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Intermediate-term Seismic Precursors for Some Coupled Subduction Zones

RENATA DMOWSKA[1] AND LUCIA C. LOVISON[2]

Abstract—The paper discusses model results and then reviews observational data concerning some aspects of the mechanics of mature seismic gaps in coupled subduction zones. The concern is with space- and time-varying stresses, as signalled by the presence and mechanisms of earthquakes in the outer-rise zones adjacent to main thrust areas of large subduction events, and down-dip from such areas, in the downgoing slab. Observations are shown to be consistent with the expectation that in mature seismic gaps, as a result of interplate boundary locking in presence of sustained gravitational driving forces, at least the deeper portions of the ocean plate in the outer-rise zones are under increased compression, and the downgoing slab is under increased tension. The observational data cover two cases of closed seismic gaps, namely the region of the Chilean Valparaiso earthquake of March 3, 1985, and the earthquake of October 4, 1983. Four other cases concern still to-be-closed gaps in northern Chile and along the coast of Guatemala, and also the Kurile Islands Trench gap and the northern New Hebrides gap. It is concluded that the intermediate-term precursor, consisting of a combination of compressional outer-rise earthquake(s) and tensional intermediate-depth, intra-plate events in the downgoing slab, which mechanically signals the latter part of the earthquake cycle, could be useful in evaluating the maturity, and hence great earthquake potential of a seismic gap.

Key words: Earthquake prediction; earthquake precursors; subduction zones.

Introduction

Recent observational and theoretical work on earthquake cycles in subduction zones (CHRISTENSEN and RUFF, 1983; SPENCE, 1987; RICE *et al.*, 1987; DMOWSKA and LOVISON, 1987; CHRISTENSEN and RUFF, 1987; DMOWSKA *et al.*, 1987; EGUCHI *et al.*, 1987) has advanced our understanding of the occurrence of certain seismic phenomena in relation to stress accumulation and release related to great underthrust events. In particular, it has been realized that temporal variations of stress, associated with earthquake cycles, occur in the subducting slab and, as well, in the area of the outer-rise, oceanward from the main zones of subduction earthquakes. These observations prompted us to model, in a simple way, stress transfer in coupled subduction

[1] Division of Applied Sciences and Department of Earth and Planetary Sciences, Harvard University, Cambridge, MA 02138.

[2] Division of Applied Sciences, Harvard University, Cambridge, MA 02138.

zones during the earthquake cycle (RICE *et al.*, 1987; DMOWSKA and LOVISON, 1987; DMOWSKA *et al.*, 1987). In the present work we propose to use information on the temporal variations in stress fields in the downgoing slab and in the outer-rise zones, provided by seismological observations, to assess the maturity of a seismic gap. That is, the occurrence of seismicity consistent with elevated stresses in both those zones is proposed as an intermediate-term precursor to some large subduction earthquakes. The proposed precursor consists of the combination of an outer-rise, compressional event and normal (tensional), intra-plate, intermediate-depth earthquake(s) down-dip from the zone of main thrust events. Before looking at observational data relating to this type of precursor, we will briefly discuss the very recent observational work on earthquake cycles in subduction zones and our own model results pertinent to the present work.

Observations of intermediate-depth, intra-plate normal earthquakes, occurring in the latter parts of the earthquake cycle down-dip from the zone of main thrust events (e.g., MALGRANGE *et al.*, 1981; ASTIZ and KANAMORI, 1986; MCNALLY *et al.*, 1986; DMOWSKA *et al.*, 1987) have been interpreted qualitatively by ASTIZ and KANAMORI (1986) as results of temporal variations in stress in the subducting slab (e.g., Figure 6, ASTIZ and KANAMORI, 1986), caused by the locking of the main thrust zone and sinking of slab. The observations suggested that in the latter part of the cycle the slab down-dip from the main thrust zone is under increased tension and such earthquakes themselves are an intermediate-term precursor of the approaching large subduction earthquake.

For the outer-rise areas, CHRISTENSEN and RUFF (1983, 1987) present a world-wide search for the occurrence and mechanisms of outer-rise events, to correlate the observations with the time of these events in the earthquake cycle in each particular area. They find that in uncoupled subduction zones only tensional outer-rise earthquakes occur, indicating the permanent dominance of the slab pull in these areas, and in strongly coupled subduction zones, both tensional and compressional outer-rise events occur, relating spatially and temporally to the distribution of large underthrusting earthquakes in each region. They observe, that in the latter case, tensional outer-rise events follow large underthrusting events as the subducting plate is temporarily in tension, due to the underthrusting motion. Compressional outer-rise events occur as compressional stress slowly accumulates oceanward of locked sections of the interplate boundary. CHRISTENSEN and RUFF (1987) find that in four cases compressional outer-rise earthquakes have been followed by large under-thrusting events which have occurred two, four, seven and nineteen years after the associated outer-rise event. The remaining occurrences of compressional outer-rise events that they searched are located in regions which are either known seismic gaps or in regions where the seismic potential is unknown. They conclude that the occurrence of compressional outer-rise events is useful for assessing the seismic potential of a region on an intermediate time scale.

EGUCHI *et al.* (1987a,b) recently reported results of seismicity observations in the

Zenisu Ridge, Japan, the outer-rise of the eastern Nankai Trough. The routine observations performed by the Japan Meteorological Agency, with minimum identified magnitude about 4, improved dramatically with the installation in 1979 of the ocean bottom seismometers, which lowered the detectability of earthquakes to magnitude 2. To clarify further the seismic features of the area, the ocean bottom seismometer array was deployed for a period of 25 days in mid-1985. The results of that last experiment showed no earthquakes in the upper layer of the plate. All recorded earthquakes were located with depths range 18 to 35 km, i.e., within the lower layer of the mechanically strong part of the lithosphere. EGUCHI *et al.* (1987a,b) speculate that large, horizontal, compressive stresses of regional origin have raised the neutral surface for the horizontal deviatoric stresses, and they associate these stresses with the strong interplate coupling between the overriding and underthrusting plates around the eastern Nankai Trough. The authors suggest that the neutral surface under the outer rise of strongly coupled subduction zones seems to become shallower prior to the occurrence of great interplate earthquakes, and deeper immediately afterwards.

Thus it is clear that the occurrence of compressional outer-rise events signals the presence of compressional stresses associated with the locking of the main thrust area, i.e., signals that the region entered the latter part of the earthquake cycle. This also means that compressional outer-rise events alone constitute a quite reliable intermediate-term precursor.

In the following section we discuss briefly our model results pertinent to the evaluation of pulsating stresses associated with the earthquake cycle in coupled subduction zones.

Stress Accumulation and Transfer in Coupled Subduction Zones During the Earthquake Cycle

Here we will discuss some implications of an elementary mechanical model of stress accumulation and transfer in coupled subduction zones during the earthquake cycle. The model has been presented in preliminary form by DMOWSKA *et al.* (1986) and subsequently extended by RICE *et al.* (1987) and DMOWSKA and LOVISON (1987), and recently discussed in depth in DMOWSKA *et al.* (1987). The model is severely simplified, but its results help focus intuitive ideas about the basic characteristics of coupled subduction zones mechanics.

The model encompasses the deformation of the oceanic plate before it submerges under the continental lithosphere, the subduction zone itself and the subducting slab, up to the depths of the order of 200–250 km (Figure 1), and consists of the elastic, gravitationally driven slab of the thickness H, having shear drag interactions with a Maxwellian viscoelastic mantle. The velocity of the subduction process, averaged over many earthquake cycles, equals V_{pl}. Great thrust earthquakes are

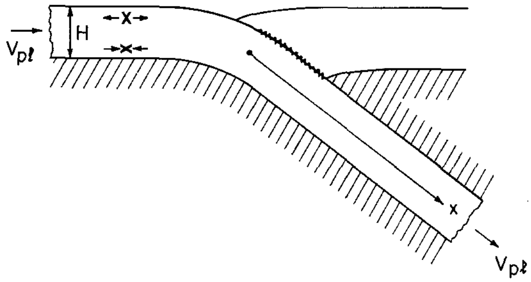

Figure 1
Subducting oceanic lithosphere with the thickness H; the velocity averaged over many earthquake cycles is V_{pl}.

simulated as periodic, sudden stress-relieving displacements at the contact zone between the slab and the overriding plate. The model treats the oceanic plate and subducting slab as a one-dimensional continuum, undergoing pulsating extensional and compressional deformation during the earthquake cycle, and having shear interactions with its surroundings, that are treated approximately by the Elsasser procedure. The model allows us to analyze the space- and time-variable stresses in the main thrust zone, in the adjacent oceanic plate, and in the subducting plate during each earthquake cycle. The modeling procedures are discussed in depth in DMOWSKA *et al.* (1987); here we present only the results concerning the areas of the outer-rise and, also, the regions of submerged slab, for depths starting with those just under the lower edge of rupture zones of large subduction earthquakes and extending down to 200–250 km depth.

It is important to recognize here that the model simulates only the cyclic part of the stresses, that is that part which pulsates over the earthquake cycle but averages in time to zero over a periodic cycle. Hence, comparison of stresses plotted for different locations along the slab should not be used to infer absolute stress levels. The magnitudes and spatial distributions of the "background" stresses, corresponding to the steady subduction of the slab (they also may be interpreted as the time-averages of stresses over several earthquake cycles), have not been determined in the model. However, the gravitational loading of subducting plates, as reviewed recently by SPENCE (1987), offers some characterization of these time averaged stresses. In particular, the presumed dominance of slab pull as a driving force suggests that steady-state stresses in the plate downdip from the thrust contact zone will correspond to an extensional regime. Also, since flexure of the plate is significant updip from the thrust contact zone (in the area of outer-rise earthquakes), we may expect that the steady-state stresses would have a strong bending component there, i.e., be extensional at shallow depth and compressional at greater depth in the plate. We will come back to that in the further discussion in this section.

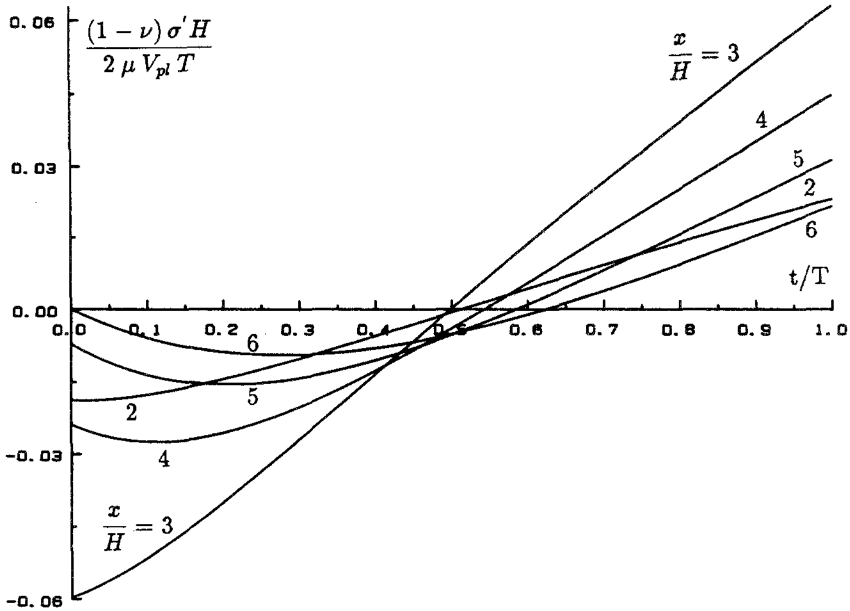

Figure 2
Extensional stress perturbation σ' in the subducting slab, downdip from the thrust contact, at x = 3H, 4H, 5H and 6H, and at location x = 2H beneath the thrust contact (from DMOWSKA *et al.*, 1987).

Figure 2 shows σ', i.e., the pulsating part of the extensional stresses, sampled at different locations along the axis x directed with the descending slab. The figure shows σ' in the dimensionless time during the whole earthquake cycle of the length T, H being the slab thickness. This figure presents stresses for locations between x = 3H (bottom edge of thrust contact) and x = 6H, that is downdip from the thrust contact zone, and also at a location beneath the thrust contact at x = 2H. The stress history here, and in the next figure, is shown on $0 < t/T < 1$ and repeats itself periodically on $1 < t/T < 2$, $2 < t/T < 3$, etc. This particular case is based on a viscoelastic relaxation time, entering the Elsasser model of shear coupling, of 0.14 T.

The most severe variations in stress occur, as expected, at the downdip edge of the thrust contact zone, at x = 3H, and the stress variations attenuate with distance downdip in the slab.

For the purposes of the present paper it is important to interpret Figure 2 in terms of seismicity. Let us recall that total stress would consist of the previously discussed time-independent mean stress (associated with an imagined steady subduction process) and the pulsating stress field, as visualized in Figure 2. Due to the dominance of slab pull in driving the subduction process, the time-independent part of the stress field will be of extensional character in the part of descending slab considered. Thus one expects the total stress to be extensional at all times. As shown in Figure 2, the magnitude of the pulsating stress (thus also the magnitude of total stress) is diminished highly by the occurrence of the large subduction earthquake (at t/T = 0), and, even, as shown by the curves x = 4H, 5H and 6H, continues to reduce within the descending slab for 15 to 30% of the next earthquake cycle. These results are consistent with the infrequent extensional seismicity in the slab during the first half or so of the cycle, whereas the increases of stress in the latter part of the cycle are consistent with the reactivation of extensional (normal) faulting in the slab. These results of our modeling, of which full discussion is presented in DMOWSKA *et al.* (1987), rationalize our expectations of occurrence of normal, intra-plate seismicity down-dip from the main thrust zones, as the cycle progresses. That is, the locking of the thrust zone and the continual sinking of the subducting plate make the presence of such normal earthquakes more probable in the latter parts of the earthquake cycle, with corresponding scarcity of such normal seismicity in the first parts of the cycle.

Now let us consider the outer-rise regions. Figure 3 shows the extensional pulsating stresses σ' updip from the thrust contact zone, in the region of oceanic plate extending from the trench (approximately at x = 0) over negative values of x towards the outer rise. As in the previous figure, the stresses shown are averaged over the slab thickness H.

In the first part of the cycle the pulsating stresses σ' are tensional, and they are compressional in the later parts of the cycle. The effect diminishes with distance and is most pronounced close to the trench. However, note that for locations closer to

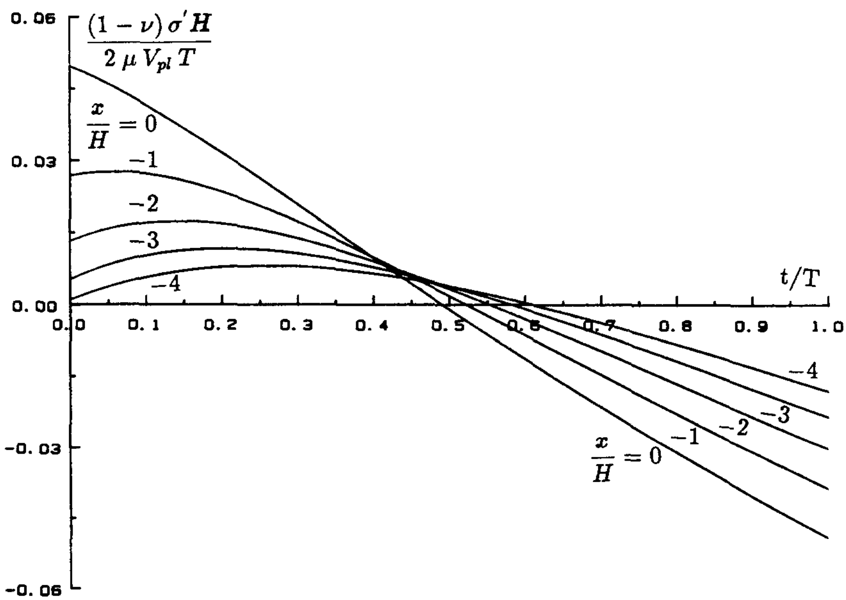

Figure 3
Extensional stress perturbation σ' updip from thrust contact, toward outer-rise; shown at x = 0, −2H, −3H and −4H (from DMOWSKA *et al.*, 1987).

the ocean (x = −H, x = −2H, x = −3H, etc.) the tensional stresses increase to 30% of the cycle, after the occurrence of the large subduction event at t/T = 0, suggesting a prolonged period of extensional seismicity. These model simulations are consistent with the seismicity observations by CHRISTENSEN and RUFF (1983, 1987), who note that many extensional outer-rise earthquakes follow large subduction events, with such events in Chile, in the area of the extremely large 1960 earthquake, continuing even until now.

As shown in Figure 3, the pulsating stresses σ' turn compressional in the latter part of the cycle, which is consistent with the absence of extensional outer-rise seismicity in the later portions of the cycle and with occasional compressive events occurring in that region of the ocean plate (CHRISTENSEN and RUFF, 1983, 1987).

The shapes of the curves in Figures 2 and 3 reflect a competition between stress diffusion from the great subduction earthquake in the beginning of the new cycle, and continued loading from ongoing subduction process. The former is important in the earlier portions of the cycle. It dominates the shape of the curves there, producing the local minima in σ' in Figure 2 and maxima in Figure 3. It is a negligible effect in the later portions of the cycle, where stress accumulation is dominated by slab sinking.

To interpret the total stresses in the areas of outer-rise, we should add our model simulations of the pulsating part of the stress field to the "background" stress field present in the oceanic lithosphere there. This is done schematically in Figure 4. The addition of time-variable pulsating stresses associated with earthquake cycle to the

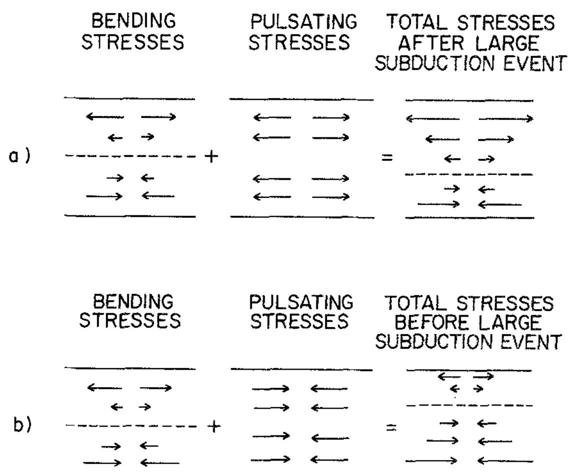

Figure 4
Schematic behavior of outer-rise stresses before and after large subduction event.

bending "background" stresses operating in the oceanic plate, and consisting of tensional ones in the upper regions of the plate and compressional ones in the lower regions, would produce the situations visualized on the right in Figure 4. Total stresses after a large subduction event would tend to be mainly tensional, with compressional ones diminished in magnitude and depressed to the deeper regions of the plate. The same total stress field would look totally different before the coming large subduction event, with compressional stresses dominating and tensional ones diminished in magnitude, and being pushed to the most shallow areas of oceanic plate, if existing at all. Also the neutral surface, separating the areas of different direction of the stress field, would be rather deep after the subduction event and very shallow towards the end of the cycle, before the next big earthquake. This picture constitutes a rationale for seismicity observations of outer-rise events, as presented by CHRISTENSEN and RUFF (1983, 1987) and EGUCHI *et al.* (1987a,b).

Observational Data

For earthquake locations, we used the International Seismological Center catalogue from 1964 to 1984, and the PDE (Preliminary Determination of Epicenters) catalogue from USGS from January 1985 to October 1986. We examined seismic events with magnitude $m_b \geqslant 4.0$.

We used the centroid moment tensor solutions of Harvard University, which are available from digital data for the period 1977 to 1987. We integrated the former data with focal mechanism solutions prior to 1977, drawn from the geophysical literature cited in each regional section. Each focal mechanism diagram corresponds to an equal area projection of the lower focal hemisphere, with the shaded areas as quadrants of compressional first motion, that is, regions of extension at the source; and white quadrants regions of shortening at the source. The dot centered in the white quadrant indicates the direction of the P-axis.

Since our purpose was to identify possible time-space variations of the type of seismicity trenchward and downdip from the thrust contact zone, we reported the focal mechanisms of the earthquakes located in these areas, and the focal mechanisms of the main thrust events, only if these occurred after 1977.

All earthquakes considered in this paper are listed in Table 1.

Central Chile: Valparaiso

Figure 5 presents the sequence of events for the region between 31.5°S and 35°S along the Central Chilean trench, as it started rerupturing the area of the large

Table 1

List of all Earthquakes Considered, Chronologically by Region.

Date	Longitude (deg)	Latitude (deg)	m_b	M_s	Depth (km)	Mech.	Ref.
		Central Chile: Valparaiso					
Aug. 16, 1906	72W	33S		8.2*–8.6		C	K72
April 6, 1943	71.27W	30.98S		*		C	K72
Mar. 28, 1965	71.10W	32.42S	6.4	8.3*	68	T	ISC/S73
Nov. 10, 1966	68.40W	31.95S	5.9		110	T	ISC/S73
Sept. 26, 1967	70.51W	33.47S	5.7		78	T	ISC/S73
July 9, 1971	71.20W	32.51S	6.5		40	C	ISC/M
Jan. 20, 1978	70.38W	34.29S	5.8	7.5	112	T	ISC/HAR
July 13, 1980	70.18W	33.46S	5.6		106	T	ISC/HAR
Oct. 16, 1981	73.10W	33.13S	6.2		18	C	ISC/HAR
Nov. 7, 1981	71.43W	32.17S	6.2	7.2	57	T	ISC/HAR
Feb. 25, 1982	73.25W	33.24S	5.0		27	C	ISC/KM
Dec. 15, 1983	70.15W	33.09S	5.9		103	T	ISC/HAR
May 9, 1984	70.27W	34.17S	5.5		113	T	ISC/HAR
Mar. 3, 1985	71.87W	33.13S	6.7	7.8	33	C	ISC/HAR
		Northern Chile: Copiapo					
Nov. 11, 1922	70W	28.50S		8.4*		C	K72
Aug. 18, 1964	71.78W	26.37S	6.1		32	C	ISC/CR83
Dec. 28, 1966	70.74W	25.51S	6.6		23	C	ISC
June 28, 1978	70.15W	27.56S	5.2		84	T	ISC/HAR
Aug. 3, 1978	70.66W	26.52S	6.3		49	T	ISC/HAR
Oct. 4, 1983	70.62W	26.55S	6.3	7.4	5	C	ISC/HAR
		Northern Chile: Coquimbo					
Nov. 11, 1922	70W	28.50S		8.4*		C	K72
July 12, 1965	68.25W	28.41S	5.6		108	T	ISC/S73
Nov. 13, 1969	71.67W	27.76S	5.7		43	C	ISC/S73
Mar. 15, 1970	69.42W	29.60S	5.8		99	T	ISC/S73
June 27, 1979	68.36W	29.51S	5.5		103	T	ISC/HAR
July 19, 1979	69.69W	29.00S	6.0		97	T	ISC/HAR
Oct. 26, 1982	71.37W	29.70S	5.6		63	T	ISC/HAR
Oct. 21, 1983	69.17W	30.70S	5.5		107	T	ISC/HAR
		Middle America: Southern Mexico–Guatemala					
Aug. 6, 1942	90.93W	13.90N		8.3*		C	K73
Oct. 23, 1950	91.82W	14.34N		7.1*		C	K73
April 29, 1970	92.59W	14.65N	5.8	7.3	56	C	ISC/LM85
Aug. 20, 1971	92.41W	13.30N	5.8	5.6	9	C	ISC/F
Feb. 22, 1978	91.46W	14.23N	5.6		93	T	ISC/HAR
Aug. 18, 1978	91.49W	14.31N	5.5		100	T	ISC/HAR
Sept. 10, 1978	91.58W	14.25N	5.6		87	T	ISC/HAR
Jan. 12, 1979	91.60W	14.32N	5.4		89	T	ISC/HAR
Aug. 17, 1981	93.78W	14.43N	5.6		23	T	ISC/HAR
Oct. 31, 1984	90.37W	14.06N	5.3		111	T	ISC/HAR
Sept. 15, 1983	93.24W	16.06N	5.7		116	T	ISC/HAR
Aug. 31, 1984	93.32W	16.06N	5.3		121	S–T	ISC/HAR
July 5, 1986	92.58W	15.47N	5.4		112	T	ISC/HAR

Date	Longitude (deg)	Latitude (deg)	m_b	M_s	Depth (km)	Mech.	Ref.
		Kurile Islands and Kamchatka					
May 1, 1915	155E	47N		$8.1^1/7.9^2$		C–S	R58/K74
Nov. 4, 1952	159.50E	52.60N			53	T	K76
Sept 15, 1962	157.11E	48.48N	6.5	8.25	3	C	UB
Mar. 16, 1963	154.83E	46.79N			0	C	SB66/CR82
Oct. 13, 1963	149.50E	44.80N		7.7*	30	C	SR87/K70
Aug. 4, 1964	151.36E	46.57N	5.7	8.5	86	T	ISC/SM
Dec. 1, 1967	154.40E	49.45N	5.9		144	C	ISC/SM
Mar. 3, 1971	153.07E	48.23N	5.7		127	C	ISC/SM
Dec. 2, 1971	153.33E	44.77N	6.2		38	C	ISC/SM
Mar. 22, 1972	153.60E	49.05N	6.2		135	T	ISC/SM
Mar. 25, 1972	153.13E	48.04N	5.8		124	T	ISC/SM
May 12, 1977	155.03E	50.12N	5.2		129	C–S	ISC/HAR
June 19, 1977	151.09E	47.12N	5.6		154	T	ISC/HAR
Dec. 20, 1977	153.06E	48.57N	5.7		153	C–S	ISC/HAR
Apr. 20, 1978	151.91E	46.65N	5.4		92	T	ISC/HAR
July 10, 1978	150.12E	48.76N	5.1		341	C	ISC/HAR
Sept. 3, 1978	154.41E	49.42N	5.5		114	T	ISC/HAR
May 23, 1979	159.69E	50.29N	5.6		3	ST	ISC/HAR
July 20, 1979	152.36E	47.41N	5.2		134	C	ISC/HAR
Aug. 23, 1981	157.37E	48.71N	6.0		2	T–S	ISC/HAR
Sept. 26, 1982	158.63E	50.10N	5.5		44	T	ISC/HAR
Sept. 26, 1982	152.21E	47.04N	5.6		113	C	ISC/HAR
Aug. 2, 1983	153.36E	45.05N	5.5		52	S–T	ISC/HAR
Aug. 28, 1983	151.53E	46.13N	5.5		53	T	ISC/HAR
May 15, 1984	153.79E	50.92N	4.8		281	T	ISC/HAR
Aug. 22, 1984	153.45E	49.04N	5.1		148	C	ISC/HAR
Apr. 10, 1985	159.42E	49.99N	5.5		28	T	ISC/HAR
May 30. 1985	154.09E	49.13N	5.5	4.9	150	T	ISC/HAR
Feb. 19, 1986	153.42E	48.58N	5.3		115	T	ISC/HAR
July 19, 1986	151.13E	47.26N	5.9		141		ISC/HAR
		Northern New Hebrides					
Jan. 22, 1964	165.96E	13.64S	6.0		50	T	ISC/JM
June 13, 1966	167.02E	12.23S	6.0		242	C	ISC/IM
Dec. 31, 1966	166.38E	11.89S	5.5	7.8^3	73		ISC/H
Aug. 11, 1970	166.56E	14.13S	6.1	7.5	20		ISC/H
Dec. 28, 1973	166.80E	14.56S	6.3	7.8	12		ISC/H
Sept. 26, 1978	166.89E	13.45S	5.6		204	C	ISC/HAR
June 19, 1979	167.19E	14.51S	5.7		126	T	ISC/HAR
July 8, 1980	166.37E	12.49S	5.9	7.8	29	C	ISC/HAR
July 17, 1980	166.04E	11.77S	5.7		33	C	ISC/HAR
Dec. 3, 1982	167.19E	13.32S	5.5		261	C	ISC/HAR
Dec. 28, 1983	166.77E	13.06S	5.5		114	T	ISC/HAR
Oct. 21, 1985	166.00E	13.60S	5.5	5.2	33	C	ISC/HAR
Dec. 21, 1985	166.52E	13.97S	6.0		43	C	ISC/HAR

References for Table 1 are: [1] magnitude M in RICHTER, 1985; [2] magnitude M in KELLEHER *et al.*, 1974; [3] magnitude in HABERMANN, 1984; * magnitude M from KELLEHER, 1972; CR83–CHRISTENSEN and RUFF, 1983; CR87–CHRISTENSEN and RUFF, 1987; F–FORSYTH, 1982; HAR–centroid moment tensor solution by the Harvard group; IM–ISACKS and MOLNAR, 1971; ISC–International Seismological Center Bulletin; JM–JOHNSON and MOLNAR, 1972; K70–KANAMORI, 1970; K73–KELLEHER *et al.*, 1973; K76–KANAMORI, 1976; KM–KORRAT and MADARIAGA, 1986; LM85–LEFEVRE and MCNALLY, 1985; M–MALGRANGE *et al.*, 1981; R58–RICHTER, 1958; S73–STAUDER, 1973; SB66–STAUDER and BOLLINGER, 1966; SM–STAUDER and MUALCHIN, 1976; SR87–SCHWARTZ and RUFF, 1987; UB–UDIAS and BAUMANN, 1969; T–tensional, C–compressional and S–strike-slip focal mechanism solutions.

August 16, 1906 (M_W = 8.2–8.6) Valparaiso earthquake. The extent of the 1906 event is marked in the right part of the figure.

The first event to rerupture part of 1906 quake was the July 9, 1971 (M_w = 7.5, m_b = 6.5) Valparaiso earthquake (e.g., MALGRANGE *et al.*, 1981; NISHENKO, 1985; KORRAT and MADARIAGA, 1986; CHRISTENSEN and RUFF, 1986; the aftershock zone shown in Figure 5). It was preceded by the normal events of March 28, 1965 (M_s = 7.5, m_b = 6.4) and September 26, 1967 (m_b = 6.7) located in the descending slab at depths of 72 km and 78 km respectively, beneath the epicenter of the July 9, 1971 underthrust event (MALGRANGE *et al.*, 1981) and signaling the tensional stresses present at that time and location. The slip of the July 9, 1971 underthrusting earthquake is estimated to be 1.6–5.7 m (MALGRANGE *et al.*, 1981), a poorly constrained value caused by uncertainties in the aftershock area. We suggest that the slip was smaller because the downgoing slab shows tensional stresses downdip from that zone in mid-1980's (earthquakes of July 13, 1980, m_b = 5.6, depth 106 km, and December 15, 1983, m_b = 5.9, depth 103 km; see Figure 5). This suggestion is also based on the fact, that this area has been partially reruptured in the March 3, 1985 Valparaiso earthquake (CHRISTENSEN and RUFF, 1986; ASTIZ and KANAMORI, 1986;

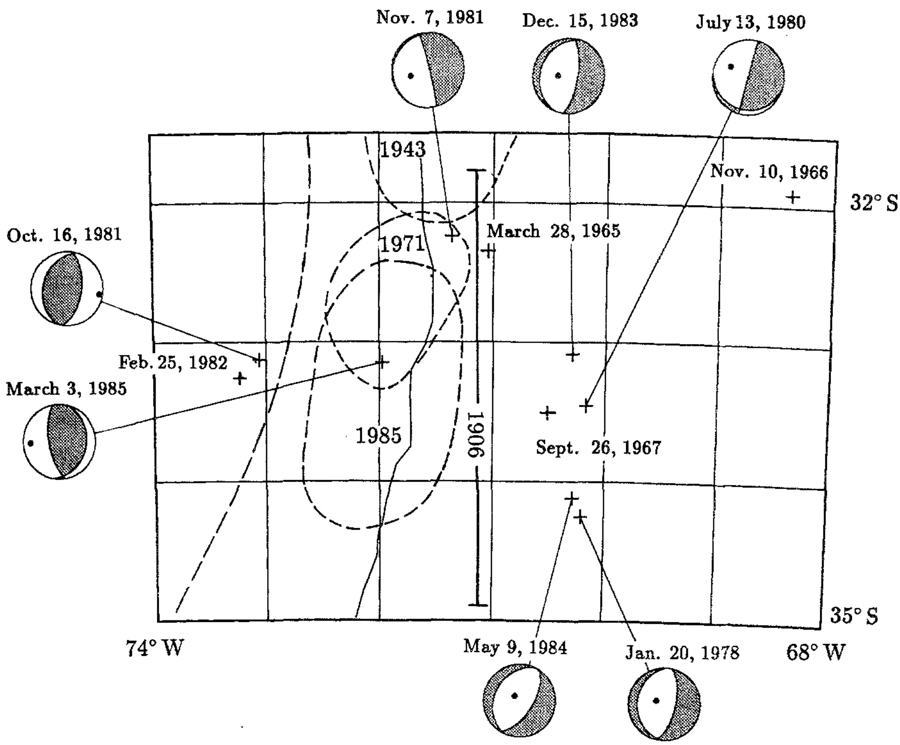

Figure 5
Sequence of events for the Central Chile Valparaiso earthquake of March 3, 1985. Aftershock areas of significant underthrust events are adapted from KELLEHER (1972) and CHRISTENSEN and RUFF (1986).

KORRAT and MADARIAGA, 1986), which was the second large underthrust quake rerupturing the 1906 earthquake zone and occurring in a place recognized as a mature seismic gap with a strong potential for large or great underthrusting event, based on historic seismicity (KELLEHER, 1972; MCCANN *et al.*, 1979; NISHENKO, 1985). In retrospect, it is possible to note signs of stress build-up in that area, namely the occurrence of compressional outer-rise events of October 16, 1981 ($M_s = 7.2$, CHRISTENSEN and RUFF, 1983; KORRAT and MADARIAGA, 1986) and February 25, 1982 ($m_b = 5.2$, KORRAT and MADARIAGA, 1986) suggesting the increased level of compressional stresses caused by the locking of the underthrust area, and, down-dip from that area, in the downgoing slab, the build-up of tensional stresses appearing in the form of intermediate-depth normal earthquakes (two previously mentioned, closer to the down-dip area of 1971 event, and two others, of January 20, 1978 ($m_b =$ 5.8, depth 112 km) and May 9, 1984 ($m_b = 5.5$, depth 113 km), all of them shown in Figure 5).

It is interesting to note one more intermediate-depth normal event present in this area, namely the earthquake of November 7, 1981 ($m_b = 6.2$, depth 57 km, Figure 5). ASTIZ and KANAMORI (1986) associate its presence with the build-up of tensional stresses in the downgoing slab in the area of the 1985 Valparaiso event, and, possibly, this is the case. However, its presence might be associated with the zone of April 6, 1943 Illapel earthquake ($M_s = 7.9$), just north of the 1971 event (southern end of aftershock zone shown in Figure 5). The 1943 event has poorly determined ends; KELLEHER (1972) estimates that the rupture extended from 30.2° to 32.2°S. The previous earthquake there was in 1880 and NISHENKO (1985) suggests that a 63-year repeat time appears to be reasonable for that area. If so, the zone of 1943 Illapel event is at present in its latter part of the cycle and the event of November 7, 1981 might be reinterpreted as the sign of the build-up of tensional stresses down-dip in the downgoing slab.

Northern Chile: Copiapo-Coquimbo

Part of northern Chile from 25.0° to 28.0°S is shown in Figure 6, with a sequence of events associated with the moderate underthrust earthquake of October 4, 1983 ($M_s = 7.4$), of which the aftershock zone is shown in the figure. The earthquake reruptured the northern end of the zone of the large November 11, 1922 Atacama Desert earthquake ($M_w = 8.5$), a complicated multiple event with a poorly known rupture zone ranging however from about 300 to 450 km (based on the extent of the coastal uplift, KELLEHER, 1972) and extending south from the area of the 1983 earthquake. The aftershock zone of the 1983 event touches to the north the area ruptured in the 1966 Taltal earthquake (estimated slip 200 cm, NISHENKO, 1985; Figure 6).

The 1983 event has been preceded by the compressional outer-rise earthquake of

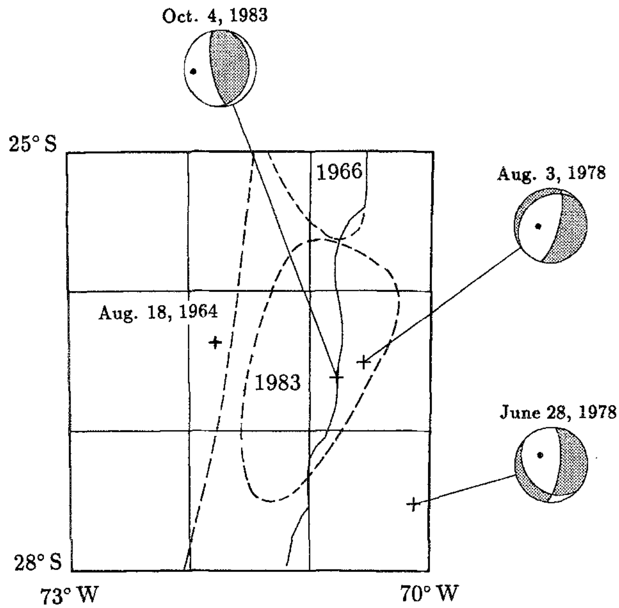

Figure 6
Sequence of events for the Northern Chile earthquake of October 4, 1983.

August 18, 1964 (m_b = 6.4, CHRISTENSEN and RUFF, 1983), showing the build-up of compressional stresses oceanward from the underthrust zone. We searched for signs of the build-up of tensional stresses down-dip from that zone, and, indeed, we found two normal events located in that area and preceding the 1983 event, namely the earthquake of June 28, 1978 (m_b = 5.2, depth 84 km) and the August 3, 1978 Copiapo earthquake (m_b = 6.3, depth 49 km, estimated average displacement 70 cm), located just below the lower edge of the thrust zone. Thus here we observe one more case of simultaneous occurrence of compressional outer-rise earthquakes and tensional, intermediate depth intraplate events, before the underthrust subduction event.

Figure 7 presents part of the northern Chile coast from 26.5° to 31.0°S, and is the southern continuation of the region shown in Figure 6, with part of aftershock zone of 1983 earthquake in the upper part of the figure. The previous large underthrust earthquake in this area was the November 11, 1922 Atacama Desert (M_w = 8.5) earthquake, of which the southern end extended to approximately 30°S. The previous seismic history in that area is poorly documented (LOMNITZ, 1970, NISHENKO, 1985). However, the region may have the potential for a large earthquake (MCCANN *et al.*, 1978, 1979). NISHENKO (1985) estimates the average recurrence time to be 104 years. As we discussed previously, the northern part of 1922 aftershock zone reruptured in the moderate earthquake in 1983 (Figure 6).

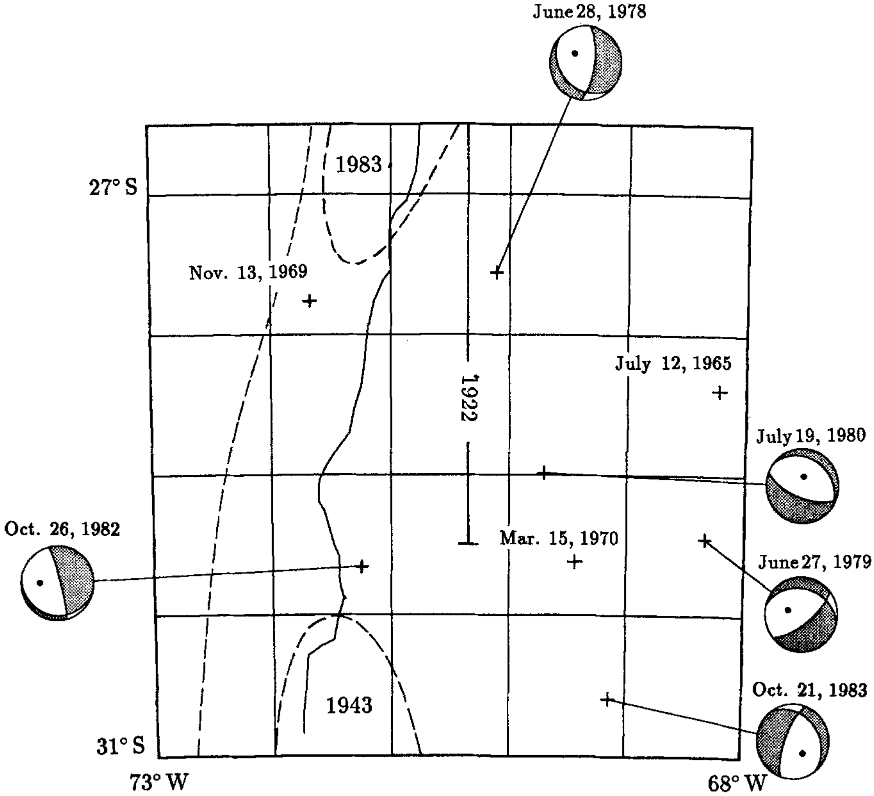

Figure 7
Sequence of events for Copiapo-Coquimbo region in Northern Chile. Aftershock area of the 1943 earthquake is adapted from KELLEHER (1972).

CHRISTENSEN and RUFF (1987) note the sign of higher compressional stresses in the oceanward area of that zone, appearing in the compressional outer-rise event of November 13, 1969 (m_b = 5.7, Figure 7). We searched the zone for possible signs of tensional stresses in the downgoing slab and found three tensional, intermediate-depth earthquakes of June 27, 1979 (m_b = 5.5, depth 103 km), July 19, 1980 (m_b = 6.0, depth 97 km) and October 21, 1983 (m_b = 5.5, depth 107 km). (Figure 7). A fourth one, of June 28, 1978 (m_b = 5.2, depth 84 km), lies down-dip from the southern end of the zone ruptured in 1983, and its appearance might be associated as well with that earthquake.

We agree with CHRISTENSEN and RUFF (1987) that the compressional outer-rise event of November 13, 1969 points to a future occurrence of a large subduction earthquake in that area. Also, the evidence we have presented of the build-up of tensional stresses in the downgoing slab strengthens the evidence that the thrust zone is at present locked and that the subduction earthquake is unavoidable.

Unfortunately, as we see from other coupled subduction zones where the compressional outer-rise and tensional intermediate-depth intraplate earthquakes were followed by a great subduction earthquake closing the cycle, their presence is a good intermediate-term precursor, but it does not suggest with any precision when the subduction event will occur.

Middle America: Southern Mexico-Guatemala

Figure 8 shows the southern end of Mexico and coast of Guatemala, with the Middle American Trench marked. Also we have marked the aftershock zones of recent large subduction events there, earthquakes of August 6, 1942 ($M_s = 7.9$), October 23, 1950 ($M_s = 7.1$) and the Chiapas event of April 29, 1970 ($M_s = 7.3$). The recurrence of earthquakes in that area has been studied by McNally and Minster (1981) and the length of cycle is 40 years for the 1942 quake area and 34 years for the 1950 earthquake zone. McNally (1981) and McNally and Minster

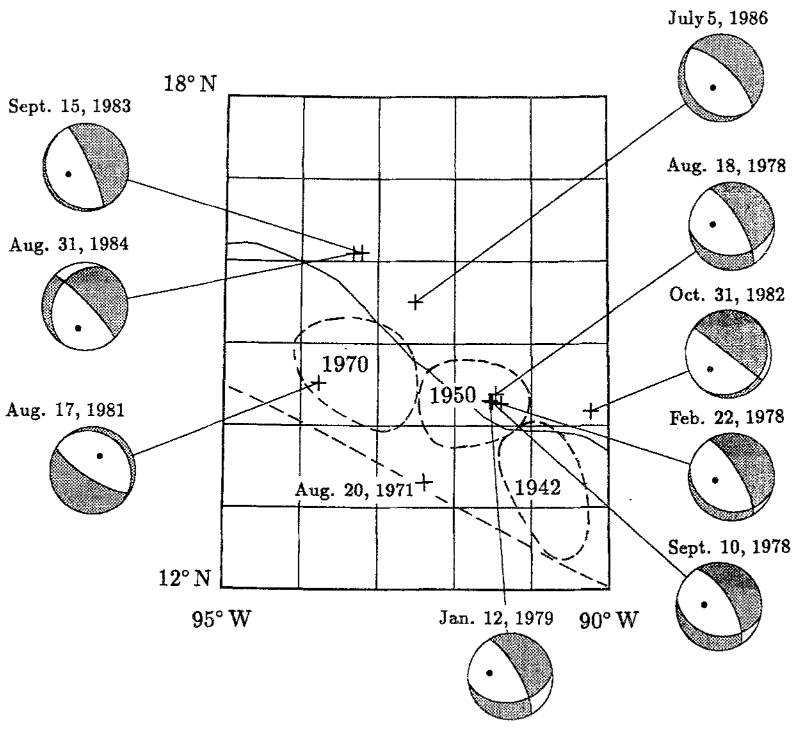

Figure 8
Sequence of events for the region of Southern Mexico and the coast of Guatemala. Aftershock areas are adapted from Kelleher *et al.* (1973).

(1981) suggest that the coast of Guatemala, between 90° and 92°W, is a mature seismic gap, and large shallow earthquake with ($M_s \geqslant 7.5$) could be expected there.

CHRISTENSEN and RUFF (1987) note two outer-rise events in this zone: a tensional event of August 17, 1981, which they associate with 1970 Chiapas earthquake, though it is difficult to judge that, as it occurred 11 years after the Chiapas event, and a compressional outer-rise earthquake of August 20, 1971, located oceanward from the junction of aftershock zones of 1950 and 1942 events. They suggest that this last earthquake may be a sign of the latter part of the earthquake cycle in that area and return of compressional stresses to the outer-rise zone, caused by locking of the interplate boundary.

We searched the down-dip area for intermediate-depth, intra-slab events, in hope of finding clues to the present state of stress there; our findings are presented in Figure 8. First, there is a series of events down-dip from southern end of aftershock zone of the 1950 earthquake, namely earthquakes of February 22, 1978 (m_b = 5.6, depth 93 km), August 18, 1978 (m_b = 5.5, depth 100 km), September 10, 1978 (m_b = 5.6, depth 87 km) and January 12, 1979 (m_b = 5.4, depth 89 km), all of which show tension in the slab, with the axis perpendicular to the trench. Also there is another normal event down-dip from the aftershock zone of the 1942 earthquake, which occurred on October 31, 1982 (m_b = 5.3, depth 111 km), further suggesting that the slab there is in tension. All these are in accord with MCNALLY (1981), MCNALLY and MINSTER (1981) and CHRISTENSEN and RUFF (1987) that the Guatemala coast between 90° and 92°W is indeed a mature seismic gap, and a large subduction event there is unavoidable.

Kurile Islands and Kamchatka

Figure 9 shows the Kurile Islands Trench Gap, which extends along the Kurile Trench from 46° to 49°N and is flanked by the aftershock zones of November 4, 1952 (M_w = 9.0) Kamchatka earthquake from the north-east and October 13, 1963 (M_w = 8.5) Kurile Islands event in the south-west. The previous earthquake in the gap occurred on May 1, 1915 (M = 7.9, KELLEHER *et al.*, 1974), however the rupture zone of this event was estimated by FEDOTOV (1965) to extend only about 100 km. Thus no record exists of an extensive rupture in the area of the gap, and, perhaps, this is unlikely, as the zone of shallow earthquakes there is unusually narrow (AVER'YANOVA, 1968; KELLEHER *et al.*, 1974), which suggests that the width of interface, that is the area where great subduction earthquakes occur, is also narrow. This narrowing of the interface occurs only along the gap segment of the plate margin; the great Kamchatka earthquake of 1952 (with the aftershock zone extending approximately 200 by 500 km) occurred along a segment of the plate margin that appears to have an unusually broad interface. Also the width of the interface broadens toward the southern Kuriles, where maximum dimensions of ruptures are larger (about 150–250 km, KELLEHER *et al.*, 1974).

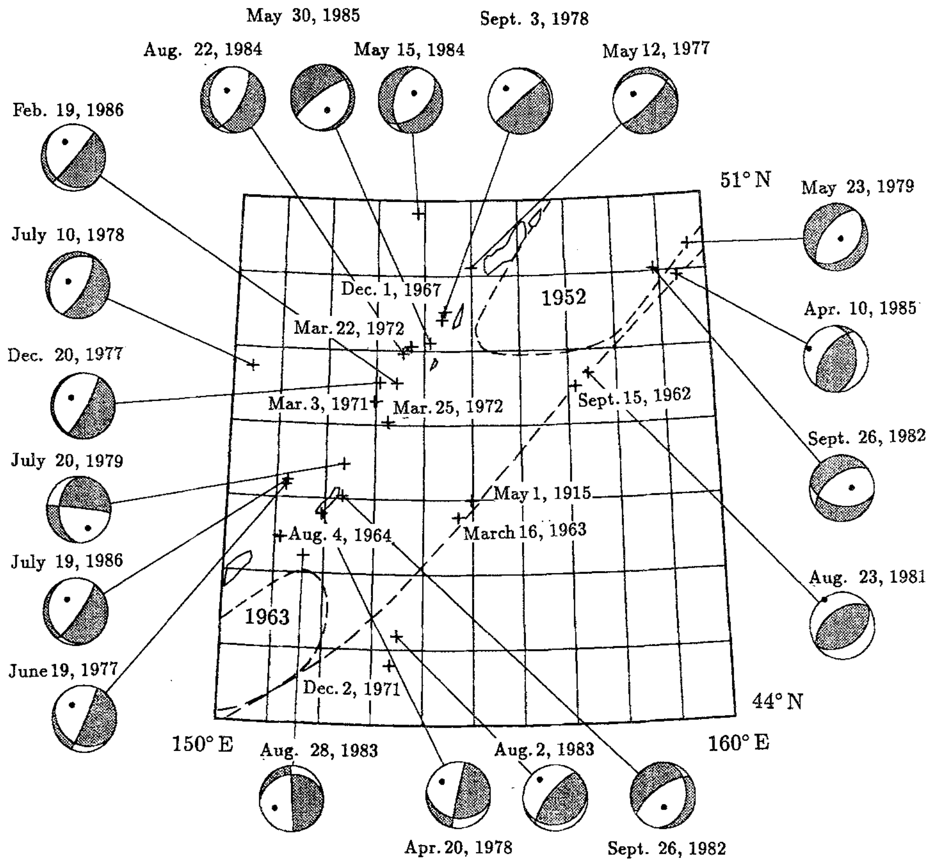

Figure 9
Sequence of events for the region of Kurile Islands Trench Gap. Aftershock areas of 1952 Kamchatka and 1963 Kurile Islands events are adapted from KELLEHER *et al.* (1974).

The Kurile Islands Trench Gap was recognized as a likely site for a future large earthquake in some earlier studies (FEDOTOV, 1965, 1967; KELLEHER, 1970; KELLEHER *et al.*, 1973, 1974). Recently CHRISTENSEN and RUFF (1987) searched the area for the presence and mechanisms of outer-rise events, with the goal of understanding space- and time variations of stresses in the outer-rise area, and their association with earthquake cycles. They note that both the 1952 and 1963 earthquakes bordering the gap have been followed by tensional outer-rise events; the area oceanward of the 1963 quake shows tensional stresses even in 1985. The situation is different for the zone of the gap itself, where they found a compressional event just in the middle of the gap (March 16, 1963, $M_b = 7.7$, see Figure 9). Also the southern end of the gap shows signs of compressional stresses (events of December 2, 1971, $m_b = 6.2$, and August 2, 1983, $m_b = 5.5$). The northern end of the gap is even more interesting: there the tensional outer-rise event of September 15, 1962 has been followed, almost in the same area, by the compressional one on August 23, 1981, showing the change of stress regime from tensional to compressional, as observed by CHRISTENSEN and RUFF (1987). Also, a similar change occurred farther north-east, around 50°N, where a tensional outer-rise event of September 26, 1982 ($m_b = 5.5$) was closely followed by the compressional one of April 10, 1985 ($m_b = 5.5$), which CHRISTENSEN and RUFF (1987) interpret as the reloading of the southern section of the 1952 region, caused by the continued loading of the gap.

We searched for possible signs of tensional stresses at the intermediate depths down-dip from the interface zone in the gap area, and the results are shown in Figure 9. In general, as shown by mechanisms of intermediate depth earthquakes within the slab in that zone (Table 1, Figure 9), tensional stresses prevail at present in the gap zone, with few exceptions (e.g., a more complex stress configuration in the southern end of the gap). Also, in some places we observe down-dip tensional events located below down-dip compressional events, which possibly are signs of the presence of a double Benioff zone there.

We feel that this observation strengthens the suggestion of CHRISTENSEN and RUFF (1987) that the potential for a large underthrusting event in this region is high. However, it is not clear to us how large earthquake(s) would rupture the gap area, because of its narrow interface, as mentioned earlier.

Northern New Hebrides: East Rennell Island Ridge

Figure 10 presents northern part of New Hebrides seismic zone, from 11.5° to 15°S. The top of the figure coincides with the sharp western bend of the seismic zone (bend placed between 11° and 11.5°S). We marked part of the aftershock zone of the December 31, 1966 earthquake ($M_s = 7.8$) and aftershock zones of the July 1980 earthquakes (July 8, 1980, $m_b = 5.9$, $M_s = 7.8$ and July 17, 1980, $m_b = 5.7$, $M_s = 8.0$). There are no clear transverse features there, but, based on seismicity observations, HABERMANN (1984) places an asperity there. This is the region where the large event of December 31, 1966 stopped after rupturing to the south from 11.8°S and 116.5°E. Both great events during July 1980 initiated in this zone. According to HABERMANN (1984) this region appears to be a simple, isolated asperity. The asperity might extend to the north, along the east edge of the 1966 and 1980 events, to include the region where the 1966 event initiated; its southern end is positioned at approximately 13°S.

The area south of it, between 13° and 14.2°S, is where the East Rennell Island Ridge intersects the subduction zone. The ridge is a major transverse feature in the New Hebrides, with 3 to 4 kilometers of relief. Both edges of this ridge appear to have stopped large earthquakes. The northern edge stopped the July 8, 1980 event, the southern edge stopped the December 28, 1973 event ($m_b = 6.3$, $M_s = 7.8$, aftershock zone shown in Figure 10). The August 11, 1970 event ($m_b = 6.1$, $M_s =$

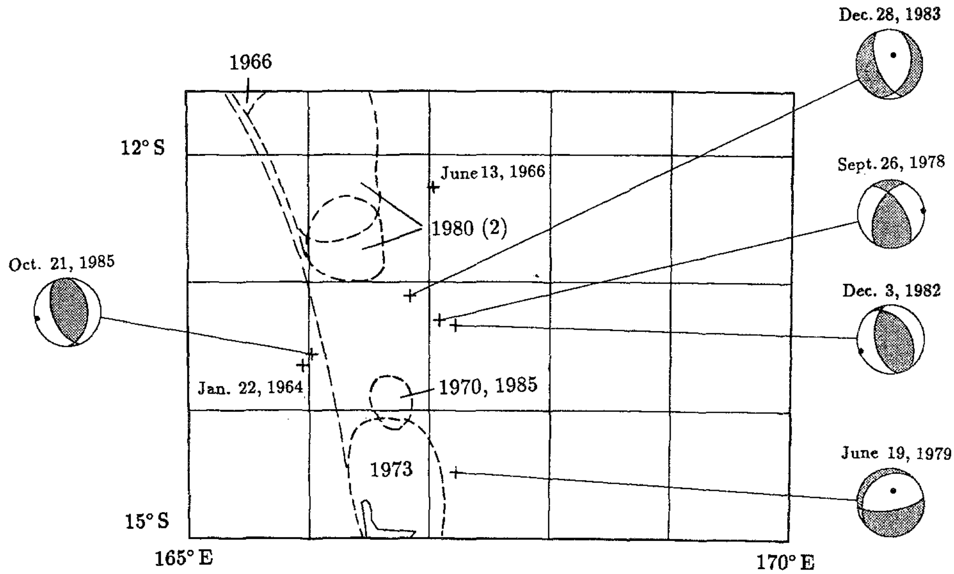

Figure 10
Sequence of events for northern New Hebrides seismic zone. Aftershock zones of significant underthrust events are adapted from CHRISTENSEN and RUFF (1987).

7.5) initiated on the southern edge of the ridge, and ruptured north into the region of intersection. This last event (aftershock zone shown in Figure 10) was followed by two large tensional outer-rise events just oceanward from the zone of the 1970 earthquake (CHRISTENSEN and RUFF, 1987), and, unexpectedly, by another moderate underthrusting event on December 21, 1985, in the same region as the 1970 earthquake.

The East Rennell Island Ridge—subduction zone intersection region is about 130 km long, and, according to HABERMANN (1984), may contain a nest of smaller asperities. CHRISTENSEN and RUFF (1987) found in this area one tensional outer-rise event of January 22, 1964 ($m_b = 6.3$), prior to the surrounding subduction events, followed by the compressional outer-rise earthquake of October 21, 1985 ($m_b = 5.5$, shown in Figure 10), suggesting that the stress state of the outer-rise has changed since that time from tensional to compressional, following the subduction of segments to the north and to the south. We searched the area down-dip from that zone trying to find possible signs of tensional stresses in the downgoing slab, and we have found a normal event on December 28, 1983 ($m_b = 5.5$, depth 114 km, Figure 10). Here we agree with CHRISTENSEN and RUFF (1987) that this zone shows signs of a mature seismic gap, and we suggest that the occurrence of the December 28, 1983 event which we have identified, strengthens such a contention.

We have also found two deeper events (September 26, 1978, $m_b = 5.6$, depth 204 km; December 3, 1982, $m_b = 5.5$, depth 261 km) with mostly compressional mechanisms (Figure 10, Table 1). ISACKS and MOLNAR (1971) found also a compressional event of June 13, 1966 (Figure 10, Table 1), which occurred half a year before and downdip from the zone of a large subduction earthquake of December 31, 1966 ($M_s = 7.8$). We think that all three compressional events occurred at depths where the pulsating stresses associated with earthquake cycles in this particular subduction zone, do not reach.

It is perhaps of interest to note here, that in our effort to find intermediate-depth earthquakes in this area, we noted an earthquake of June 19, 1979 ($m_b = 5.7$, depth 126 km, Figure 10), positioned down-dip from the underthrust event of 1973. This event occurs in the beginning of the new cycle in this area, and shows the tension almost parallel to the trench, in agreement with our mechanical interpretation of the present situation in that zone.

Conclusions

The main concern of this work is an intermediate-term precursor to large earthquakes in coupled subduction zones, consisting of a combination of compressional outer-rise event(s) and tensional, intermediate-depth earthquakes in the downgoing slab. If such earthquakes are present in a seismic gap, they signal that the gap is in the latter part of the earthquake cycle, i.e. that the interplate interface, where main thrust events occur, is locked, causing regional compressional stresses to dominate in the outer-rise area, and tensional stresses to be increased at intermediate depths in the downgoing slab.

We have discussed results of a simple mechanical model of the subduction process, along with observational data from different places around the globe. The data cover two cases of closed seismic gaps, namely the region of the Chilean Valparaiso earthquake of March 3, 1985, and the area of an October 4, 1983 earthquake. Four other cases concern still to-be-closed gaps in northern Chile and along the coast of Guatemala, and also the Kurile Islands Trench gap and the northern New Hebrides gap.

We conclude that the presence of both compressional outer-rise events and increased number of tensional intermediate-depth earthquakes in the downgoing slab strengthens the evaluation of the gap maturity. The material presented here is limited and further work on the areas of other seismic gaps is needed. The precursor discussed by us could be useful in evaluating the maturity and hence great earthquake potential of a seismic gap.

Our observational results also show some local deviations from a simple subduction process associated with geometrical irregularities along the strike, and also with loading coming from the sides of the seismic gap, and we acknowledge the importance of these factors for future models of subduction process.

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