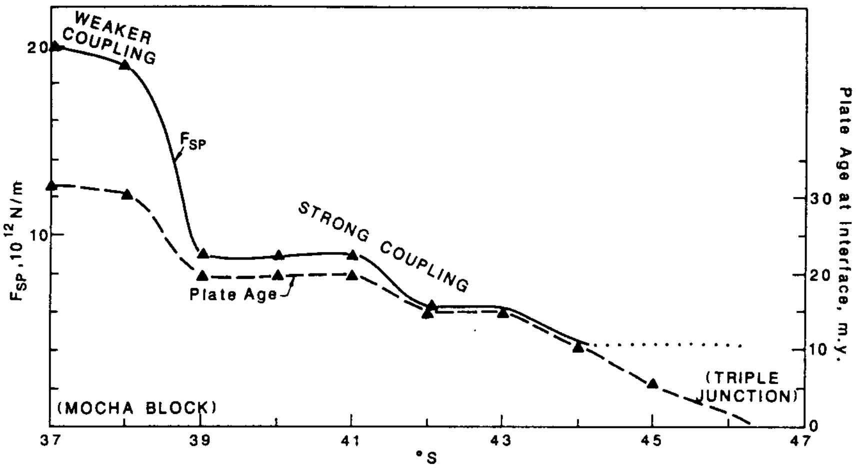

Fig. 6. Age of Nazca plate and estimates of F_{SP} acting at rupture zone of the 1960 Chile earthquake. F_{SP} estimates are taken parallel to the strike of the coast and are based on the relationships of *McKenzie* [1969] and *Davies* [1983]. The dotted line suggests slab pull force due to plate subducted beneath the triple junction.

should be concentrated at the slab bend zone, but the lack of earthquake activity there implies that the stresses due to that bending (as well as residual stresses from bending at the trench and outer rise) are dissipated locally at the slab bend. Possible sagging of slab between zones of relatively high viscosities is an alternative for the origin of double seismic zones [*Sleep*, 1979], but this hypothesis is inconsistent with double zones being best defined in rapidly subducting plates. In the slab pull model the tensional earthquakes directly reflect slab pull forces, whereas the compressional earthquakes reflect local resistance to rapid downdip plate motion. The cause of this compression is like that described earlier for slab pull stresses contributing to compressional stresses at an interface thrust zone.

Many intermediate-depth earthquakes that fail by other than downdip tension can be explained in ways such as that for the Aleutian arc or by their presence in the upper layer of double seismic zones. This leaves a predominating group of intermediate-depth earthquakes that fail by downdip tension, as is originally suggested by *Isacks and Molnar* [1971].

Precursors Downdip of Locked Interface Thrust Zones

The repeat time for large, shallow earthquakes along the Middle America Trench, offshore of Mexico, is 35–40 years [*Gonzalez et al.*, 1984; *Nishenko and Singh*, 1984]. Analysis of a complete catalog of smaller earthquakes in this zone showed an onset of normal-faulting earthquakes directly downdip from the interface thrust zone, beginning about 15 years prior to several large, shallow earthquakes [*Gonzalez et al.*, 1984; *McNally et al.*, 1986]. Other examples of large normal-faulting earthquakes downdip from interface thrust zones that preceded large gap-filling earthquakes have been reported for Chile [*Malgrange and Madariaga*, 1983] and Peru [*Dewey and Spence*, 1979; *Beck and Ruff*, 1984]. Such normal-faulting activity, beginning at the final $\frac{1}{2}$-$\frac{1}{3}$ of the repeat time and continuing to near the time of the main shock, would be expected as the slab pull force extends the subducted plate and increases tensional stresses at the plate's shallow regions.

Inversions of geodetic data from areas affected by great earthquakes occurring at Japanese subduction zones suggest slip at depths downdip from the interface thrust zones, before great earthquakes. In modeling these geodetic data, *Thatcher and Rundle* [1979, 1984] have considered a precursory, slowly acting slip downdip from the interface thrust zone.

The precursory normal-faulting earthquakes occurring just downdip of locked interface thrust zones are direct evidence for precursory tension downdip of great subduction earthquakes. These data are consistent with the influence of the slab pull force in loading stresses at a locked interface thrust zone and support the hypothesis that the slab pull force is an important contributor to stresses causing great subduction zone earthquakes.

STRESS-GUIDE EFFECTS

The downdip tension of typical earthquakes in the depth range of 50–200 km is interpreted here as due to the summed slab pull force of the deeper plate. In a deeply subducted plate, often there is a hiatus of earthquakes beginning at depths of 300 ± 50 km and extending 50 to 200 km downdip [*Isacks and Molnar*, 1971; *Abe and Kanamori*, 1979; *McGarr*, 1977; *Giardini and Woodhouse*, 1984; *Vassiliou et al.*, 1984]. With very few exceptions, earthquakes deeper than the hiatus show compressional failure, and a relative maximum in seismic energy release exists at depths of 500–650 km.

Mantle Resistance to Slab Pull

The usual explanation for the hiatus is that it reflects lowered deviatoric stresses [*Isacks and Molnar*, 1971; *Davies*, 1980; *Vassiliou et al.*, 1984; *Wortel*, 1986], although a localized weakening of subducted plate due to the exothermic olivine to β spinel phase change may be an alternate explanation [*Rubie*, 1984; *Wortel*, 1986]. Compressional stresses at depths beginning at 250–350 km may be caused by increased resistance to slab penetration due to a viscosity increase near these depths and to phase or chemical changes in the host mantle, beginning at depths of about 400 km and 650 km [*Isacks and Molnar*, 1971; *Christensen and Yuen*, 1984; *Anderson and Bass*, 1986]. *Vassiliou et al.* [1984] took the 650-km discontinuity to be a strong resister to plate penetration, without regard to the origins of plate motion, and successfully modeled much of the stress and seismicity of subducted plates.

In addition to the results of Vassiliou et al., the slab pull

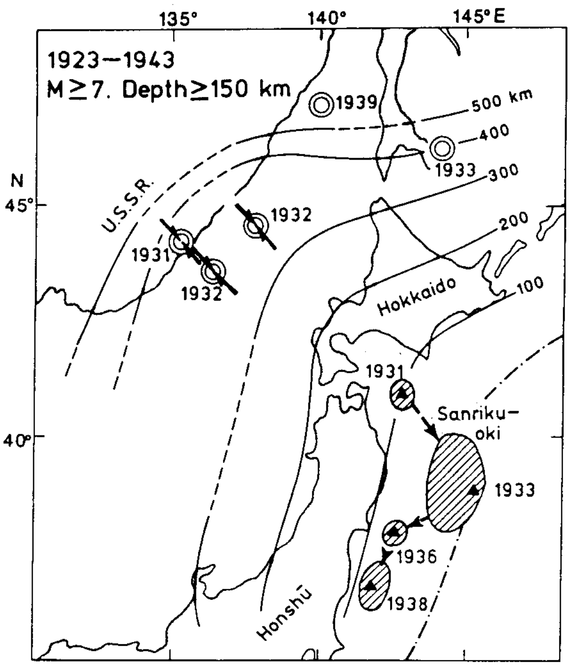

Fig. 7. Three $M > 7$, $h \approx 300$ km thrust earthquakes that preceded the great 1933 Sanriku normal-faulting earthquake [after *Mogi*, 1973]. Other shallow earthquakes of $M > 7.7$, whose occurrences migrate southward through the 1933 earthquake zone, are also shown by hatched areas, but these were thrust-faulting earthquakes.

force may contribute to the stress regime of the entire plate [*Davies*, 1980, 1983; *Wortel*, 1986]. Consider the stresses due to the slab pull force, as one moves downdip within the subducted slab, beginning just below the slab bend. At first the deviatoric stresses would be large and tensional because of the excess mass beneath (Figure 3). Moving deeper, there would be a diminished slab pull force, owing to diminished excess mass beneath and an increased downdip push from the increased excess mass above. Below a transition depth of low deviatoric stress, the deviatoric stress becomes increasingly compressional because of the weight of the plate above. Thus the factors of slab pull (plate weight) acting in a stress guide and resistance to plate motion at the discontinuities beginning at about 400 km and 650 km combine to explain the observed intraplate earthquake activity of tensional failure in the depth range of 50–200 km, the low seismicity at 300 ± 50 km, and the compressional failure at greater depths.

Mogi's Thrust Earthquakes at Depths of 260–350 km, Precursory to Great, Shallow Earthquakes

Mogi [1973] noted that within 3 years prior to the occurrence of the five great ($M_S > 7.8$) shallow earthquakes at and near northern Japan during the interval 1930–1968, there occurred one or more large ($M \geq 7$) earthquakes at depths of 260–350 km, directly downdip from the hypocenter of each pending great shallow earthquake. Where focal mechanisms are known, these intermediate-depth earthquakes exhibit thrust faulting. The great shallow earthquakes are typical interface thrust events, except for the great, normal-faulting 1933 Sanriku earthquake. Figure 7 shows the large intermediate-depth earthquakes that preceded the Sanriku earthquake. Mogi notes that all large intermediate-depth earthquakes occurring during this time-space interval were related empirically in this way to the great shallow earthquakes.

Take the shallow subduction zone to be locked and the subducted plate at 50 to 200 km depth to be under tension. Further plate sinking in the zone of lowest mantle viscosity will cause an increase in shallow tension. This episode of sinking plate will increase the stress in the deeper plate, because motion of deeper plate is resisted by the higher viscosity mantle there. The deeper compression, whose time history is describable by a Maxwell model, may cause the observed deeper compressional earthquakes (Figure 8*a*). Continued plate sinking would further increase shallow tensional stresses, eventually loading sufficient stress at the locked shallow zone to lead to the observed subduction zone earthquakes. The greatest extension would be immediately downdip from the most strongly coupled zone (the controlling asperity) for each great, shallow earthquake, and correspondingly, the increased compression would be immediately downdip from each main shock epicenter, consistent with the observed data. It is concluded that because subducted plate is a stress guide, temporal changes in tension in the upper part of subducted plate can induce stress changes (and occasional faulting) in the deeper subducted plate, if the period of plate sinking in the lowest viscosity part of the upper mantle is faster than the relaxation time of the deeper, resisting mantle.

Slab Pull and Postearthquake Strain Propagation

Trench, Normal-Faulting Earthquakes Following Decoupling, Interface Thrust Earthquakes

Normal-faulting earthquakes at oceanic trenches and outer rises have been observed for days to many months following decoupling interface thrust earthquakes [*Stauder*, 1968; *Sykes*, 1971; *Spence*, 1977; *Chapple and Forsyth*, 1979; *Hanks*, 1979; *Christensen and Ruff*, 1983]. These normal-faulting earthquakes nearly always have axes of least-compressive stress that trend downdip, regardless of the direction of local relative plate motion. The largest known such normal-faulting earthquake (M_S 7.5) was immediately trenchward of the epicenter of the 1965 Rat Island earthquake [*Spence*, 1977]. An excellent example of postearthquake normal faulting was shown for the 1960 Chile earthquake series (Figure 5). Such earthquakes are explainable by a trenchward migrating, tensional strain pulse (Figure 8*b*), whose amplitude is a function of the local seismic slip at the interface thrust zone and whose speed is controlled by the effective viscosity of the mantle. When this tensional strain reaches the trench and outer rise, it adds to existing bending stresses and causes the observed normal-faulting earthquakes.

A slab pull origin of this tensional strain is consistent with the finding that the shapes of outer rise and trench systems are best explained in the absence of large, horizontal compressional stresses [*Chapple and Forsyth*, 1979; *Caldwell et al.*, 1976; *Parsons and Molnar*, 1976; *Melosh and Raefsky*, 1980] but perhaps are due to bending moments originating at or below interface thrust zones. Such a temporally and regionally localized tensional strain would help even a weak ridge push force to produce a trenchward diffusion of oceanic plate [*Melosh*, 1976].

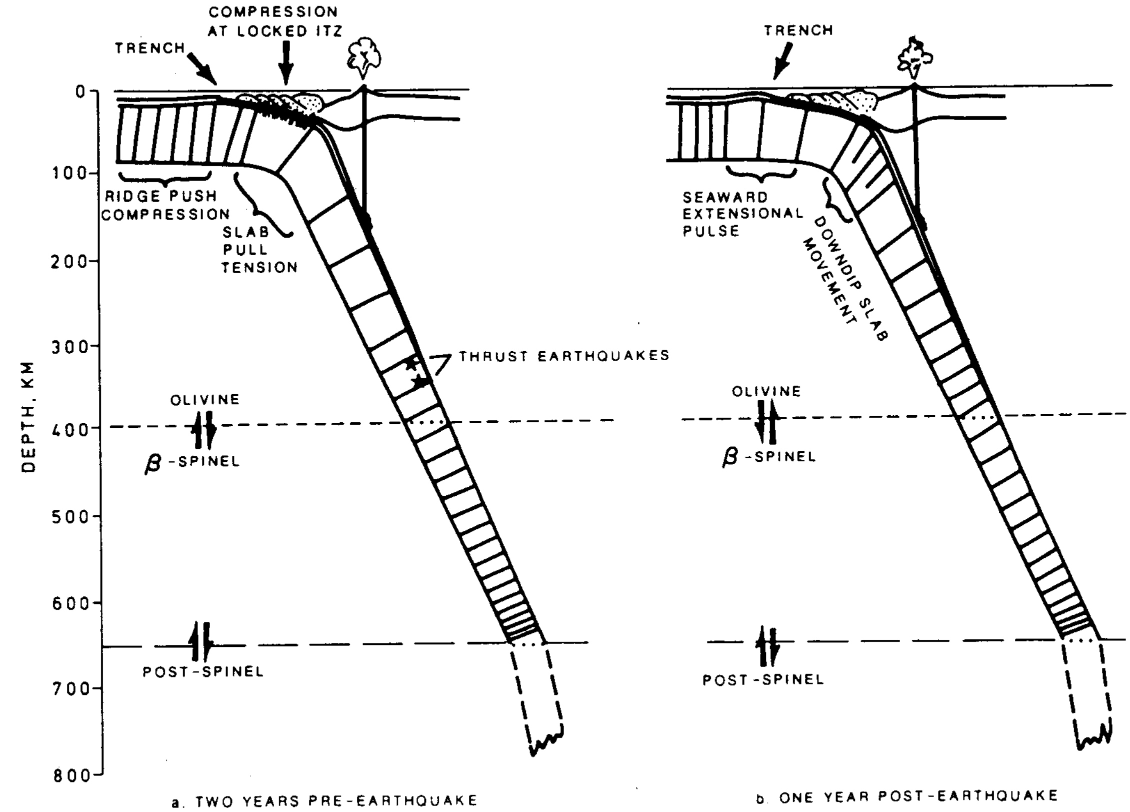

Fig. 8. Schematic diagram of temporal deviatoric stresses (*a*) 2 years before decoupling interface thrust zone earthquake and (*b*) 1 year after decoupling interface thrust zone earthquake. Relative extensional stress is indicated by widely spaced lines normal to the slab surface; relative compressional stress is indicated by closely spaced lines. Greatest tensional stresses are just beneath the interface thrust zone and prior to the decoupling earthquake; greatest compressional stresses are at the interface thrust zone. Entrained mantle flow is not indicated.

Strain Pulse Moving Downdip From a Decoupled Earthquake Zone

Following a decoupling earthquake, the stretched downdip slab responds like a stretched, damped spring after one end of the spring has abruptly been released (Figure 8*b*). Because the postearthquake, downdip propagating strain pulse is returning the plate stresses toward lithostatic values, brittle fracture would tend not to occur unless mantle resistance to plate motion causes high compressional stresses near the plate's upper surface, similar to the mechanism by which slab pull can lead to compression at a locked interface thrust zone or at the upper layer of double seismic zones. Evidence for such strain pulses includes earthquakes with downdip compression axes following the decoupling earthquakes of 1960 in Chile [*Astiz and Kanamori*, 1985], of 1965 at the Aleutian arc [*Spence*, 1977] and of 1977 in Indonesia [*Spence*, 1986]. Further evidence for downdip propagating postearthquake strain is provided by geodetic data, where *Savage* [1983] inferred rapid slip to a depth of near 100 km, well below the downdip end of the interface thrust zone. The forces resisting downdip plate movment will determine how much the tension in the upper 200 km of plate is diminished by such a downdip propagation of strain.

DISCUSSION

The Shumagin seismic gap, Alaska, has been the subject of much study [*Davies et al.*, 1981]. The updip part of the interface thrust zone at the Shumagin gap is seismically quiescent, and interface thrust earthquakes presently are only at the downdip end of the main thrust zone. Source modeling for two of the largest of these earthquakes indicate stress drops greater than 60 MPa, and these earthquakes have been interpreted as the breaking of two small but strongly loaded asperities [*House and Boatwright*, 1980]. Within the subducting plate and below the slab bend, at depths between 45 and 120 km, there are two discrete earthquake subzones that show extensional or compressional deformation. The sense of deformation within these two subzones reversed between 1979 and 1981 [*Reyners and Coles*, 1982; *Hauksson et al.*, 1984], correlative with a sudden change in surface tilt [*Beavan et al.*, 1984]. Interpretation of the seismicity and tilt data by *Beaven et al.* [1984] and *Hauksson et al.* [1984] indicates major stress changes downdip from the main interface thrust zone. If these phenomena are precursory to the large earthquake that some investigators expect to fill the Shumagin seismic gap, then it is implied that detailed seismic monitoring for precursors to other interface thrust earthquakes should include the regions just updip and downdip from the slab bend. *Savage et al.* [1986] suggest that the Shumagin seismic gap is not storing significant strain. In this case the preceding observations indicate processes of largely decoupled subduction and indicate that the slab pull force acting through the slab bend is important in such processes.

The possible association of excitations of the Chandler wobble with great earthquakes is well known, but only about 10% of the required excitation power can be accounted for by the global static deformation field of great earthquakes [*Dahlen*, 1973; *O'Connell and Dziewonski*, 1976; *Mansinha et* *al.*, 1979; *Gross*, 1986; *Chao*, 1985]. The plate motions described in this paper, which precede and follow decoupling earthquakes, may provide much of the missing excitation power for the Chandler wobble and changes in length of day.

Summary and Conclusions

This synthesis of subduction seismotectonics indicates that gravity expressed through the negative buoyancy of subducted lithosphere causes most earthquakes at shallow subduction zones and in subducted lithospheres. These earthquakes are caused by interactions between the forces that drive and resist plate motions. The slab pull force and the weaker ridge push force cause localized compression at interface thrust zones, because plate motion due to these forces is resisted there. The large slab pull force of very old plate tends to pull the subducting plate vertically away from the interface thrust zone and lead to low coupling, whereas the smaller slab pull force of younger plates allows the strength of subducting plate to resist bending and lead to strong coupling. The rupture of the great 1960 Chile earthquake began at an interface zone bounding oceanic plate <30 m.y. old, which is locally the oldest oceanic plate (implied relatively weak coupling); the rupture then propagated southward along an interface bounding increasingly younger oceanic lithosphere, terminating near the subducting Chile rise.

Subducted plate at mantle depths of 50–200 km shows pervasive normal faulting and occasional great earthquakes, due to pulling by the excess mass of deeper plate. In the slab pull model the extensional failure of earthquakes in the lower seismic layer of double seismic zones represents this usual pattern of seismicity, whereas the compressional failure of earthquakes in the upper seismic layer is interpreted as due to downdip plate motion being resisted by the bounding mantle. Deeper than about 200 km, because subducted plate acts as a stress guide, the push from excess mass above balances the pull from beneath, resulting in near-zero deviatoric stress and a hiatus of earthquake activity. Below this hiatus, earthquakes are caused by compression in the plate due to the push from excess mass above being resisted by the higher viscosity mantle phases beginning at depths of about 400 km and 650 km.

Prior to a great earthquake, stress accumulates at a locked interface thrust zone owing to the subducted plate's slow sinking and the oceanic plate's being pushed trenchward. The sinking plate will cause the greatest extension and occasional precursory normal-faulting earthquakes at shallow mantle depths. *Mogi* [1973] has noted large thrust earthquakes at depths of 260–350 km within 3 years of five great, shallow earthquakes at and near northern Japan. The sinking of the upper plate in the zone of lowest mantle viscosity, at 50–200 km depth, may load high stresses at the depths Mogi observed, because of the higher mantle resistance to plate motion at those depths.

A general feature of subducted plates at 20–40 km beneath the earth's surface is a 20°–50° increase in plate dip. This aseismic slab bend is interpreted as a pivot for the summed slab pull force of the deeper plate. Great thrust earthquakes nucleate at the subduction interface just above the slab bend, whereas below the slab bend, earthquakes occur within the subducted plate. Because the slab pull force promotes a slow seaward movement of the subducted plate, and because oceanic plate is subducting, the slab bend must be a continually developing feature. The dissipative deformation within the slab bend and the bending moment there may be important resistances to the slab pull force being delivered to an interface thrust zone.

A decoupling earthquake will release subducted plate that was extended by slab pull forces, and a subsequent strain pulse will propagate downdip, returning the subducted plate to a less stressed state. Because the slab pull force typically is greater than the ridge push force, a postearthquake tensional pulse will propagate trenchward. This tensional pulse leads to accelerated plate bending and occasional normal-faulting earthquakes at the trench and eases trenchward diffusion of the oceanic plate by the ridge push force. Following rehealing of a decoupling earthquakes's rupture zone, the earthquake and subduction cycles will repeat. The slab pull model indicates that the zones just updip and downdip of the slab bend should be examined for precursors to great subduction zone earthquakes.

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