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The Asperity Model and the Nature of Large Subduction Zone Earthquakes

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Abstract. Regional variations in the rupture characteristics of large shallow earthquakes in circum-Pacific subduction zones are interpreted in the context of the asperity model of heterogeneous stress distribution on the fault plane. It is assumed that the degree of seismic coupling between the downgoing and overriding plates is reflected in the maximum earthquake rupture dimensions in each region, and that gross features of the regional stress distribution can be inferred from the rupture process of large earthquakes. The results of numerous studies of the historic record and detailed source process of large subduction zone events are summarized for each region. The systematic variation in maximum rupture extent in different zones indicates that four fundamental categories of behavior are observed. These are (1) the Chile-type regular occurrence of great ruptures spanning more than 500 km; (2) the Aleutian-type variation in rupture extent with occasional ruptures reaching 500 km in length, and temporal clustering of large events; (3) the Kurile-type repeated failure over a limited zone of 100–300 km in length in isolated events; and (4) the Marianas-type absence of large ruptures. The rupture processes associated with each category have distinctive features. The great earthquakes of category (1), such as the 1964 Alaskan event, are often preceded by a prolonged period of increased seismicity, and the body wave source process is characterized by a long duration ($\geqslant 120$ sec) time function. The earthquakes in category (2) are usually preceded by seismic quiescence and may occur as doublets or multiplets. The body wave source time functions of these events tend to consist of several long duration (30 to 60 sec) discrete ruptures. Category (3) events commonly have a precursory quiescence followed by extensive foreshock activity, and the body waves are complicated since they result from a sequence of short duration (<30 sec) ruptures. Interpreting these features with the asperity model indicates that the stress distribution in categories 1,2,3, and 4 are characterized by very large asperities and strong coupling, large but discrete asperities, numerous smaller asperities, and an absence of significant asperities, and large component of aseismic slip, respectively.

1. Introduction

The nature of the stress distribution on the plate contact in a subduction zone determines the seismic behavior associated with the subduction process. It is reasonable to assume that the strength of coupling between the plates is characterized by the size of earthquakes in the zone, and that analysis of the seismic rupture process will provide some detail of the stress distribution. Quantification of the seismic coupling has an important bearing on the plate driving mechanism, mechanical properties of the fault zone material,

and occurrence of aseismic slip. The observed variations in earthquake size and rupture mode indicate the complexity and heterogeneity of subduction zone stress regimes, which require a general classification system. In this review we draw upon the results of numerous studies of subduction zone events to develop a categorization of subduction zones based upon similarities in their rupture characteristics and provide a physical interpretation of the stress regime associated with each category.

Some general inferences about the state of stress in subduction zones have been made based on maximum rupture lengths in each zone. KANAMORI (1971b, 1977) proposed that the degree of coupling on the fault plane is reflected in maximum earthquake dimensions, with strong coupling resulting in very large rupture zones, and decoupling associated with the absence of large earthquakes. KANAMORI (1971b) proposed a model of gradual thinning and weakening of the ocean-continent lithospheric boundary to account for the range in degree of coupling around the northwest Pacific. LAY and KANAMORI (1981) used maximum rupture zone size to categorize the circum-Pacific subduction zones and sought to relate characteristic rupture dimensions to the gross features of the stress regime in each region.

The proposed relationship between maximum rupture length and coupling on the fault plane is largely qualitative and is somewhat obscured by the complexity of subduction regimes. The width of the lithospheric interface has been correlated with maximum length of rupture zone by ISACKS *et al.* (1968) and KELLEHER *et al.* (1974), with regions of broad interface contact having the longest rupture zones, and a corresponding greater degree of coupling. However, the presence of topographic features on the subducted seafloor (KELLEHER and MCCANN, 1976, 1977) and the presence of transverse features such as ridges, submarine canyons, or changes in the strike of the trench that indicate segmentation of the subduction zone (MOGI, 1969b; VOGT, 1973; CARR *et al.*, 1974; KELLEHER and SAVINO, 1975; VOGT *et al.*, 1976; SPENCE, 1977; CHUNG and KANAMORI, 1978a, b) complicate the relationship between contact area and rupture length. The effect on the stress regime produced by these features is difficult to predict, and must be determined from the seismic behavior of the zone. It is encouraging that a correlation between convergence rate and lithospheric age with maximum earthquake size in each zone was found by RUFF and KANAMORI (1980a). This correlation suggests that factors which may affect the subduction zone geometry and normal stress on the fault plane do correlate with maximum earthquake size, which is used to infer the degree of coupling, despite the complexity of the various subduction zones.

It is possible to resolve some of the major features of the stress distribution in subduction zones by analysis of the seismic signals of large events and the associated seismicity in each region. Many events from each region must be studied before any confidence can be placed in generalizations about the stress regime associated with a particular zone. LAY and KANAMORI (1980) studied body waves and surface waves of recent large events in the Solomon Islands, and suggested that relative body wave complexity can be used to infer the relative degree of stress heterogeneity, particularly in conjunction with foreshock and aftershock analysis. Other efforts to elaborate on the nature of failure of large earthquakes have demonstrated the complexity and regional variations of the rupture process. Relative timing of body wave arrivals has indicated the multiple rupture nature of some events, as manifested in body wave complexity (IMAMURA, 1937; MIYAMURA *et al.*, 1965; WYSS and BRUNE, 1967; TRIFUNAC and BRUNE, 1970; NAGAMUNE, 1971; FUKAO and FURUMOTO,

1975; WU and KANAMORI, 1973), while synthetic seismograms have been used in more detailed analysis of other complex events (FUKAO, 1972; CHUNG and KANAMORI, 1976, 1978a; ABE, 1977; KANAMORI and STEWART, 1978; RIAL, 1978; FUKAO and FURUMOTO, 1979; STEWART *et al.*, 1981; EBEL, 1980; BOATWRIGHT, 1980). The detailed source analyses of large earthquakes using body and surface waves include events in the Solomon Islands, New Hebrides, Japan trench, Kurile Islands, Aleutian Islands, Alaska, Middle America, and South America. We will use these results along with studies of seismicity to further characterize regional variations in subduction zone stress regimes.

In many of the studies of large earthquakes it has been found that the body wave moment and inferred source area are smaller than the surface wave determinations, particularly for multiple rupture events (e.g. LAY and KANAMORI, 1980; EBEL, 1980; BOATWRIGHT, 1980; CHAEL and STEWART, 1982). There are also many observations of clustering of foreshocks and aftershocks within rupture zones, and other seismicity patterns which indicate localized stress concentrations along subduction zones as reviewed by KANAMORI (1981). These observations suggest that there are areas of high and low breaking stress on the fault, and that the distribution of these areas governs the rupture process. The asperity model proposed by KANAMORI (1978a) provides a useful framework in which to discuss and categorize the variations in stress concentration and degree of coupling determined from the seismic record.

In this review we will consolidate the seismological observations pertinent to the stress distribution in each zone, then categorize the zones on the basis of systematic features in the large earthquake record, and finally interpret these categories with the asperity model to infer the nature of coupling in each zone.

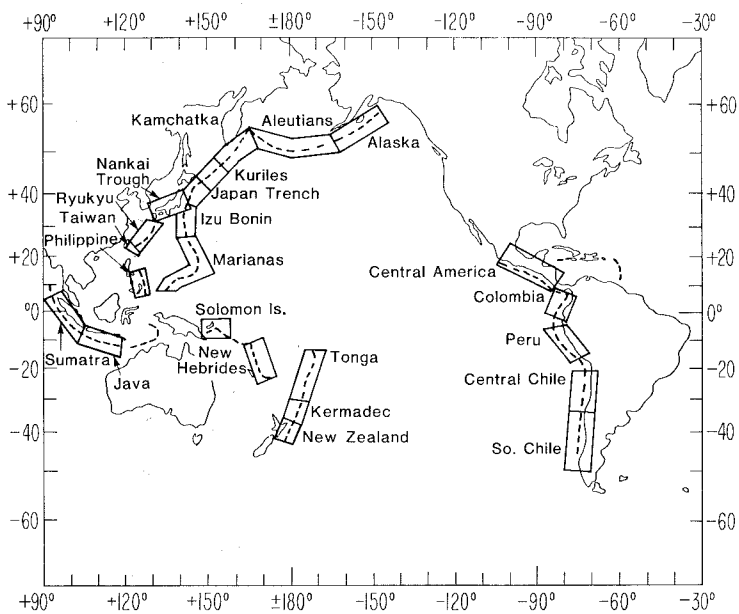

Fig. 1. Map showing the circum-Pacific subduction zones considered in this review.

2. Regional Characteristics of Large Earthquakes

The historical record of large earthquakes in many tectonic regions has been reviewed by MCCANN *et al.* (1979), who emphasized assessment of the seismic potential of each zone. In this section, our emphasis is on compiling the observations of the rupture process of large subduction zone events, the regional seismicity associated with these events, and features in the trench morphology which can be related to the nature of coupling in each region. Despite the limitations of the historic record in most circum-Pacific regions, almost all subduction zones have had at least one sequence of large earthquake activity recently enough to have been analysed using seismicity patterns and seismic waves. We will summarize the results of such studies for each of the subduction zones indicated in Fig. 1.

2.1 Solomon Islands

LAY and KANAMORI (1980) conducted a study of six large events in the Solomon Islands. These events are believed to have resulted in rupture of the entire length of the Solomon trench northwest of the Woodlark ridge and a significant portion of the New Britain trench, thus constituting a full sequence in the failure history of this region. Through detailed surface wave and body wave analysis, focal mechanisms and dislocation parameters for each event were determined. The source mechanisms and two-day aftershock zones of each event are shown in Fig. 2. An unusual characteristic of these large Solomon Islands earthquakes is their tendency to occur in pairs of events with close temporal and spatial relations. The three doublets that occurred in 1971, 1974, and 1975 are the six largest events in the region over a 25 year span. Each doublet member was separated by only a few hours in time, or a few days in the case of the 1971 events which propagated around the corner at the junction of the Solomon and New Britain trenches. Southeast of the 1974 events the Woodlark ridge enters the Solomon trench, disrupting the bathymetry and seismicity.

The body waves from these events consist of impulsive arrivals usually indicating either one or two dislocation sources having 10 to 15 sec rupture durations. The source dimensions inferred from the body waves were found to be substantially smaller than the areas of total rupture surface indicated by aftershocks. A summary of the body wave and surface wave source areas for two of the doublets is given in Table 1. These events, and what little is known about the historic record, indicate that maximum rupture zone lengths of 150–250 km are typical in this complicated subduction zone (MCCANN, 1981).

The Solomon trench shallows out in the region where the Woodlark ridge intersects it, but is well-defined southeast of the ridge along the island Guadalcanal. This region has similar behavior to the northwest part of the trench with large earthquakes occurring as doublets in 1977 and 1978. These events had maximum rupture extents similar to the northwestern events. The historic record of the Woodlark ridge region does not clearly define whether this portion of the trench fails seismically, or not. It seems that this is a clear instance of a transverse feature segmenting a trench, but whether subduction is locked up here due to density contrast as in the model of KELLEHER and MCCANN (1976), or whether there is a large component of aseismic slip due to inherent ductility of the thermal anomaly is not clear. The similarity of activity on either side of the ridge does suggest the latter possibility. The trench is segmented by the subducting ridge, the junction

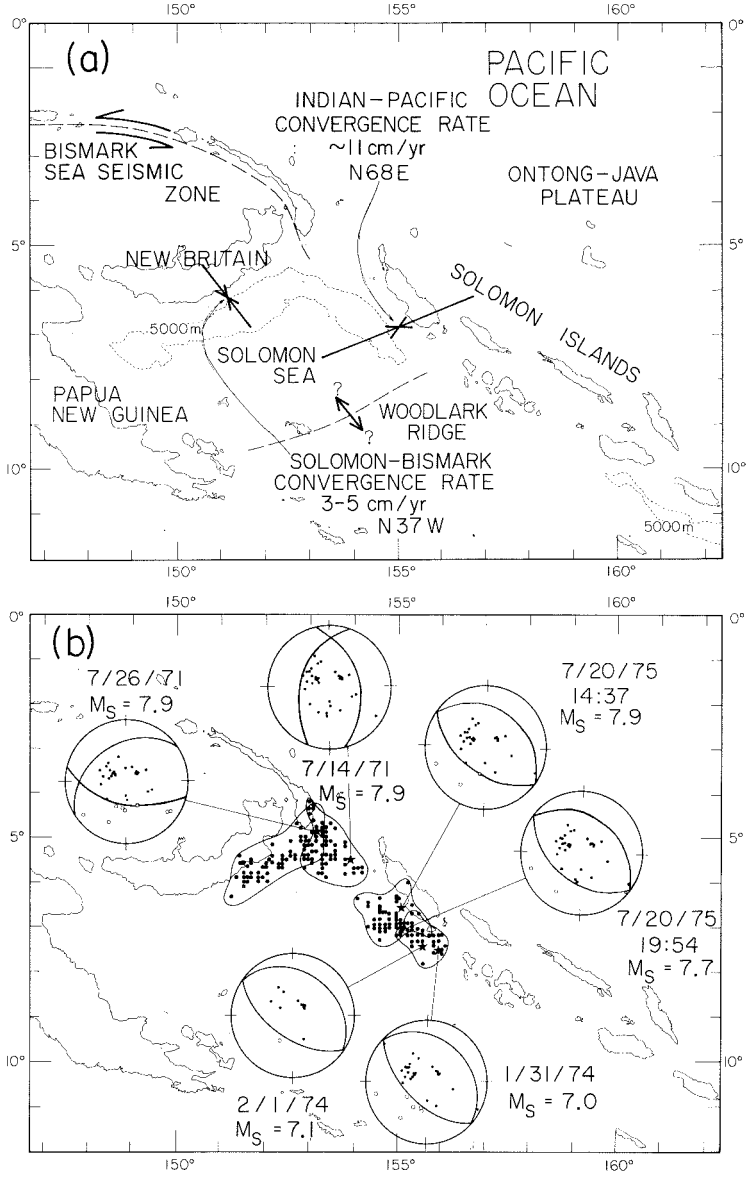

Fig. 2. (a) Tectonic features of the Solomon Islands region. (b) Fault mechanisms of three large Solomon Islands doublets determined by LAY and KANAMORI (1980).

with the New Britain trench, and the poorly defined transition to a strike slip environment east of Guadalcanal. These major features have delimited the rupture zones of large earthquakes, as in the 1974 and 1971 sequences. It is significant that the 1971 doublet does appear to have overcome the geometric variation at the trench junction. Though bathymetry in the region is not very detailed, there is no indication of bathymetric features

Table 1. Large, shallow circum-Pacific subduction zone

Region	Date	$m_B(T)$*	M_s	M_w	M_t†	M_o (10^{27} dyne-cm)	Length (km)	Width (km)	Average displacement D(m)
Solomon Is.	7/14/71	7.4(8)	7.9	8.0		12.0	200.	70.	1.3
	7/26/71	7.4(8)	7.9	8.1		18.0	300.	70.	1.3
	1/31/74		7.0	7.3		1.0	~36.	75.	0.6
	2/01/74		7.1	7.4		1.4	~40.	75.	0.6
	7/20/75		7.9	7.7		3.4	90.	90.	0.5
	7/20/75		7.7	7.4		1.2	45.	90.	0.5
Nankai Trough	12/23/1854		[8.4]g		8.25		230.	70.	~4.0
						46.0	265.	70.–100.	~4.0
	12/23/1854		[8.4]		8.25		230.	70.	~4.0
	12/07/44	7.8(15)	8.2	8.1	8.2	15.0	120.	80.	3.1
				8.2		28.0	190.	75.	4.0
	12/20/46	7.8(9)	8.2	8.1	8.0	15.0	120.	80.	3.1
				8.5		66.0	300.	85.	5.0
						78.0	300.	90.	10(5–18)
South Honshu	1703					37.–41.	200.	70.	6.–7.
	9/01/23	7.7(9)	8.2	7.9	8.0	7.6	130.	70.	2.1
				8.0		11.0	85.	55.	6.7
				7.9		8.4	95.	54.	4.8
	11/25/53	7.7(8)	7.9	7.9	7.9	9.0	75.	55.	
Central Honshu	6/15/1896		7.9		8.6	63.(t)a	300.	160.	1.3
	8/05/1897		8.1	8.0		12.0	100.	50.	~3.5
	3/02/33Nb	8.3(12)	8.5	8.4	8.3	43.0	185.	100.	3.3
	5/23/38	7.2(10)	7.4			4.0	75.	40.	2.7
	11/05/38	7.4(9)	7.7			7.0	100.	60.	2.3
	11/05/38	7.3(10)	7.8			4.8	100.	60.	1.6
	11/06/38N	7.4(11)	7.7			3.8	85.	45.	2.0
	11/06/38N	7.1(7)	7.1			3.2	95.	45.	1.5
	5/16/68	7.6(12)	8.1	8.2	8.3	28.0	150.	100.	4.1
	6/12/68	7.2(9)	7.2			0.5	80.	30.	0.5
	6/12/78		7.7			3.1	70.	34.	1.9/2.4
						2.2			
Hokkaido	3/04/52	8.0(9)	8.3	8.1	8.2	17.	180.	100.	1.9
				8.2		23.	130.	100.	
	6/17/73	7.3(7)	7.7	7.8	8.0	6.7	100.	60.	1.6
Kuriles	11/06/58	8.2(6)	8.1	8.3	8.2	44.0	150.	70.	5.1
	10/13/63	7.7(14)	8.1	8.5	8.4	75.0	200.–300.	150.	3.0
			8.0				250.	70.	3.1–4.6
	10/20/63	7.1(14)	7.3		7.9		85.	70.	0.7–1.1
							100.	60.	
	8/11/69	7.9(10)	7.8	8.2	8.2	22.0	180.	85.	2.9
	6/10/75		7.0				100.	60.	
						2.0@200s	50.	35.	
						0.5@ 50s			
						1.7–2.1@200s			
	3/23/78		7.5	7.4		1.5	90.	55.	
	3/24/78		7.6	7.6		2.8	105.	75.	
Kamchatka	2/03/23	7.7(7)	8.3	8.3	8.8		220.	90.	
	1/04/52	7.9(8)	8.3	9.0	9.0	350.0	650.	200.	5.
							700.	240.	
							450.	175.	
	5/04/59	7.8(9)	7.7	8.2	8.0		160.	110.	
	11/22/69	7.0(7)	7.3	7.8	7.8	7.0	110.	60.	
	12/15/71		7.8	7.9	7.8	10.0	120.	60.	
Aleutians	3/07/29N	7.7(8.4)	7.7	7.8	8.1	6.7@100s			
	4/01/46	6.6(5.1)	7.4		9.3	11. @100s			
		7.2(8)				370.(t)			
						150.(t)	400.	120.	
	3/09/57	7.7(13)	8.1	9.1	9.0	150.(t)	900.	100.	
	2/04/65	7.7(10)	8.2	8.7	8.6	140.	500.	150.	2.5
							420.	50.	7.5
	3/30/65N	7.4(10)	7.5	7.6	7.8	3.4	80.	50.	1.2

earthquakes for which the source process has been investigated.

Stress drop $\Delta\sigma$ (bar)	Rupture velocity v_r (km/s)	$\tau(s)$**	$t_c(s)$**	l(km)§	$M_o^{t\#}$ (10^{27} dyne-cm)	Body wave time function characteristics	Reference
19.		5	9	27–54	0.7	Single Event	LAY and KANAMORI, 1980
17.	2.5	5	11	33–66	0.4	2 Subevents	" "
12.						Single Event	" "
12.						Complex Event	" "
9.		5	9	27–54		2–3 Subevents	" "
9.		4	6	18–36	0.3	Single Event	" "
							ANDO, 1975
							ISHIBASHI, 1981
							ANDO, 1975
33.							KANAMORI, 1972a
							ISHIBASHI, 1981
33.							KANAMORI, 1972a
							ANDO, 1975
							FITCH and SCHOLZ, 1971
						3 Subevents (?)	YOSHIYAMA, 1950
							MATSUDA *et al.*, 1978
18.							KANANAMORI, 1971c
							ANDO, 1974
45.	~2.0	~5.0					MATSU'URA *et al.*, 1980
							ANDO and SENO, 1981
						3 Subevents (?)	USAMI, 1956
					.19		KANAMORI, 1972b
							SENO, 1979
42.	3.2						KANAMORI, 1971a
57.				60.–75.			ABE, 1977
33.							" "
23.							" "
38.							" "
28.							" "
32.	~3.5						KANAMORI, 1971d
						2 Subevents, 45 sec apart	NAGAMUNE, 1971
	2.0 initial					3 Subevents	FUKAO & FURUMOTO, 1975
	3.5–4.5						
12.							YOSHIOKA & ABE, 1976
95./145.	3.2	2		37/34	3.0	2 Subevents	SENO *et al.*, 1980
							KANAMORI & GIVEN, 1981
17.							KANAMORI, unpubl.
							KASAHARA, 1975
35.							SHIMAZAKI, 1974
78.							FUKAO & FURUMOTO, 1979
23.	2.7–4.5						KANAMORI, 1970a
						2 Subevents	NAGAMUNE, 1971
		~5	30	~50	19	>3 Subevents (30 sec each)	RUFF & KANAMORI, 1980b
	3.5						BEN-MENAHEM & ROSEMAN, 1972
							" "
	2.0					Multiple Rupture	FUKAO, 1979
29.	3.8						ABE, 1973
	3.5–4.5					Single Event	FUKAO & FURUMOTO, 1975
	2.0					Multiple Rupture	FUKAO, 1979
				.18@30s		" "	SHIMAZAKI & GELLER, 1977
						" "	TOKEMURA *et al.*, 1977
							REGAN & KANAMORI, 1980
							" "
							SENO & EGUCHI, 1981
							KANAMORI, 1976
	3.0						BEN-MENAHEM & TOKSOZ, 1972
							KELLEHER *et al.*, 1973
							SENO & EGUCHI, 1981
							KURITA & ANDO, 1974
							" "
							KANAMORI, 1972b
							" "
							HATORI, 1981
							" "
30.	4.0					Multiple Event	WU & KANAMORI, 1973
	3.0						BEN-MENAHEM & ROSEMAN, 1972
			60	120	5.0	Two 60-sec ruptures	RUFF & KANAMORI, 1980b
33.							ABE, 1972a

Table 1. (continued)

Region	Date	$m_B(T)$*	M_s	M_w	M_t†	M_o (10^{27} dyne-cm)	Length (km)	Width (km)	Average displacement D(m)
Alaska	11/10/38	8.2(13)	8.3	8.2	8.4	28.(t) 21.@100s 35.(t)	400. 250.	100. 100.	
	7/10/58	7.4(10)	7.9	7.7	8.1	4.0	300.	10.	4.0
	3/28/64	7.9(13)	8.4	9.2	9.1	750.0	500.	300.	7.0
	7/30/72		7.6			4.0	170.	10.	
	2/28/79		7.7			6.5	60.–80.	5.0–6.5	.6–6.
Java	8/19/77		8.1	8.3		100.	160.	100.	
Kyushu	4/01/68	7.3(5)	7.5			1.8	56.	32.	1.6
Taiwan	3/12/66	7.6(10)	7.5			1.6	27.	28.	2.1
Mindanao	8/16/76		7.8	8.1		19.0	160.	80.	3.0
Middle America	8/23/65	7.7(15)	7.6	7.5		1.7	105.	47.	0.7
	8/02/68	7.2(12)	7.1	7.3		1.0	84.	50.	0.5
	4/29/70		7.3	7.4		1.2	130.	96.	0.2
	1/30/73		7.5	7.6	8±	3.0 3.0	90. 90.	70. 63.	1.4 1.1
	11/29/78		7.8	7.6		3.2	85.	65.	1.2
	3/14/79		7.6	7.6		2.7	70.	57.	1.4
Colombia	1/31/06	8.2(9)	8.7	[8.8]	8.7	[200.]	~500.	~200.	[5.2]
	5/14/42	7.7(8)	7.9	[7.6]		[3.2]	~100.	~60.	[1.3]
	1/19/58	7.4(11)	7.8	7.7		5.2	~100.	~60.	2.3
	12/12/79		7.7	8.2	8.1	29.0	230.	120.	2.7
Peru	5/24/40	7.9(8)	8.0	[8.2]		[25]	~230.	~100.	
	8/24/42	7.9(11)	8.1	[8.2]		[27]	~170.	~150.	
	10/17/66	7.5(16)	7.8	8.1	8.2	2.0	140.	80.	2.6
	5/31/70N	7.5(8)	7.6	7.9		10.	130.	60.–80.	1.4–1.8
	10/03/74	7.7(4)	7.6	8.1	8.1	15.			
Chile	5/22/60	7.9(11)	8.5	9.5	9.4	2700.	800.	200.	24.
	6/06/60	7.1(7)	6.9			5.6			
	3/28/65N	7.4(15)	7.5			1.8			1.6
	7/09/71		7.5			5.6	100.	70.	1.6–5.7
New Hebrides	8/11/65	6.8(10)	7.0			0.7			
	8/11/65	7.1(10)	7.5			3.0	~100.	~50.	1.5
	8/13/65	6.8(15)	7.0			1.7			
	10/25/80		7.2			3.0	45.	45.	

*$m_B(T)$ is the long period body wave magnitude as given by GELLER and KANAMORI (1977); GELLER *et al.* (1978); ABE and KANAMORI (1980); or ABE (1981).
†M_t is the tsunami magnitude given by ABE (1979).
**τ is the body wave time function rise time.
xxt_c is body wave time function duration.

separating the 1971, 1975, and 1974 failure zones.

The distribution of seismicity for $m_b \geqslant 4.0$ in the region is shown as projected along the Solomon trench in Fig. 3. There is an apparent period of quiescence before large doublet events in 1971, 1974, 1977, and 1978 as well as preceding a large single event in 1966, as has also been noted by MCCANN (1981). Note that the 1975 zone which lies between the 1971 and 1974 doublets experienced an increase in seismicity prior to failure. The rather even distribution of events of $M_s \geqslant 7.0$ along the trench indicates that the whole trench has failed, and there is a suggestion of migration of activity from northwest to southeast commencing in 1971, though the 1975 events deviate from this. JONES and MOLNAR (1979)

Stress drop $\Delta\sigma$ (bar)	Rupture velocity v_r (km/s)	$\tau(s)$**	$t_c(s)$[xx]	l(km)[§]	M_0'[#] (10^{27} dyne-cm)	Body wave time function characteristics	Reference
							HATORI, 1981
							DAVIES *et al.*, 1981
							ANDO, 1977
28.	3.5						KANAMORI, 1970b
			120	140–200	100.	Single Source	RUFF & KANAMORI, 1980b
	3.5					7 Subevents	WYSS & BRUNE, 1967
							PAGE & GAWTHROP, 1973
	2.5				1.2		LAHR *et al.*, 1980
	2.2		4./9./6.	12/27/17	1.2	3 Subevents	BOATWRIGHT, 1980
							STEWART, 1977
60.							SHIONO *et al.*, 1976
66.	1.9					2 Subevents	SUDO, 1972b
17.	2.5						STEWART & COHN, 1979
6.		5	10	25[§§]	1.9	Single Event	CHAEL & STEWART, 1982
4.		5	10	25[§§]	0.8	Single Event	" "
1.		7	10	25[§§]	0.5	2 Subevents	CHAEL & STEWART, 1982
					0.46	4 Subevents	YAMAMOTO & MITCHELL, 1978
10.						Multiple Event	REYES *et al.*, 1979
7.					0.5–1.0	Multiple Event	CHAEL & STEWART, 1982
8.	2.5	4	10	25[§§]	1.9	Single Event	STEWART *et al.*, 1981
10.	3.0	7	10	25[§§]	1.0		CHAEL & STEWART, 1982
							KANAMORI & MCNALLY, 1981
							" "
							" "
	2.0						KANAMORI & GIVEN, 1981
			60	~120.	20	One 60-sec Rupture	RUFF & KANAMORI, 1981
							KELLEHER, 1972
							ABE, 1972b
42.						Multiple Event	LOMNITZ & CABRÉ, 1968
23.–35.	2.0–3.0						ABE, 1972b
						Multiple Event	LOMNITZ, 1971
							STEWART, pers. comm.
				15		6–9 Subevents	SPENCE *et al.*, 1980
	3.5						KANAMORI & CIPAR, 1974
						Multiple Event	NAGAMUNE, 1978
						~15 Subevents	KANAMORI & STEWART, 1979
91.		~2.	6	17[§§]	10.		MALGRANGE *et al.*, 1981
38.		4	20	37–41[§§]	56.	2 Subevents	" "
		3	4	16	1.6	Single Event	EBEL, 1980
		1	5–7	14–18	1.6	2 Subevents	" "
		1	11			2 Subevents	" "
							VIDALE & KANAMORI, 1981

[§]l is the body wave source dimension; either a characteristic length or radius of a circular rupture (indicated by §§).
[#]M_0' is the body wave moment.
[c]Numbers enclosed in brackets are inferred from macroseismic data or aftershock area.
[a]Moments marked by (t) are based on tsunami source dimensions.
[b]Events indicated by N are normal faulting events.

found that the Solomon Islands region is distinctive in not having pronounced foreshock activity before large events. MCCANN (1981) has noted that seismicity in this region may be the most active in the world.

The historical seismicity of large events in this region is shown in Fig. 4 (LAY and KANAMORI, 1980; MCCANN, 1981). For the western portion of the trench, the recurrence rate appears to be about 25 years, while a slightly longer rate of about 40 years is indicated for the southeastern portion of the trench. The earlier events are not well-documented but do cluster closely in time as in 1944–1945, 1919–1920, and 1931. A high convergence rate of 11 cm/yr in the Solomon trench is given by MINSTER *et al.* (1974), which may be as-

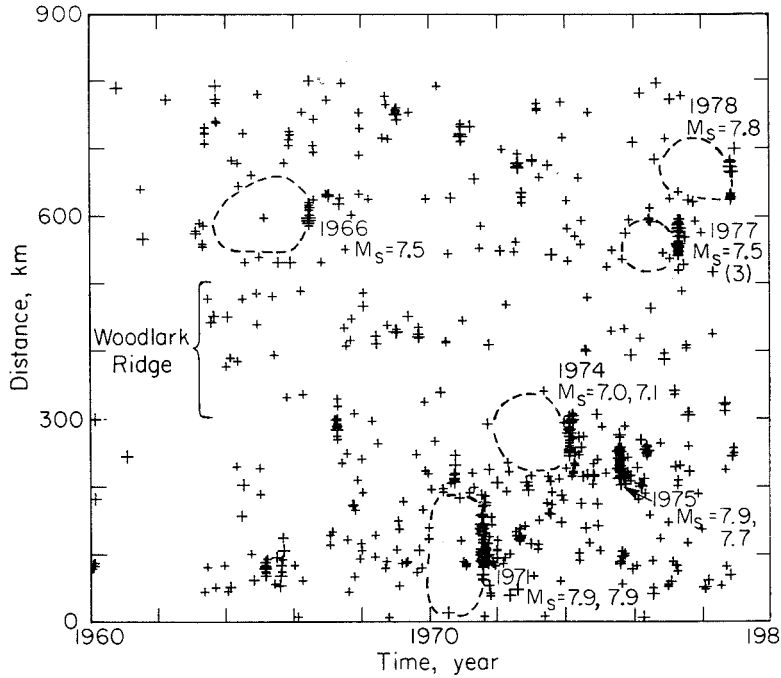

Fig. 3. Space-time plot of seismicity for the Solomon Islands region for events with $m_b>4.0$ and depths less than 60 km obtained from the NOAA catalog. Distance is measured along the Solomon trench from the New Britain trench junction.

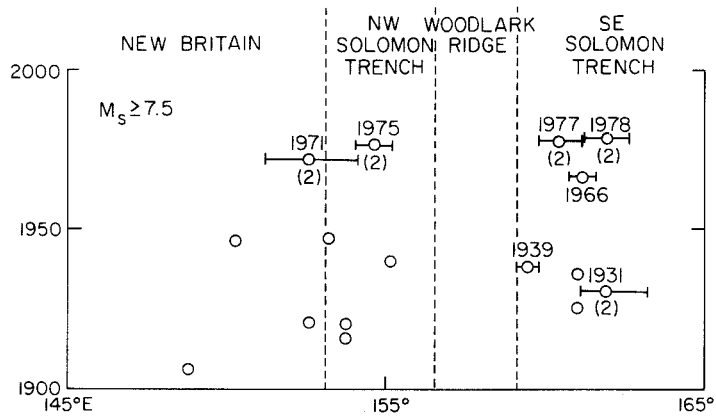

Fig. 4. Historic record of $M_s \geq 7.5$ events in the New Britain and Solomon trenches. Aftershock zone estimates are from LAY and KANAMORI (1980) and MCCANN (1981).

sociated with the short recurrence interval. Also, the overriding plate here is the Ontong-Java Plateau, which has a thick (30–40 km), semi-continental crust (FURUMOTO *et al.*, 1976). The width of the contact zone defined by seismicity is fairly narrow, consistent with the small upper bound on rupture length as suggested by KELLEHER *et al.* (1974), but the nature of the overriding plate may influence the stress distribution differently from other island arcs.

2.2 *Japan*

Of all circum-Pacific subduction zones, the historical record of large earthquakes is most complete in the Nankai trough and Japan trench regions. The Philippine Sea plate underthrusts Japan toward the northwest in the Nankai trough, beneath Shikoku and southern Honshu. This region has an exceptional history of great earthquakes, as documented by IMAMURA (1928) and ANDO (1975), and summarized in Fig. 5(b). The average repeat time for great earthquake sequences rupturing the entire trench is 170 years with a large standard deviation of 68.9 years (ANDO, 1975; RIKITAKE, 1976). There is substantial regularity in the earthquake sequences, with each trench segment failing after a similar time interval and frequent clustering of the major events. These aspects indicate the uniformity of the subduction process along this trench, however there are large variations in repeat time as indicated by the large standard deviation. There is also a tendency for the entire trench to occasionally fail in a single event, as apparently occurred in 887 and 1707, with total rupture lengths of ~500 km.

MOGI (1969b) noted that the 1946 Nankaido event rupture zone is bounded by transverse structural boundaries. The segmentation of the Nankai trough indicated in Fig. 5(b) was suggested by ANDO (1975) on the basis of transverse geologic and bathymetric structures on the underthrust and overriding plates. The southern end of the trough is bounded by the Kyushu-Palau ridge, and the northern end abuts the South Honshu ridge (KELLEHER and MCCANN, 1976). Both of these ridges intersect the convergent zone at discontinuities in strike of the zone. DEN (1968) and NISHENKO and MCCANN (1979) have discussed the correlation between the rupture zones of the 1944 and 1946 events and the submarine basins bordering the Kii Peninsula, offsets of the inner margin of the Nankai trough, and transverse structures in the trench slope. MOGI (1969b) discussed how aftershocks of the 1944 and 1946 events expanded northeast and northwest respectively from their epicenters which lie adjacent to one another at the tip of the Kii Peninsula. This peninsula appears to have separated rupture zones in previous sequences such as the 1854 events which occurred 32 hours apart on December 23 and 24. However, the 1707 event appears to have broached this boundary in a single, perhaps complex event which shook Kyoto for 15 min (ANDO, 1975).

KANAMORI (1977) found a seismic slip rate of 3.5 cm/yr in the shallowly dipping (10°) Nankai subduction zone. The lack of constraint on the Philippine plate motion makes comparison of this seismic slip rate with the total convergence rate in the trench somewhat uncertain. The convergence rate may vary from west to east along the trench, ranging from 4.3 to 3.3 cm/yr (SENO, 1977; SENO and EGUCHI, 1981). This indicates that nearly 100% of the convergence may be accommodated seismically. This zone may be strongly coupled due to the recent onset of subduction and the young age of the underthrusting Shikoku Basin (20–30 my) (KANAMORI, 1972a; SENO, 1977). However, the 1946 Nankaido event has a seismic moment obtained from long period surface waves substantially smaller than that inferred from geodetic analysis, which may indicate substantial aseismic slip during these events (ANDO, 1975; KANAMORI, 1972a; FITCH and SCHOLZ, 1971). The eastern portion of this region did not fail in the 1944 event and remains today as the heavily monitored Tokai seismic gap. There may be a geometric boundary segmenting this portion of the region due to curvature of the trench.

The 1944 and 1946 events were studied by surface wave analysis (KANAMORI, 1972a),

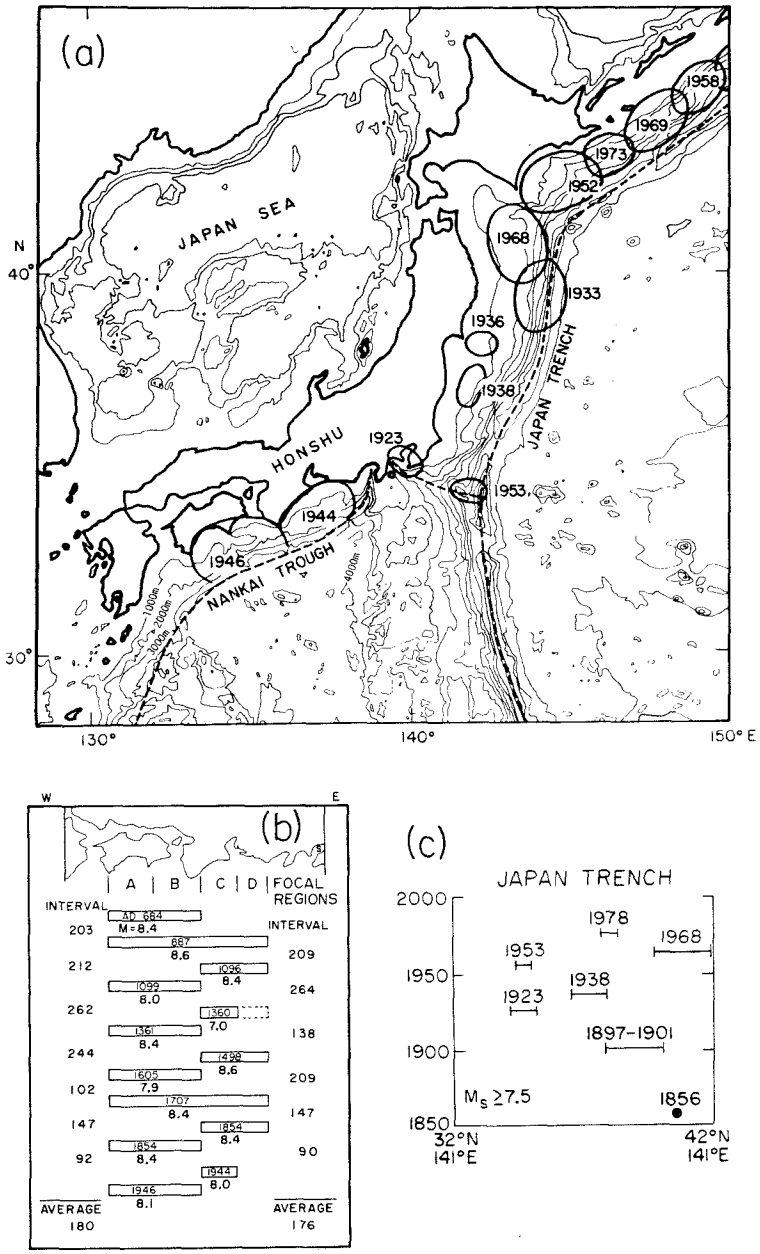

Fig. 5. (a) Map of large earthquake rupture zones in the Nankai trough and Japan trench since 1900. (b) Historic record of great earthquakes in the Nankai trough (based on Ando. 1975). (c) Historic record of $M_s > 7.5$ earthquakes in the Japan trench and Boso-Oiso region since 1850. The 1953 event may not have a thrust mechanism. The 1933 Sanriku normal-faulting event is not included.

but the only body wave analysis in the region was done by YOSHIYAMA (1950) (Table 1). Yoshiyama compared seismograms of the 1946 Nankaido event with one of its aftershocks and inferred that the main event had three pulses. The initial P-wave amplitudes were not as large as expected for the magnitude of the event ($M_s=8.2$), being about the same as for the aftershock ($M_s=7.2$). The third pulse appeared to be the largest. No foreshocks are reported for the 1944 or 1946 events, and there was a period of quiescence for 20 years preceding the two events (MOGI, 1969a). KANAMORI (1972a) considered the relatively sparse aftershock activity of these events to reflect homogeneity of stress on the fault plane.

The convergence between the Pacific and Asian plates along the Japan trench also has an extensive recorded seismic history (Fig. 5c). There is less regularity of great earthquake occurrence compared with southwestern Japan, and a significant north to south decrease in rupture zone dimensions, which historically do not exceed 150 km (UTSU, 1974). NISHENKO and MCCANN (1979) noted that the size and distribution of terraces along the inner wall of the Japan trench tend to correlate with tsunami source areas and rupture zones, reflecting the variation in subduction process along the trench. KELLEHER and MCCANN (1976, 1977) suggested that this variation is associated with subduction of seamounts along the southern portion of the trench from 35.5° to 37.5°N. Segmentation of this arc is also indicated by CARR *et al.* (1973) who used deep and intermediate depth earthquakes to delineate transverse discontinuities in the subducting slab. Their arc segments of 100 to 300 km in length tend to coincide with rupture zones of great earthquakes such as the 1938, 1952, and 1968 events. KELLEHER *et al.* (1974) noted that the width of interface along the trench is rather wide but larger earthquake ruptures may not develop due to weak coupling on the contact. KANAMORI (1971a) associated low seismic velocities under Japan with high temperatures and partial melt indicating zones of weakness on the plate contact. KANAMORI (1971a, 1972b, 1977) proposed that there is increased decoupling on the subduction interface toward the south from Hokkaido, which accounts for the decrease in great earthquake occurrence. South of the Japan-Izu Bonin trench junction large earthquakes are infrequent and do not exceed $M_s=7.2$ in the seismic record.

The Boso region of the southern Japan trench has been the source area of several large historic tsunamigenic events in 1605, 1703, 1923, and 1953 (IIDA *et al.*, 1967; SOLOVIEV and GO, 1974). The 1703 and 1923 events have been interpreted as the result of low-angle right lateral faulting with a thrust component along the Sagami trough (KANAMORI, 1971c; ANDO, 1971; MATSUDA *et al.*, 1978). The 1703 earthquake ruptured a zone about 200 km long, while the 1923 event ruptured over a length of 130 km. These rupture zones overlap, however the recurrence interval for the region is estimated at greater than 800 years by MATSUDA *et al.* (1978). This estimate is based on the height of Holocene terraces in the Southern Boso peninsula, as well as on the lack of any record of previous events in the last 1,000 years. The 1923 event has been studied in some detail, with an average slip inferred from seismological data of 2.1 m (KANAMORI, 1971c). This is only 1/2 to 1/3 of the slip estimated from geodetic data (ANDO, 1971, 1974; MATSU'URA *et al.*, 1980), which indicates that a component of aseismic slip may have occurred preceding or during the 1923 event. SCHOLZ and KATO (1978) and THATCHER and RUNDLE (1979) have made further studies of the geodetic data for this event. A large earthquake occurred in 1605 further eastward along the Sagami trough, which may also partially overlap the 1703 rupture zone (HATORI, 1975).

The November 26, 1953 Boso-Oki earthquake ($M_s = 7.5$) was found to be a multiple rupture of three shocks by USAMI (1956) who studied P-wave arrival times in Japan. This event has been interpreted as hinge faulting within the Pacific plate with predominantly normal faulting on a fault plane striking northwest and dipping steeply to the southwest (ANDO and SENO, 1981). This mechanism may arise due to disruption of the slab at the junction between the Honshu and Izu-Bonin arcs.

The seismic record north of the Boso region and south of the 1938 Shioya-Oki earthquakes is rather sparse. Events in 1257 ($M = 7.0$ or larger), 1677 ($M = 7.4$), and 1909 ($M = 7.7$) are the largest in this region (UTSU, 1974). The 1677 event just north of the Boso region has a tsunami source area as large as the events to the south, while other tsunami source areas in the vicinity are much smaller. Deep sea terraces are very small and discontinuous in this area (KELLEHER and MCCANN, 1977).

The zone of the 1938 sequence, which included thrust events with $M_s = 7.4$, 7.8, and 7.7, had not had a large earthquake for 800 years (UTSU, 1974; ABE, 1977). ABE (1977) studied this sequence, which also included two large normal events, and determined a seismic slip rate of .4 cm/yr compared with a plate convergence rate of 9 cm/yr, indicating that the plate motion here is largely aseismic. Abe also modeled long period P and S waves of the initial event in the sequence and found a body wave source dimension of 60–75 km, comparable with the tsunami source area and aftershock distribution. The low angle thrust events appear to have triggered intraplate normal faults, which is consistent with the model of progressive decoupling suggested by KANAMORI (1977).

Further evidence for decoupling north of the 1938 zone was given by KANAMORI (1972b) who analyzed the 1896 Sanriku tsunamigenic earthquake, and attributed the efficient generation of long period waves to weak coupling on the fault plane. KANAMORI 1971a, 1977) also relates the 1933 Sanriku normal event with lithospheric detachment and weak coupling on the main thrust contact. Previous large events in the Sanriku region occurred in 869, 1611, and 1677, but there is not a continuous north-south trending zone of thrust earthquake activity in this region.

SENO *et al.* (1980) investigated the rupture process of the 1978 Miyagi-Oki earthquake off central Honshu using surface and body wave data. They found that the rupture occurred in two stages, with discrete rupture zones defined by the body wave complexity and aftershock distribution. A gap in the aftershock distribution lies between the two zones. This complexity was interpreted as resulting from non-uniformity of the rupture character along the plate contact.

South of Hokkaido there is a greater regularity in great earthquake occurrence, with events recorded in 1677, 1763, 1856, and 1968 (UTSU, 1974). For the apparent recurrence interval of 97 years, and assuming 3 to 4 m of displacement in each event, the seismic slip rate is about 4 cm/yr, or about 40% of the convergence rate (ABE, 1977). NAGAMUNE (1971) and FUKAO and FURUMOTO (1975) studied the 1968 Tokachi-Oki event. The latter analysis of the body waves from this event indicates that it was a multiple shock with perhaps three distinct events of increasing moment. The events occurred progressively south and west of the first epicenter, and then were followed by a larger smoother dislocation on the fault which propagated in the opposite direction. NAGAMUNE (1971, 1978) identified two large sources separated by 107 km in this event, and suggested a structural boundary between the source regions. The surface waves, which were also studied by

KANAMORI (1971d), indicate the northward propagation of the smooth dislocation, and give a moment of 2.8×10^{28} dyne-cm for a fault area of 100×150 km^2 (Table 1). The observation that the initial body wave rupture areas appear smaller than the total rupture

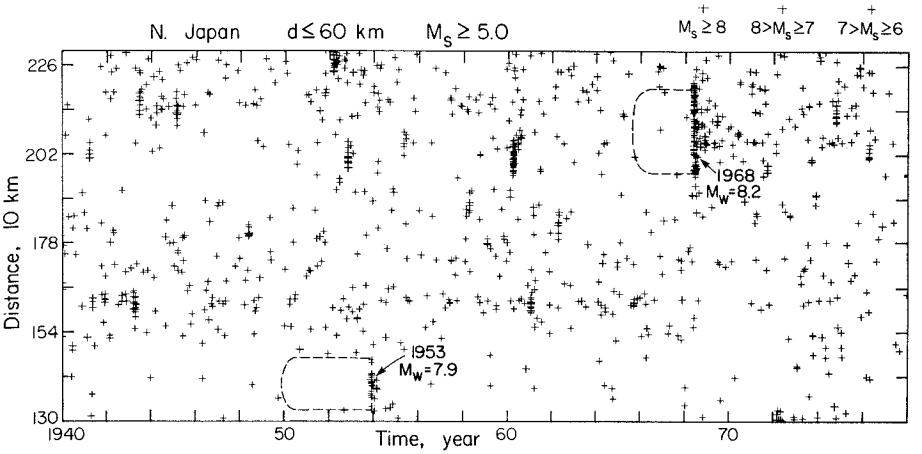

Fig. 6. Space-time plot of seismicity for northern Japan obtained from the JMA catalog. Earthquakes smaller than 5 are not plotted (from KANAMORI, 1981).

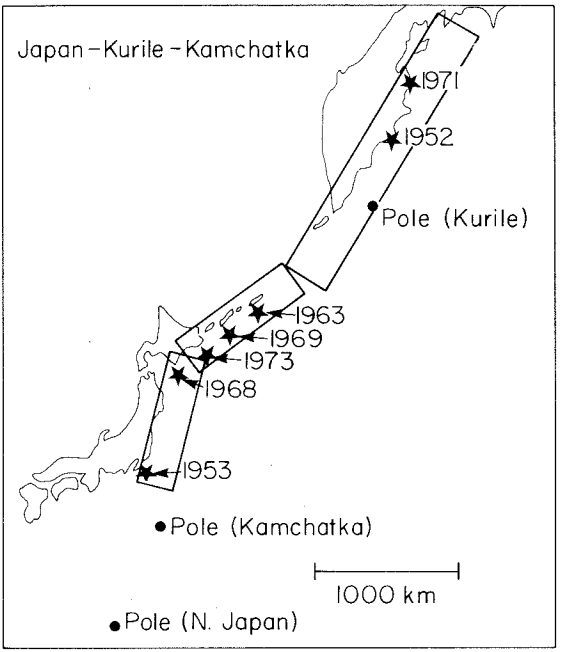

Fig. 7. Index map for the Kurile Is., Kamchatka and Northern Japan. All the events shallower than 60 km which occur in each box are shown in Fig. 6 and Fig. 9. The location of the poles is arbitrary and the scale refers to the middle of the figure, (from KANAMORI, 1981).

surface was interpreted by FUKAO and FURUMOTO (1975) as nucleation of cracks at weak spots along the fault plane subsequently coalescing into large cracks. The northern terminus of this event lies at the junction of the Kurile and Japan trenches, and the distribution and mechanism of aftershocks suggest contortion of the lithosphere at this junction (KANAMORI, 1971d; SASATANI, 1976).

There are relatively few foreshocks reported before large events in the Japan trench. Figure 6 shows a space-time plot of seismicity obtained from the catalog published by the Japan Meteorological Agency ($M_{JMA} > 5$) for the region shown by the index map in Fig. 7. The largest earthquake during the time period shown is the 1968 Tokachi-Oki event. The activity during about 3 years before the mainshock is considerably lower than during the preceding period (MOGI, 1969a; KANAMORI, 1981; HABERMANN, 1981a). No obvious foreshocks were reported for this earthquake, although NAGUMO *et al.* (1968) recorded a number of very small events during several days before the mainshock with an ocean-bottom seismograph that had been deployed in the epicentral area.

KATSUMATA and YOSHIDA (1980) have considered the seismicity in the 1968 rupture zone from 1948 to the present. They identify an elliptical area that was relatively aseismic from 1948 to 1963 and then became active 4.5 years before the main event while the surrounding zone became quiescent. This ellipsoidal area spans only 1/8 of the region eventually covered by aftershocks, and lies between the location of the two body wave sources identified by NAGAMUNE (1971). The ellipsoidal area has been practically devoid of aftershock activity. KATSUMATA and YOSHIDA (1980) interpret this as a localized stress concentration or asperity. HAMAGUCHI and HASEGAWA (1975) studied a large number of the aftershocks of this event that have very similar waveforms and concluded that these occurred at the same location under the same mechanical conditions, giving further evidence for stress concentrations along the fault.

2.3 Kurile Islands

Historical data for great earthquakes in the Kurile Islands region is also rather complete for at least two sequences of rupture of the entire trench (Fig. 8b). It is clear that there is a tendency for the large earthquakes to occur repeatedly within well-defined zones along the trench with maximum rupture length of 200 to 300 km, though there is substantial variation in repeat time. Events have tended to cluster recently, but evidence for large scale migration of activity within the trench is weak. MOGI (1968a, c) does relate the recent activity in this region to migration from further south heading northward toward the Aleutians.

Significant segmentation of the Kurile trench is generally indicated only by secondary features such as volcanic activity concentrations and seismicity. MOGI (1968b, 1969b) has discussed how the aftershocks of the 1952 Tokachi-Oki event along Hokkaido tended to remain within transverse boundaries defined by submarine-canyons for six days after the main event, before extending to adjacent areas. The southern extent of the 1952 zone abuts the Japan-Kurile trench junction, without overlapping the zone that failed in 1968. North of the 1918 and 1963 rupture zones, the Kurile trench has a less active seismic history, with a large event in 1915 with a moderate inferred rupture length of 100 km, being the only large historic earthquake. The background seismicity suggests that along this portion of the trench the plate interface narrows, which has been associated with the lack

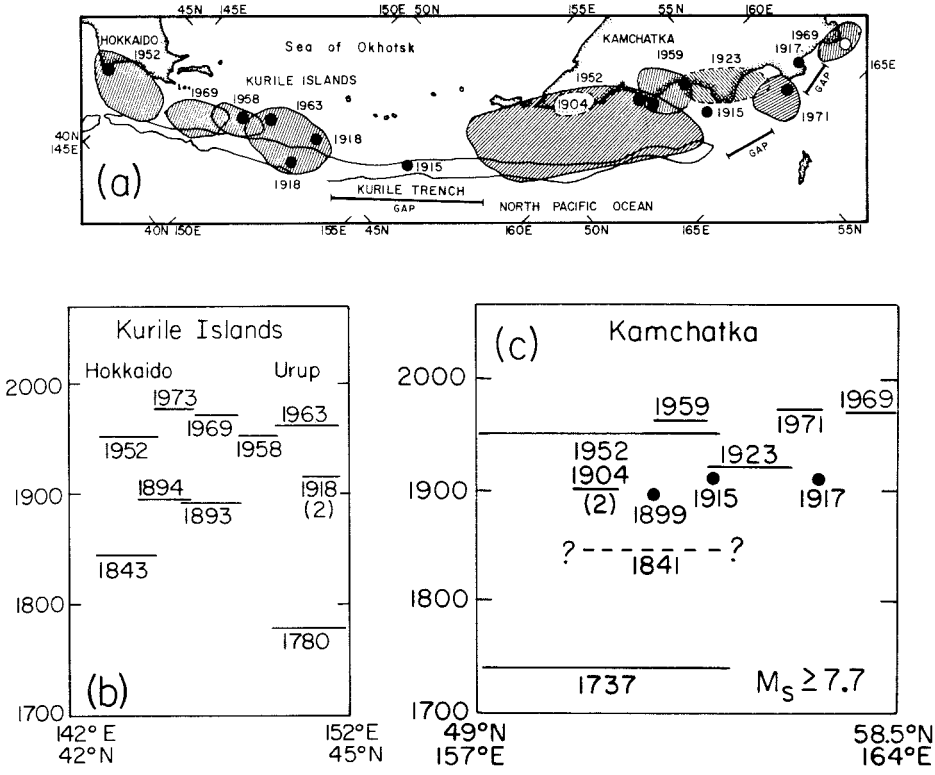

Fig. 8. (a) Map of large earthquake rupture zones in the Kurile Islands and Kamchatka since 1900. (b) Historic record of large earthquakes along the Kurile Islands. (c) Historic record of large earthquakes along Kamchatka.

of great earthquakes (KELLEHER *et al.*, 1974). The seafloor subducting along the whole length of the trench is generally smooth and devoid of rises, ridges or seamounts (KELLEHER and MCCANN, 1976). ISACKS and BARAZANGI (1977) inferred a discontinuity of the subducting lithosphere between the northern and southern Kurile Islands at about 45°N but no finer scale features could be resolved in the seismicity distribution.

Like the 1968 Tokachi-Oki event, the 1952 event did not have large magnitude foreshock activity, though there was some increase in small magnitude activity two years before the main event (KATSUMATA and YOSHIDA, 1980). However, further north, the 1963 and 1969 events did have pronounced preseismic activity, and the 1973 Nemuro-Oki event had several recorded foreshocks. KANAMORI (1978b) interpreted this increase in foreshock activity as reflecting the difference in mechanical coupling along the subduction zone. The transition to aseismic slip toward the Japan trench apparently suppresses foreshock activity. The space-time plot of seismicity for the Kurile Islands region using data from the NOAA catalog is given in Fig. 9. The index map is shown in Fig. 7. This seismicity plot shows that the 1963 event ($M_w = 8.5$), 1969 ($M_w = 8.2$) and possibly the 1973 ($M_w = 7.8$) events were preceded by a period of quiescence. For the 1963 and 1969 events there appears to have been a period of increased activity prior to the quiescence. Thus, the seismicity

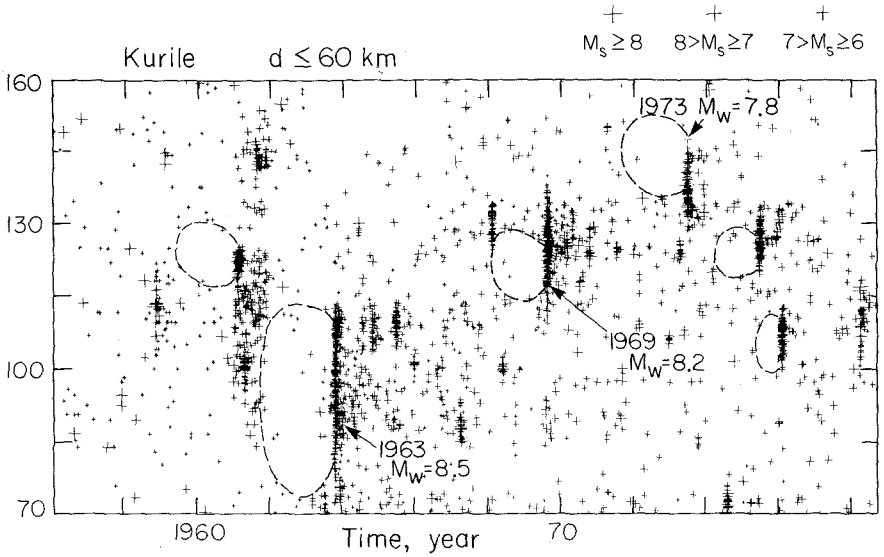

Fig. 9. Space-time plot of seismicity for the Kurile Islands region obtained from the NOAA catalog, (see Fig. 7 for location). Dotted curves encircle the period of quiescence. The magnitude range for the larger symbols are indicated in the figure. The magnitude range for the events smaller than 6 are not indicated, but the size of the symbols is approximately proportional to the magnitude (from KANAMORI, 1981).

pattern associated with large events in this region appears to be one of swarm-quiescence-foreshocks (KANAMORI, 1981).

The details of the seismicity prior to and following the 1952, 1963, 1969, and 1973 events have been studied by KATSUMATA and YOSHIDA (1980). In each case they identify a core region similar to that found for the 1968 Tokachi-Oki event, in which there is a localized zone of quiescence before and after the main event. This zone also has a precursory increase in activity before the mainshock while the surrounding zone becomes quiescent. From these observations they define a five stage process in the large earthquake cycle. The core is considered to be a localized zone of homogeneous strain field, or an asperity, with dimensions that are a small fraction of the ultimate rupture zone.

The seismic slip rate determined from studies of each of the large Kurile events is about 2.5 cm/yr (KANAMORI, 1977), only 1/4 of the plate convergence rate of about 9 cm/yr (MINSTER *et al.*, 1974). This indicates that a large amount of aseismic slip is occurring, which is consistent with crustal deformation in Hokkaido which indicates only 2.7 cm/yr of strongly coupled convergence (SHIMAZAKI, 1974a).

The occurrence of tsunami earthquakes along this trench also indicates weak coupling on the fault contact. SHIMAZAKI (1975) analyzed a very low stress drop event that occurred in 1968 within the fault zone of the 1969 event. TAKEMURA *et al.* (1977), SHIMAZAKI and GELLER (1977), and FUKAO (1979) have analyzed another slow event in this region which occurred on June 10, 1975, and an aftershock of the 1963 event which occurred on October 20, 1963 with slow earthquake characteristics. These latter two events are believed to have resulted from rupture branching up into the sedimentary wedge, loaded by previous large thrust earthquakes (FUKAO, 1979). The body waves of these events are very complicated

relative to normal foreshock and aftershock pulses.

The rupture process of the 1952 earthquake has not been extensively studied, but appears to be a complicated rupture from inspection of body wave arrivals. A long period mechanism for this event has been reported by KASAHARA (1975). The smaller 1973 Nemuro-Oki event north of the 1952 event has been more carefully studied by SHIMAZAKI (1974b). This event was smaller than the 1894 event that preceded it in the same region, and there is some question whether the entire gap between the 1952 and 1969 zones failed. The 1894 event was associated with four small foreshocks 5 to 17 hours before the main shock, and 3 large events of $M_s \geqslant 7.0$ during the three years before the main shock. The 1973 event appears to have been larger than the foreshocks of the 1894 event, but is clearly not as large as the main event.

The 1969 Kurile event was studied by ABE (1973) who used surface waves G_4 and R_4 to constrain the source, and by FUKAO and FURUMOTO (1975) who investigated the body wave complexity. The surface waves indicate a rupture velocity of 3.5–4.5 km/sec with failure spreading bilaterally over a fault area of 180×85 km^2. The source area is inferred from aftershocks and tsunami source. FUKAO and FURUMOTO (1975) showed that the 1969 event was not a multiple event, its initial impulsive break corresponding to the main faulting. However, the intense foreshock activity for this event indicates a general heterogeneity of stress on the fault plane.

FUKAO and FURUMOTO (1979) investigated the 1958 Etorofu event with surface wave analysis, and determined that it had a high stress drop relative to other large events along the arc. This event seems to have occurred in a portion of the trench that had not failed in the previous sequence of activity. The large m_B (8.2) of this event is associated with the large stress drop, and though detailed body wave analysis has not been done, it appears that the rupture initiated with a substantial failure, as did the 1969 event. The unusually high stress drop is interpreted as indicating that the fracture strength along the trench varies substantially, which yields the variable recurrence interval observed. The aftershock region of this event slightly overlaps that of the 1963 and 1969 events, but the main shock and aftershocks are slightly deeper (50–90 km), thus the overlap may not be significant (KELLEHER and SAVINO, 1975).

The great October 13, 1963 ($M_w = 8.5$) earthquake was studied with surface wave analysis by KANAMORI (1970a), who found that rupture propagated northeast over a fault length of 200 to 300 km. The body waves of this event are very complex, and few are on scale (FUKAO, 1979). NAGAMUNE (1971, 1978) identified at least two sources in the S wave complexity for this event with the strongest rupture occurring 37 seconds after the main event. The second source was located 130 km northeast of the epicenter by body wave and Love wave timing. He also inferred the presence of a structural boundary between the source regions. Several of the teleseismic P waves have been modeled by RUFF and KANAMORI (1980b), who find that the source time function is characterized by five or more discrete ruptures with durations of 10 to 30 sec, with corresponding source dimensions of ~ 50 km each.

The northeastern end of 1963 fault zone lies along the Bussol Strait. North of this region lies the portion of the trench with limited historic record. WYSS and HABERMANN (1979) have identified a seismicity decrease in this region which commenced in 1967 and terminated in 1971 (HABERMANN, 1981a). This quiescent period has been associated with

an aseismic slip event by HABERMANN (1981b).

2.4 Kamchatka

While almost all recent large earthquakes near Japan and in the Kurile islands have been studied, less attention has been paid to the vicinity of Kamchatka. This region demonstrates substantial variation in rupture mode, with large and small earthquakes contributing to rupture of the trench. The seismic activity since 1900 apparently constitutes one rupture sequence for the entire region, as indicated in Fig. 8(c). Though many rupture zones appear to overlap, they actually fill in the rather broad plate interface. The 1952 event ruptured the zone where the interface width is greatest, being bounded on the southwest by an appreciably narrower contact zone (KELLEHER *et al.*, 1974). The February 3, 1923 earthquake ($M_w=8.3$) was found to have a tsunami magnitude of $M_t=8.8$ (ABE, 1979), thus this event may have ruptured most of the northern part of the trench. The April 13, 1923 event in the same region was inferred to be a tsunami earthquake by ABE (1979) from the discrepancy between M_s (7.2) and M_t(8.2). The tsunami source areas of these two events are shown to cover the entire trench north of the 1952 zone to the 1971 source region where the extension of the Aleutian arc along Komandorski Island intersects the trench (SOLOVIEV and GO, 1974). Other recent events appear to be substantially smaller, and tend to fill in the convergent zone with less spatial regularity than observed in the Kuriles. The aftershock zone of the 1952 event has something of a dumbell shape, wrapping around the rupture zone of the 1904 events (KELLEHER and SAVINO, 1975), and the 1959 and 1971 events fill in areas along the margins of the larger 1923 and 1952 events.

Large earthquakes are reported to have occurred in the vicinity of the 1952 event in 1737 and 1841 by FEDOTOV (1965). He indicates a relatively small rupture area for the 1841 event lying just northeast of the 1904 events, and does not show the extent of rupture in the 1737 event. IIDA *et al.* (1967) documented the evidence for a very large tsunami on October 17, 1737, generated in southeast Kamchatka. From various early references they inferred that there was a sequence of three earthquakes during 15 min of shaking, culminating with a large event sending 30 m tsunami waves into the Kurile and Kamchatka regions. The tsunami source area for this event given by SOLOVIEV and GO (1974) overlies the 1952 zone, but their work gives no source estimate for the 1841 event, probably due to the sparse data on local tsunami of that event. The possibility that the 1841 ($M_s=8.4$) event was as large as the 1952 ($M_s=8.3$, $M_w=9.0$, $M_t=9.0$) event is indicated by ABE's (1979) determination of its tsunami magnitude, $M_t=9.0$. This large value is based principally on a large tsunami record at Hilo, Hawaii of 4.6 m, though most Kamchatka records indicate a local tsunami of only 1 m. SOLOVIEV (1978) does document a 15 m tsunami report for the 1841 event in Kamchatka, which is consistent with the large $M_t=9.0$. On the basis of intensity data, KONDORSKAYA and SHEBALIN (1977) assign a magnitude of 8.4 to the 1841 event, which is larger than their value of 8.3 for the 1737 event.

For the 1952 Kamchatka event, KANAMORI (1976) determined a moment of 3.5×10^{29} dyne-cm, and assuming a fault area of 200×650 km^2, he obtained a 5 m seismic displacement. Use of a slightly smaller rupture zone (200×500 km^2) indicated by aftershock relocations given by KELLEHER and SAVINO (1975) yields a displacement of 7 m. Assuming that the 1841 event did not rupture the 1952 zone, while the 1737 did, yields a recurrence period of 216 years and an inferred seismic slip rate of 2.3–3.2 cm/yr. The plate convergence

rate along Kamchatka given by MINSTER *et al.* (1974) ranges from 9.0 cm/yr at the southern end of the 1952 rupture zone to 8.6 cm/yr near the 1971 earthquake zone. If the 1841 rupture did cover the 1952 segment, the average recurrence interval is 108 years which gives a seismic convergence rate of 4.6–6.4 cm/yr. Thus, the amount of aseismic slip could range from 1/3 to 2/3 of the total convergence motion. The shorter recurrence interval would be consistent with that observed in the Kuriles and western Aleutians, and tsunami evidence does tend to favor a substantial rupture zone for the 1841 event. The occurrence of the 1923 tsunami event is consistent with KANAMORI's (1977) association of such events with regions of substantial aseismic convergence, though it is not clear whether this event occurred on the plate interface or in the sedimentary wedge like the Kurile tsunami earthquakes.

Detailed source studies of Kamchatka events are lacking as indicated in Table 1. The long period surface waves of the great 1952 event were studied by BEN-MENAHEM and TOKSOZ (1963) and KANAMORI (1976), yielding long period focal mechanisms and evidence for southwestward rupture propagation. This is consistent with the epicenter location in the northern lobe of the rupture zone. The event was preceded by three years of preshock activity clustering in the epicentral region and near the southwestern end of the zone, while there was a general quiescence over the remainder of the future rupture area (KELLEHER and SAVINO, 1975). The clustering began about three years prior to the main event, but poor depth resolution prevents definite location of these events on the fault plane. Inspection of the Pasadena strain records for this main shock indicates an S-wave complexity comparable to that of the 1960 Chile rupture, suggestive of a multiple source event, but no further body wave details are known.

Somewhat similar foreshock activity was observed for the December 15, 1971 (M_s= 7.8) earthquake, which ruptured over a 100 km zone. This event was preceded by few earthquakes in the rupture zone from 1960 to 1970, with a cluster of 17 events occurring in 1969, then quiescence until the rupture (WYSS *et al.*, 1978). The cluster with magnitudes ranging from 4.3 to 5.5 occurred over a well-defined area with a 40 km diameter about 36 km from the epicenter. These events were noted to have low ratios of P wave to S wave energy by FEDOTOV *et al.* (1977).

2.5 *Aleutians*

The historic record of large events in the Aleutians is indicated in Fig. 10. Recent activity consists of the extensive ruptures of the 1957 and 1965 earthquakes, while the previous failure sequence of events around the turn of the century had smaller individual rupture lengths (SYKES, 1971; SYKES *et al.*, 1981). This temporal variation in rupture mode may shed some light on the nature of large earthquake occurrence. SYKES *et al.* (1981) suggested that the turn of the century earthquakes commenced with smaller initial events than the 1957 and 1965 ruptures, thus failing to trigger complex rupture over more extensive area. There is ample evidence for transverse segmentation of the arc which may also influence the variable rupture history.

The main thrust interface along the Aleutian trench on which great earthquakes occur extends to a depth of 40 km and is about 120 km wide in the west and 150 km wide along the Alaskan peninsula (DAVIES and HOUSE, 1979). From east to west there is an increasing component of strike slip in the convergence direction, which may increase the effective

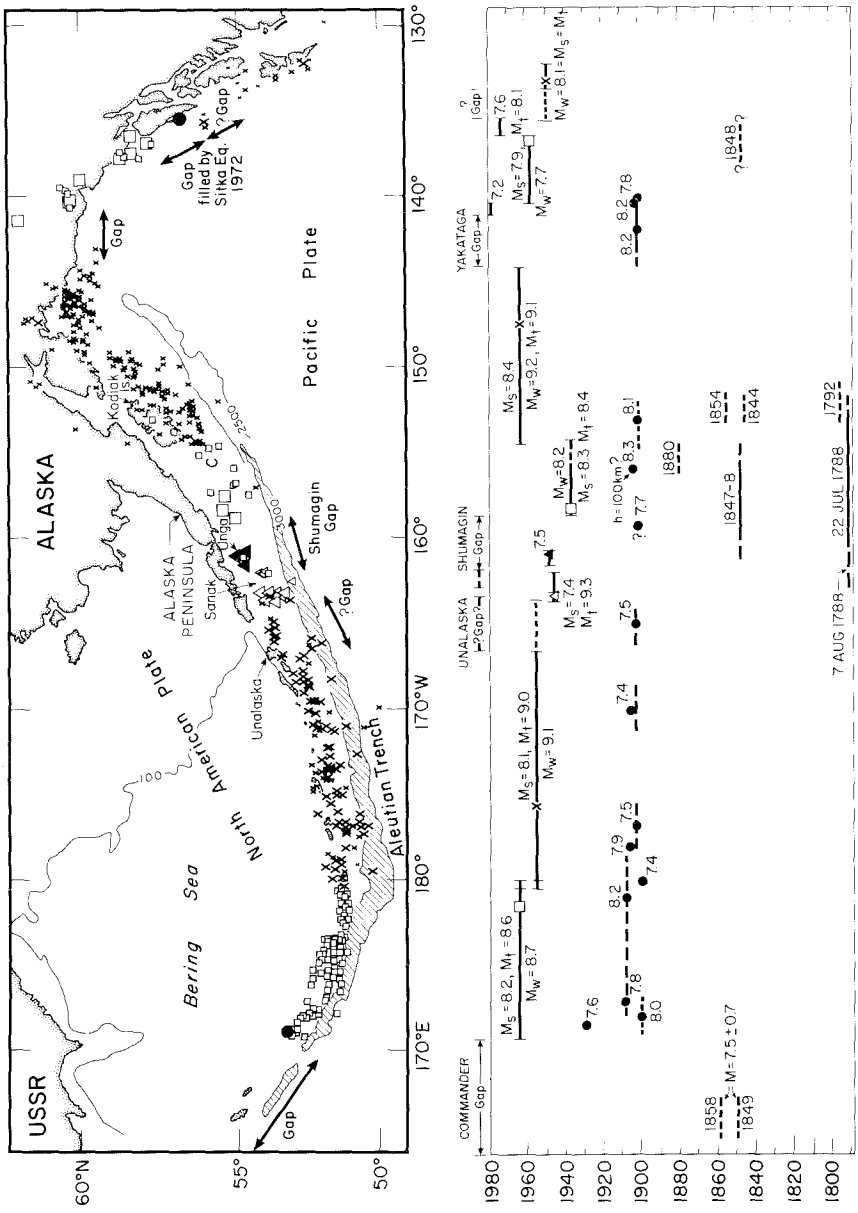

Fig. 10. Above: Rupture zones of earthquakes of magnitude $M > 7.4$ from 1925–1971 as delineated by their aftershocks along the Aleutian and Alaska zone. Contours in fathoms. Various symbols denote individual aftershock sequences as follows; crosses, 1949, 1957, and 1964; squares, 1938, 1958, and 1965; open triangles, 1946; solid triangles, 1948; solid circles, 1929, 1972. Below: Space-time diagram showing lengths of rupture zones, magnitudes and mainshock locations for known events of $M > 7.4$ from 1784–1980. Large normal-faulting events in 1929 and 1965 are ommitted, (from SYKES *et al.*, 1981).

contact width (KELLEHER *et al.*, 1974). The component of strike slip motion is largest along the Komandorski island region, where the trench is poorly developed. This area was the site of large events in 1849 and 1858, but did not have earthquakes during the turn of the century sequences to the east, and is thus labeled the Commander Gap by SYKES *et al.* (1981).

The Aleutian terrace, while not extending into the Commander Gap, is the largest in the world, being virtually continuous along the rupture zones of the 1957 and 1965 earthquakes. NISHENKO and MCCANN (1979) suggested that the greater continuity of the terrace along the 1957 rupture zone relative to the western extension may indicate that the eastern portion of the trench fails in great earthquakes more frequently than the Rat Island region. The previous sequence in the 1957 zone, which occurred in 1901 and 1902 had relatively small events ($M_s = 7.5$), but these occurred only 20 hours apart indicating that there may have been some interaction between them.

There is relatively smooth ocean floor seaward of the Aleutian trench, and an unbroken volcanic line along the arc, except where the Emperor seamounts converge with the arc along the Commander Gap near the junction with the Kamchatka trench (KELLEHER and MCCANN, 1976). SYKES (1971) associated the occurrence of large ruptures with relatively simple tectonic regions as in the Aleutians. There are, however, major transverse canyons in the trench and lineations in the overriding Aleutian Basin which may segment the seismic zone.

The trend of the trench is offset by 40° near the Amchitka Pass, which abruptly separates the 1957 and 1965 rupture zones (JORDAN *et al.*, 1965; MOGI, 1969b; SYKES, 1971). The rupture zone of the 1957 earthquake delineated by aftershocks is almost split near the intersection of the Amukta Pass and the trench. There is a slight change in strike of the trench near this junction also. This boundary appears to have interfered with aftershock migration following the 1957 event, apparently stalling the rupture propagation (MOGI, 1968b). The northeast margin of the 1957 zone terminates near the Alaska peninsula.

The seismicity preceding the 1965 and 1957 events is shown in Figs. 11b and 11c, with a base map given in Fig. 11a. There is a distinctive increase in seismicity over several years prior to the main events, and significant foreshock activity before the 1965 event. This kind of increase in activity before large earthquakes concentrated near either end of the rupture zone has been pointed out by KELLEHER and SAVINO (1975). There is some indication of a precursory swarm of magnitude $M_s > 6$ events from 1928 to 1936 in the 1957 rupture zone. There was substantial smaller magnitude swarm activity in the central portion of the eastern lobe of the rupture zone prior to the 1957 event, with a large amount of activity distributed along the entire rupture zone at depths of 25–40 km. This localized downdip activity may suggest that the coupling is strongest at these depths and weaker beneath the trench (HOUSE *et al.*, 1981). This event ruptured over a length of about 1,200 km, though the exact extent of rupture is uncertain (DAVIES *et al.*, 1981). Unfortunately, detailed body wave and surface wave analysis of this event has not been performed. HATORI (1981) has determined a tsunami source area with a 900 km length for the 1957 event, and using the tsunami magnitude he estimated a moment of $1.5 \cdot 10^{29}$ dyne-cm.

The 1965 rupture nucleated near the Amchitka Pass on the margin of the 1957 rupture. ABE (1972c) inferred from P wave travel time residuals of the Longshot nuclear explosion

on Amchitka Island that the subducting lithosphere underlying this junction is torn, effectively decoupling the western and central Aleutians.

The 1965 Rat Island earthquake appears to have ruptured three or four large block segments bounded by transverse canyons in the overriding plate that appear to be tectonically controlled (MOGI, 1969b; SPENCE, 1977). The rupture rapidly extended over the entire zone and then the zone of activity broadened as suggested by the intensive aftershock distribution (MOGI, 1968b). The aftershocks roughly outline large tectonic blocks, which are also defined by geologic structures as summarized by SPENCE (1977). The Aleutian terrace is relatively fragmented west of the Amchitka Pass, and there are distinctive geologic, bathymetric, and seismic patterns outlining these structures, indicating that they are tectonic rather than erosional features. The turn of the century sequence of smaller events indicates that these blocks may have ruptured independently.

WU and KANAMORI (1973) analyzed the Rat Island earthquake and inferred that it was a complicated rupture. Relative timing of body wave arrivals indicates that successive ruptures in the first 21 sec of failure occurred progressively south of the initial epicenter, with larger rupture propagation westward being indicated by the surface waves. Only the first few seconds of the body waves could be analyzed. RUFF and KANAMORI (1980b) have analyzed several diffracted P waves from this event recorded by WWSSN instruments and

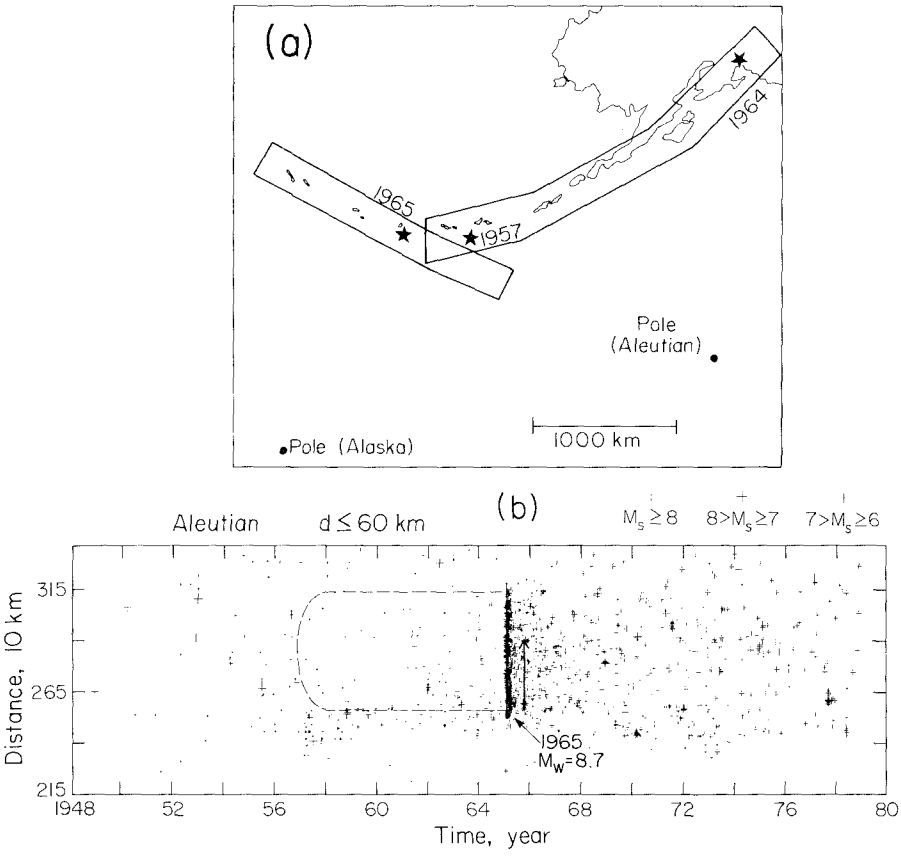

Fig. 11

have found that the source time function for the first 120 sec of rupture is represented by two distinct pulses of about 60 sec duration each. The very long duration of these time functions is indicative of rather homogeneous stress release over a large region during failure.

On the other hand, the pronounced foreshock and aftershock activity and complex rupture process of the Rat Island event may indicate a complicated stress distribution. SPENCE (1977) suggested that the oblique convergence direction may have facilitated rupture transmission across the tectonic blocks, though this would also have been effective in the turn of the century sequence when discrete ruptures occurred. Spence also argued that the greatest aftershock activity occurred in the vicinity of the north trending fracture zones, with the topographic irregularities providing high energy release elements, and mid-portions of the blocks being lower energy release elements. This may be supported by the lack of large thrust aftershocks in the central area of the Rat Island block, and by the aftershock distributions reflecting the tectonic subdivision. Larger scale interactions between tectonic blocks may be indicated by the east to west propagation in the Alaska-Aleutians region discussed by KELLEHER (1970). It is tempting to associate the discrete long duration pulses in the 1965 Rat Island source time function found by RUFF and KANAMORI (1980b) with failure of these discrete blocks. If this is the case, it suggests that

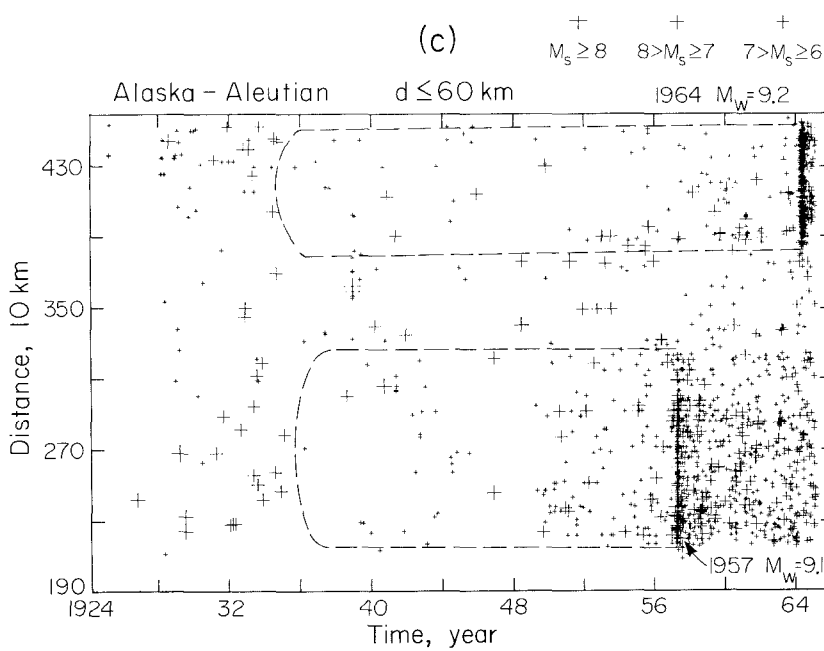

Fig. 11. (a) Index map for Alaska and the Aleutians. See the caption of Fig. 7 for details. (b) Space-time plot of seismicity for the rupture zone of the 1965 Rat Island earthquake. For details, see the caption of Fig. 9. (c) Space-time plot of seismicity for the rupture zones of the 1964 Alaskan and the 1957 Fox Island earthquakes. For details, see the caption of Fig. 9 (from KANAMORI, 1981).

the mid-portions of the blocks were in fact the sites of much of the energy release, with uniform rupture and stress release accounting for the lack of aftershocks in these regions.

ABE (1972a) studied a large normal fault event which occurred on March 30, 1965 trenchward of the Rat Island mainshock. On the basis of the aftershock area and large size of this event Abe concluded that it ruptured throughout the whole lithosphere. The main thrust event in 1965 may have temporarily decoupled the plate contact allowing the subducting slab to develop tensional stresses leading to the normal fault event for that case as well as for normal fault mechanism aftershocks of the 1957 Aleutian and 1964 Alaskan events. The large 1929 Aleutian normal fault event in the eastern lobe of the 1957 rupture zone was not preceded by a large thrust event, nor was the 1933 Sanriku earthquake. Abe suggested that the coupling on the interface contact is much stronger in the Aleutians than in Japan, and though the large normal fault events indicate localized decoupling, this decoupling is transient in the Aleutians.

KANAMORI (1972b) analyzed the 1946 Aleutian tsunami earthquake ($M_s = 7.4$, $M_t = 9.3$) which occurred east of the 1957 rupture zone. The tsunami source area of this event is 400 km long (HATORI, 1981), and the moment estimated from the tsunami magnitude is $1.5 \cdot 10^{29}$ dyne-cm. This source region is larger than that defined by the aftershock area, which has a length of 125 km (SYKES, 1971). The very long time constant source mechanism of this event may be associated with weak coupling on the thrust plane. This event is located in the transitional zone between the Aleutian trench and the Alaskan peninsula, a region of small rupture zones. This area is a relic triple junction, and is distinctive in having the Unimak seamount lying on the inner margin of the trench. The occurrence of the tsunamigenic event adjacent to the large 1957 rupture indicates that the nature of coupling on the fault plane varies along the arc.

The average plate motion between the Pacific and North America at the Rat Island epicenter (178°E, 51°N) is 7.5 cm/yr in a N50°W azimuth (MINSTER *et al.*, 1974). The subduction rate normal to the trench at 178°E is 4.7 cm/yr decreasing to zero westward along the arc. For the 1965 event, WU and KANAMORI (1973) obtained an average displacement of 2.5 m in a direction N51°W. For a recurrence period of 60 years this would indicate a seismic convergence rate of 4.2 cm/yr, or 56% of the plate rate. SPENCE (1977) argued that the average displacement is greater, on the order of 12 m, based on restricting the aftershock zone width to the subterrace zone. His number would completely accommodate the plate convergence seismically even for a recurrence interval of 150 years. Similarly, much, if not all, of the convergence in the 1957 rupture zone can be accommodated seismically if the recurrence period is as short as 60 years, assuming an average slip comparable to the 1965 event.

2.6 Alaska

The subduction regime south of Alaska is also the site of extensive historic ruptures, notably the 1938 and 1964 events (Fig. 10). Along the Alaskan peninsula the Aleutian trench shallows and the plate interface broadens. Maximum rupture lengths are on the order of 50 to 100 km (1946, 1948), or 100 to 250 km (1938), though there is some indication that more extensive ruptures of up to 600 km length have occurred as in 1847 and 1788 (SYKES *et al.*, 1981). This region may have also failed between 1899 and 1903, indicating an average repeat time of 50 to 75 years, though the lack of tsunami data for the 1847 event,

and uncertainty in rupture extent of many early events precludes a more definite determination. Details of the 1938 event are not known, though the aftershock area (250×100 km^2) and tsunami source region are fairly accurate (DAVIES *et al.*, 1981). Using estimates of the recurrence interval (40 yr) and the seismic moment of the 1938 event based on fault area (3.5×10^{28} dyne-cm), 100 sec surface waves (2.1×10^{28} dyne-cm) and the tsunami magnitude of $M_t = 8.4$ (5.0×10^{28} dyne-cm), DAVIES *et al.* (1981) infer that this part of the trench converges without aseismic slip.

The 1964 earthquake ruptured on a shallow (9°) dipping thrust fault over a region 360×750 km^2. The rupture zone is bounded on the southwest by transverse structural boundaries on the Alaskan peninsula (BURK, 1965; MOGI, 1969b; SYKES, 1971), and on the northeast near the transition to the strike slip regime along the Queen Charlotte-Fairweather fault zone. The ocean floor has much more topography than along the Aleutian trench, being covered with the Gulf of Alaska seamounts, and a thick accretionary prism.

Seismic activity prior to the 1964 event clustered near the epicenter in the northeastern end of the zone and at the southern terminus of the rupture (TOBIN and SYKES, 1966; KELLEHER and SAVINO, 1975). In Fig. 11c it is clear that there was a prolonged period of increased activity in the rupture zone of the 1964 event over 10 years preceding the main shock. There also appears to have been a cluster of activity prior to 1936, similar to that before the 1957 Aleutian event. KATSUMATA and YOSHIDA (1980) identify a large region in the northeastern end of the rupture zone which was deficient in foreshock and aftershock activity, which they associate with the core region in their model.

WYSS and BRUNE (1967) detected seven subevents in the body wave phase of the 1964 earthquake, with epicenters progressing from 35 to 250 km southwest from the epicenter with an apparent rupture velocity of 3.5 km/sec. KANAMORI (1970b) studied the surface waves of this event and found that they were consistent with a southwestward rupture propagation. In a more detailed analysis, RUFF and KANAMORI (1980b) have modeled diffracted P waves from this event, and find that the source time function has a very long duration of >120 sec. This unusual time function suggests very uniform rupture over a large region, which in turn suggests a fairly uniform state of stress on the fault plane and strong coupling. This long duration time function indicates a source dimension of several hundred kilometers, which approximately corresponds to the anomalous zone identified by KATSUMATA and YOSHIDA (1980).

KANAMORI (1977) discussed the uncertainty in the recurrence time for this region, concluding that a repeat time of 200 years would indicate purely seismic slip and strong coupling, whereas longer recurrence intervals of 500 to 1,350 years suggested by uplifted terraces (PLAFKER, 1969, 1972; PLAFKER and RUBIN, 1978) would indicate a large amount of aseismic slip. SYKES and QUITTMEYER (1981) obtained repeat times of 187 and 623 years for Alaska, assuming 100% and 30% ratios of seismic slip to total convergence, respectively.

The region northeast of the 1964 zone appears to have ruptured in two great earthquakes in 1899 in what is now identified as the Yakataga gap (MCCANN *et al.*, 1979). This region has a complex tectonic behavior with both subduction and strike slip faulting (PEREZ and JACOB, 1980). The 1979 St. Elias earthquake ruptured a portion of this gap on one of the shallow dipping, imbricate faults in the transition between the Alaskan thrust

and Fairweather strike slip zones. BOATWRIGHT (1980) analyzed long period P and S waves of this event and detected three distinct subevents propagating southwest, as well as a small initial event. The rupture lengths of these subevents indicated by pulse rise times are 12, 27, and 17 km. The total rupture area indicated by aftershocks is 60 km^2, with about 1/2 of the total area being covered by the body wave source areas. The cumulative body wave moment of 1.2×10^{27} dyne-cm is 1/4 the surface wave moment. Boatwright interpreted this complexity as barriers stopping the rupture then failing in successive events. There was a high density of aftershocks in the regions where the ruptures arrested.

Other events that have been studied along the Fairweather fault system include the 1948 bilateral rupture, the 1958 right lateral rupture that extended 350 km (KELLEHER and SAVINO, 1975; ANDO, 1977), and the 1972 Sitka event which partially ruptured a gap between the 1948 and 1958 zone. These events are largely strike slip and will not be discussed other than to mention the observation of KELLEHER and SAVINO (1975) that there is a northwestward gradation in frequency and magnitude away from the Juan de Fuca spreading center, much as along the San Andreas system north from the Gulf of California spreading sytem. They attributed this phenomena to changing lithospheric structure away from the spreading center, perhaps due to cooling and thickening of the lithosphere.

2.7 *Middle America*

The recent sequence of large earthquake activity in the Middle American trench has been investigated in some detail. Large events in this region are characterized by relatively short maximum rupture lengths (Fig. 12), ranging from 100 to 200 km in length (KELLEHER *et al.*, 1973). There is no evidence for failure of the entire trench in a single rupture in the historic record. The short length of rupture zones in this region has been associated with the relatively young age of the thin underthrusting plate (KELLEHER *et al.*, 1973) and with the narrow plate interface contact (KELLEHER *et al* , 1974). Other factors may also control the maximum rupture dimensions. CARR *et al.* (1974) cited evidence for segmentation of the underthrusting plate based on spatial volcano clustering, which loosely defines seven segments along Mexico and Central America. Large bathymetric features including the Orozco fracture zone, Tehuantepec ridge, and Cocos ridge are found on the Cocos plate, with large trench earthquakes occurring near the Cocos ridge but not near the other intersections (KELLEHER and MCCANN, 1976).

Though RIKITAKE (1976) determined a mean return period in the trench of 34.5 years with a small standard deviation of 3.6 years, it is clear from an inspection of Fig. 12, that there is substantial spatial and temporal variability in large earthquake occurrence along the trench. One of the areas with the greatest regularity is the Oaxaca region. MCNALLY and MINSTER (1981) suggested that the variation in plate convergence rate of from 5–8 cm/yr along Mexico to 7–9 cm/yr along Central America produces some of the rupture variability, with only regions near Oaxaca and Jalisco having local seismic slip rates equaling the plate convergence rate.

Figure 13 shows the space-time plot of seismicity along Mexico, which highlights the quiescence preceding the Oaxaca earthquake in 1978 ($M_w = 7.6$). This anomaly was first noted by OHTAKE *et al.* (1977), and served as the basis for their forecast of the event. The mainshock was preceded by several foreshocks which migrated toward the epicenter (PONCE *et al.*, 1977). Precursory quiescence was also reported for the 1965 ($M_w = 7.5$) and

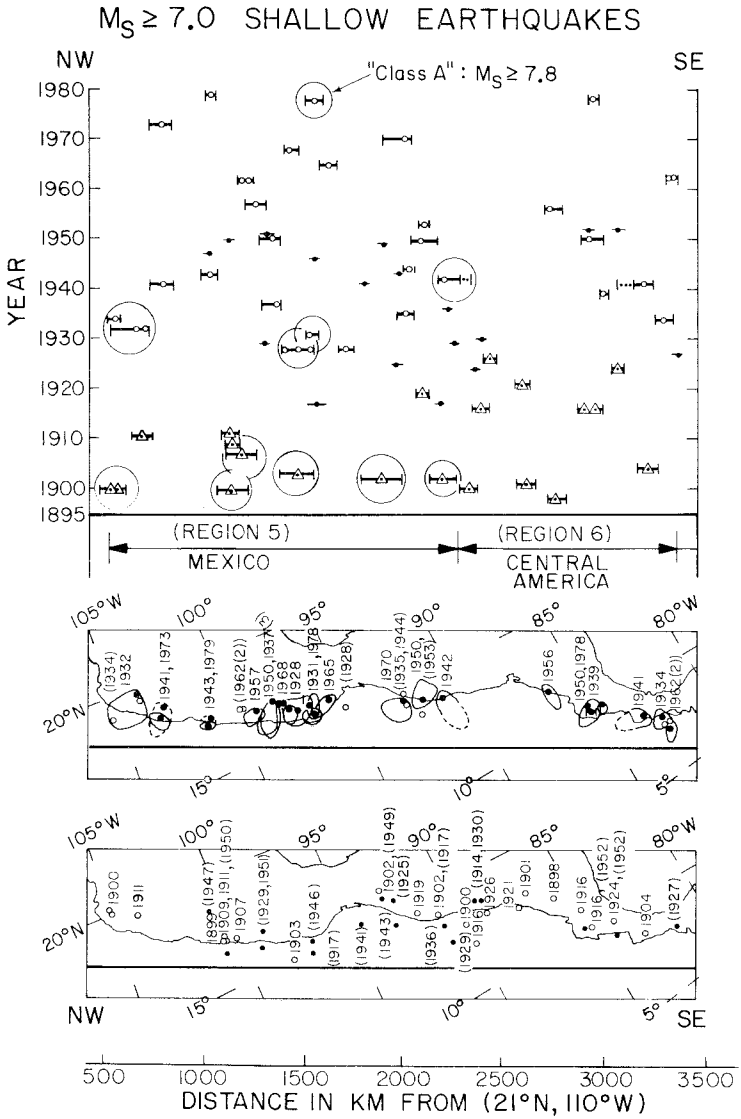

Fig. 12. Historic record of $M_s > 7.0$ shallow earthquakes along Middle America, (from McNally, personal communication).

1968 ($M_w = 7.3$) Oaxaca events by OHTAKE *et al.* (1978), though this is not obvious in Fig. 13. A detailed study by TAJIMA and MCNALLY (1981) of the spatio-temporal patterns of seismic activity before the three Oaxaca events has shown that the 1965 event was preceded by quiescence, the 1968 event by clustered foreshock activity, and the 1978 event by 43 months of quiescence during which only one event with $m_b \geqslant 4.0$ occurred within the rupture zone. Tajima and McNally attributed the difference in seismicity pattern before the 1968 and 1978 events to variation in the degree of structural heterogeneity along the

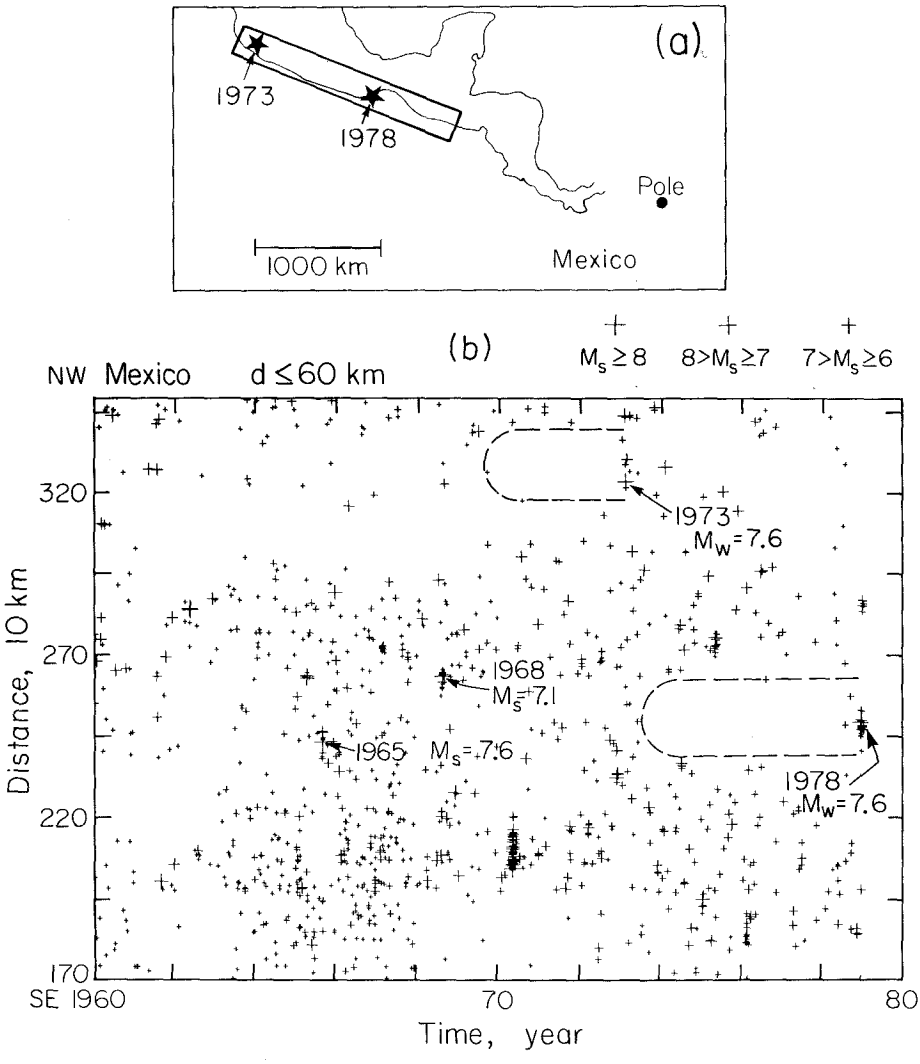

Fig. 13. (a) Index map for Mexico. See the caption for Fig. 7 for details. (b) Space-time plot of seismicity for Mexico. For details, see the caption of Fig. 9 (from Kanamori, 1981).

trench.

The source mechanisms of the 1965, 1968, and 1978 Oaxaca earthquakes have been determined by STEWART *et al.* (1981) and CHAEL and STEWART (1982). These studies analyzed the body and surface waves, and found that the events were very similar to one another. The body waves are simple impulsive signals, suggesting smoothly propagating rupture with rupture durations of about 10 sec (Table 1). The inferred body wave source areas are 1/3 to 1/2 of the aftershock areas. The 1979 Petatlan earthquake was found to be very similar to the Oaxaca events (CHAEL and STEWART, 1982). Another study of the Oaxaca body waves conducted by REICHLE *et al.* (1978) indicates that the Oaxaca

earthquake initiated with a high stress drop event over a small region, and then rupture spread into a larger area of lower average stress.

Previous large earthquakes in the Oaxaca area occurred in 1928–1931 and 1899–1903. The earliest sequence had a cumulative moment of 21.2×10^{27} dyne-cm released in two events of $M_s = 7.9$ and $M_s = 8.1$. The 1928 earthquakes resembled a swarm, involving four events of $M_s = 7.5$, 8.0, 7.4, and 7.6, and combined with the 1931 event of $M_s = 7.8$ give a cumulative moment of 20.6×10^{27} dyne-cm. STEWART *et al.* (1981) proposed that the 1928 sequence was caused by triggering of events due to contemporaneous uniform conditions of stress along the subduction zone. The cumulative moment of the 1965, 1968, and 1978 events is 8×10^{27} dyne-cm which indicates that more activity may occur. SINGH *et al.* (1980) suggested that only a portion of the Oaxaca seismic gap failed in 1978.

The rupture process of the 1970 and 1973 events appears to have been more complex than that of the other large Middle American trench earthquakes, perhaps due to their location in the vicinity of suspected triple junctions (CHAEL and STEWART, 1982). The 1970 event near the southeast coast of Mexico was studied by YAMAMOTO and MITCHELL (1978) and CHAEL and STEWART (1982). Yamamoto and Mitchell modeled the long period P waves of a double source foreshock, single source aftershock, and a complex mainshock. Four events were identified within the 15 sec before the main rupture for which the teleseismic moment was 4.6×10^{26} dyne-cm. Small rupture areas and low rupture velocity were characteristic of these earthquakes. Similar complex rupture process was found for other shallow earthquakes in this region, indicating that the stress distribution is distinct from the Oaxaca area. CHAEL and STEWART (1982) modeled the mainshock as a double event, with the total body wave moment being 1/2 of the surface wave moment.

REYES *et al.* (1979) analyzed the source mechanism of the 1973 Colima ($M_s = 7.5$) earthquake, which occurred over the same region as the 1941 event. They did not present a detailed body wave analysis, but the P wave pulses are more complicated than the Oaxaca events. Epicenter locations indicate that rupture progressed from southeast to northwest, with many aftershocks clustering at the southeastern boundary. The 10 bar stress drop of this event is relatively low, perhaps indicating moderate coupling on the fault plane. For a 5.6 cm/yr convergence rate (MINSTER *et al.*, 1974), the average slip of this event of 144 cm indicates a repeat time of 25 years, close to the actual 32 year interval, thus this area appears to converge almost entirely by seismic slip, though just to the south of this region where the Orozco fracture zone subducts, the aseismic component may be greater.

2.8 Colombia

The Colombian subduction zone has a relatively brief historic record, but appears to have a temporal variation in rupture extent similar to the Aleutian zone. Smooth seafloor north of the Carnegie ridge subducts to the east along Colombia and Ecuador (KELLEHER and MCCANN, 1976). This region is the site of the great 1906 ($M_s = 8.7$, estimated $M_w = 8.8$) earthquake which is believed to have ruptured over 500 km (KELLEHER, 1972). Subsequently, there has been a northeastward progression of events in the same zone in 1942 ($M_s = 7.9$), 1958 ($M_s = 7.8$), and 1979 ($M_w = 8.2$) as shown in Fig. 14(b). These three events appear to have covered the 1906 rupture zone with the possible exception of the northeastern corner.

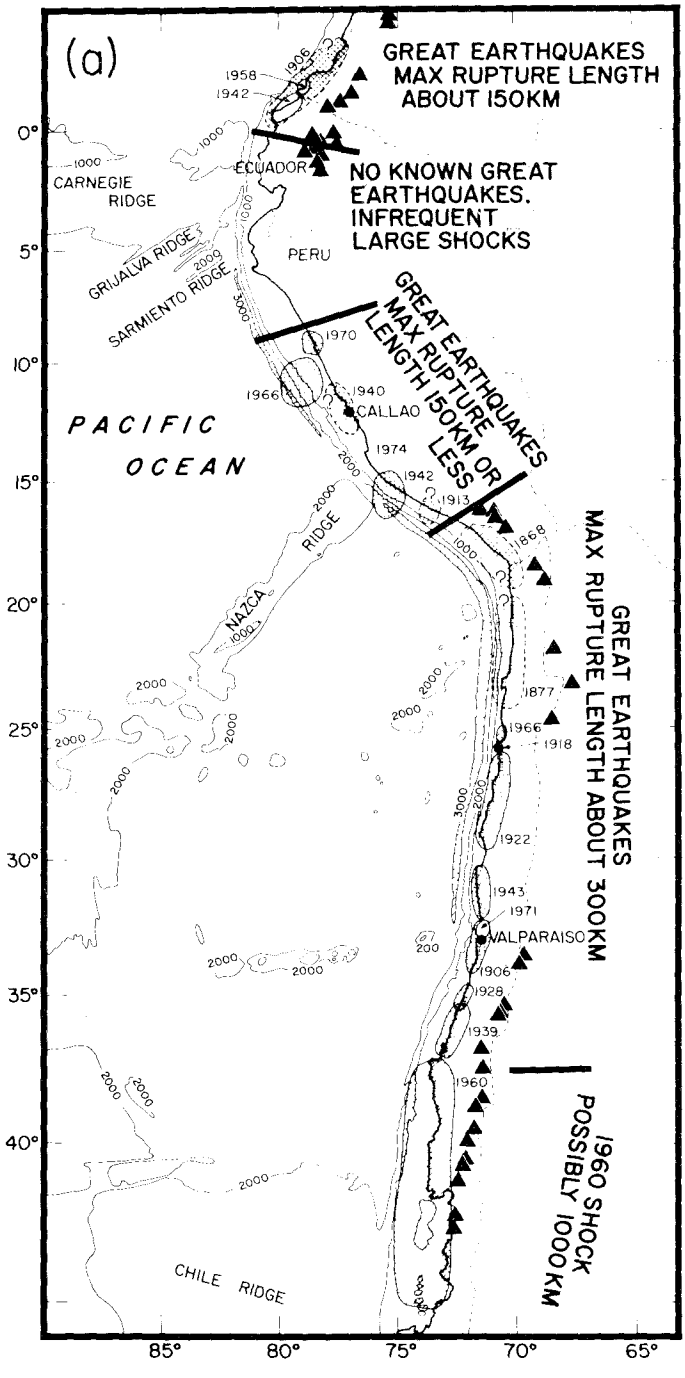
(a)
GREAT EARTHQUAKES
MAX RUPTURE LENGTH
ABOUT 150KM
NO KNOWN GREAT
EARTHQUAKES.
INFREQUENT
LARGE SHOCKS
GREAT EARTHQUAKES
MAX RUPTURE
LENGTH 150KM OR
LESS
GREAT EARTHQUAKES
MAX RUPTURE LENGTH ABOUT 300KM
1960 SHOCK
POSSIBLY 1000 KM
CARNEGIE
RIDGE
GRIJALVA RIDGE
SARMIENTO RIDGE
ECUADOR
PERU
CALLAO
PACIFIC
OCEAN
NAZCA
RIDGE
VALPARAISO
CHILE RIDGE
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Fig. 14

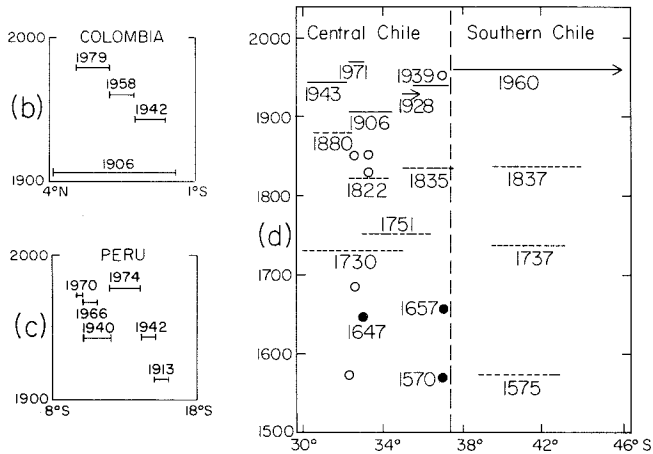

Fig. 14. (a) Map of large earthquake rupture zones along South America (from McCann *et al.*, 1979). (b) Historic record of large events along Colombia. (c) Historic record of large events along Peru. (d) Historic record of large events along Chile (from Kelleher, 1972).

Kelleher (1972) suggested that the 1906 rupture nucleated in the south, where the epicenter is located along the Carnegie ridge, and ruptured northeastward to the bend in the Colombian trench near 4°N. The extent of rupture was determined from reported water level changes and macroseismic intensity data, and is moderately well-constrained. Kanamori and McNally (1981) addressed the possibility that this was a normal-fault trench event by comparing Wiechert seismograms from Göttingen, Germany of the 1906 Colombian and the 1933 Sanriku events. The very sharp onset of the Sanriku event is different from the gradual buildup of the complex P wave train of the Colombian event, and the first motion polarity of the 1906 record is consistent with thrust faulting. In addition, the Love waves of the 1960 event show an amplitude asymmetry between G_2 and G_3 consistent with northeastward rupture propagation. Modeling of these surface waves yields a moment of 8×10^{28} dyne-cm, which is 1/3 of the moment Kanamori (1977) estimated based on the rupture zone dimensions. This discrepancy may be due to the poor long period response of the Wiechert instrument.

Kanamori and Given (1981) and Kanamori and McNally (1981) have conducted studies of the recent Colombian events. Kanamori and Given (1981) performed a moment tensor inversion of long period Rayleigh waves of the 1979 event. In addition to determining the moment and fault orientation, they used the source group delay apparent in the Rayleigh waves to infer that the rupture progressed unilaterally in a direction N40°E over a 230 km length. This dimension is in good agreement with the aftershock extent one day after the mainshock. The aftershock zone extended slightly in the following weeks.

Kanamori and McNally (1981) estimated the moment of the 1958 event to be 5.2×10^{27} dyne-cm ($M_w = 7.7$) by comparing the Rayleigh wave amplitudes of the 1958 and 1979 events. The aftershock area of this event determined by Kelleher (1972) can be used to infer a slightly smaller seismic moment of 2.8×10^{27} dyne-cm, using the empirical relations in Kanamori (1977). The similarity in size of the 1942 and 1958 aftershock zones is evidence that the moments of these two events were similar.

The total moment of the 1942, 1958, and 1979 events is 3.7×10^{28} dyne-cm, only 20% of the 1906 moment, despite the fact that these events ruptured the same area. This suggests that the coseismic displacement in a great earthquake is greater than that resulting from a series of smaller earthquakes rupturing the same zone. SYKES and QUITTMEYER (1981) suggest the generality of such behavior based on a rupture model in which the stress drop increases with rupture length.

2.9 Peru

Between the Colombian seismic zone and the Peruvian trench there is a large region for which no great earthquakes are recorded (Fig. 14). Here the Carnegie, Grijalva and Sarmiento ridges intersect the coasts of northern Peru and southern Ecuador (KELLEHER and MCCANN, 1976). These ridges bound the Peruvian trench in the north while the Nazca ridge intersects the trench near 17°S in the vicinity of the 1942 rupture. This ridge marks the northern extent of the Chilean volcanic line. The Nazca plate subducts under central Peru with a dip of about 10° to a depth of 150 km, and then may extend almost horizontally, yielding a very wide interface (e.g. KELLEHER *et al.*, 1974; BARAZANGI and ISACKS, 1976, 1979; ISACKS and BARAZANGI, 1977; STAUDER, 1975). It is also proposed that the angle of dip is closer to 30°, and reaches 100 km depth before extending horizontally over a 300 km distance (e.g. JAMES, 1978; HASEGAWA and SACKS, 1981; SNOKE *et al.*, 1979). This very broad contact may accumulate strain for centuries or may subduct aseismically, producing the gap in northern Peru. KULM *et al.* (1977) summarized the geologic nature of the trench and the basin morphology.

In a review of the historical seismicity of Peru, LOMNITZ and CABRÉ (1968) documented seven major events along Peru in 1586, 1655, 1687, 1746, 1828, 1940 ($M_s = 8.4$), 1966 ($M_s =$

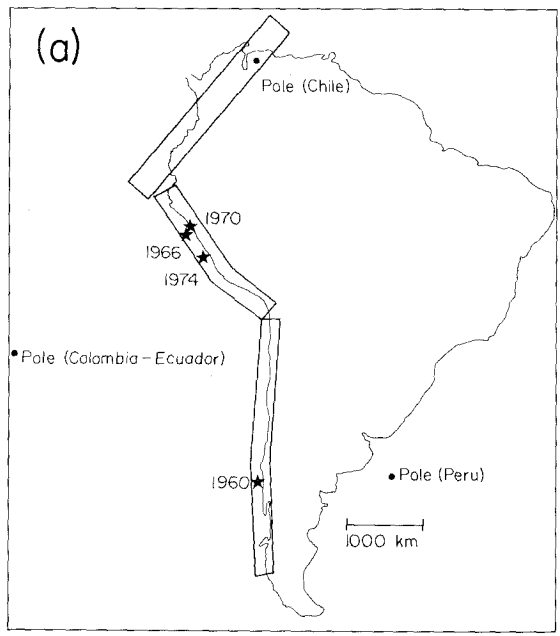

Fig. 15

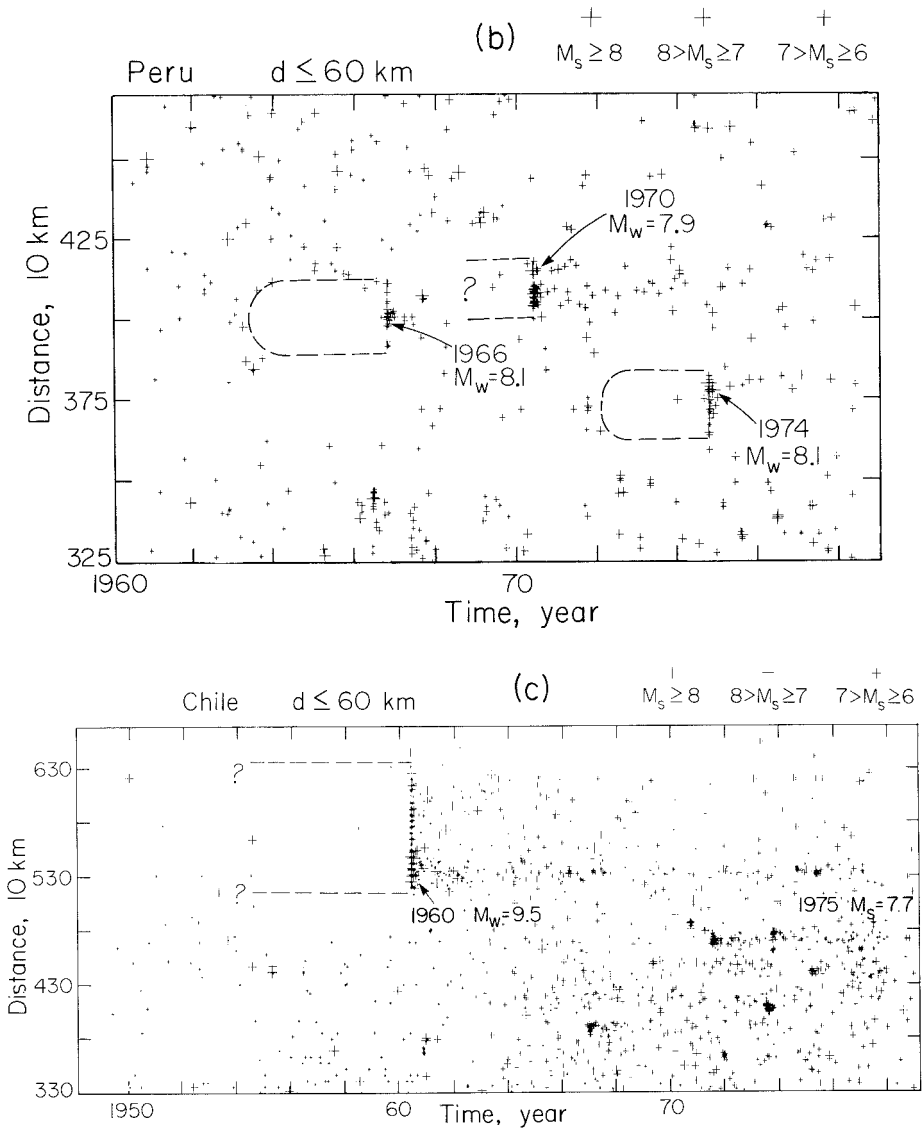

Fig. 15. (a) Index map for South America. See the caption for Fig. 7 for details. (b) Space-time plot of seismicity for Peru. For details, see the caption of Fig. 9. (c) Space-time plot of seismicity for Chile. For details, see the caption of Fig. 9. (from KANAMORI, 1981).

7.5), and the other recent activity in the trench shown in Fig. 14. These events clearly have an irregular failure sequence, with most rupture lengths being less than 150 km. The recent seismicity in the Peru trench is shown in Fig. 15, which shows an indication of quiescence preceding the 1966 and 1974 ($M_w = 8.1$) events. Neither of these earthquakes had foreshock activity detected by the world wide network, and the recorded activity is too sparse to reliably characterize the seismicity patterns of the region.

From an examination of strong motion and short period records, LOMNITZ and CABRÉ (1968) inferred complex faulting in the 1966 event, as also suggested by LAY and KANAMORI (1980), though detailed body wave analysis has not been done. ABE (1972b) studied the long period source of the 1966 thrust event and the 1970 normal event. As for the 1965 Aleutian events, the occurrence of the normal event following a thrust event indicates a complex subduction process. ISACKS and BARAZANGI (1977) and LOMNITZ (1971) showed that the 1970 event was also a multiple rupture with at least four subevents occurring during the first 20 sec. The source complexity of the 1966 events is also demonstrated by their aftershock distribution. DEWEY and SPENCE (1979) show that aftershocks of these as well as the 1974 earthquake have distinct clusters in separate zones occurring along the thrust interface as well as within the slab.

The 1974 event ($m_b=6.3$, $M_s=7.6$) which occurred on October 3, and its principal aftershock of November 9, were very complicated multiple ruptures. SPENCE *et al.* (1980) analyzed the short period body waves and accelerograms of these events, and detected 6–9 subevents in the main event and 3–4 in the aftershock. The October 3 sources appear to have originated in a volume less than 15 km in diameter, based on relative subevents timing. These events occurred just north of where the Nazca ridge underthrusts Peru. Detailed analysis of the body waves was not done, though it was noted that the first subevents in each earthquake are smaller than successive ruptures, producing a significant m_b-M_s discrepancy. The duration of the mainshock P wave signal is 81 sec on short period records and 99 sec on the Lima accelerogram. Long period surface waves of the 1974 event have been studied by Stewart (1980, personal communication), but yield little additional resolution on the source process. Though more detail of the stress distribution in the region must await further analysis, it appears that Peru has a very heterogeneous stress distribution.

2.10 Chile

South of the Peru trench the coastline of South America has an abrupt bend in the region south of where the Nazca ridge intersects the coast. The geometry of the convergent zone indicates that the underthrusting slab must be contorted in this region. HASEGAWA and SACKS (1981), and ISACKS and BARAZANGI (1977) suggested that the slab is actually torn, based on an abrupt change in depth to the Benioff zone in the region. Southward, along the Chile trench there are substantial variations in rupture lengths and other evidence for segmentation of the seismic zone (Fig. 14). STAUDER (1973) inferred that the process of subduction in the Chile trench is not a smooth continuous motion on the basis of the discontinuous nature of the regional seismicity and focal mechanisms. He suggested that the slab is segmented into tongues that subduct somewhat independently. CARR *et al.* (1974) and SWIFT and CARR (1974) studied seismicity in the Chilean Benioff zone at deep and intermediate depths. They distinguished seven segments along Chile trench where the seismic zone at depth has different strike and dip. The breaks in the deep seismic zone, which are most apparent below 300 km depth correlate with abrupt changes in topography and tectonic structures on the continent and in the trench, including offsets in the volcanic line. CARR *et al.* (1974) asserted that the margins of great earthquake rupture zones correspond well with the segment boundaries recognized in the morphology of the inclined seismic zone and surface features such as the coastline, trench and volcanic linea-

ments. BARAZANGI and ISACKS (1976) and ISACKS and BARAZANGI (1977) studied the seismicity distribution in the Chile trench and arrived at a more conservative segmentation of the trench into three regions, with northern central Chile having low-angle subduction. They could not find substantial evidence for pervasive offsets and changes in dip of the Benioff zone on the scale of 50 to 100 km suggested by CARR *et al.* (1974) and SWIFT and CARR (1974).

Most of the seafloor lying along the trench is smooth north of the Chile ridge, which lies at the southern end of the 1960 rupture zone (MOGI, 1969b). There is no obvious bathymetric feature at the northern end of the 1960 zone, although the coast curves to the northeast there (KELLEHER and MCCANN, 1976). The Juan Fernandez ridge strikes perpendicular to the trench in central Chile, near the 1906 earthquake. The submarine morphology, continuity of benches, and evidence for geologic segmentation along the trench are summarized by KULM *et al.* (1977).

The Chilean seismic zone has a very gentle dip north of the Juan Fernandez ridge similar to the Peru trench, whereas the northern and southern portions of the Chile trench have a steeper dip of 30° (ISACKS and BARAZANGI, 1977). A gap in intermediate depth activity also distinguishes central Chile from the north, but it is not clear whether the slab is torn or continuous at intermediate depths. KELLEHER *et al.* (1974) found that the plate interface indicated by shallow seismic activity broadens south of 38°S, and is 200 km wide throughout the region ruptured by the 1960 event. This interface is substantially wider than along the rest of the trench. The contact narrows in the zone between the 1960 earthquake and the 1922 events, from 30°–37°S. Smaller rupture lengths are found in this region as indicated in Fig. 14. The 1922 rupture occurred in an area of slightly broader interface contact. North of the 1922 zone, in the region of the 1868 and 1877 events, the contact is of intermediate width, slightly greater than in the 30°–37°S range. The tsunami source regions for these events given by SOLOVIEV and GO (1974) are comparable to the 1906 Colombian source dimensions, thus large events do occur in this region though with an apparently longer recurrence interval. The tsunami record indicates that a large event occurred in the 1868 zone in 1604 right at the bend in the coastline (MCCANN *et al.*, 1979).

Details of earthquakes occurring in the central and southern Chile trench are sparse though there is a well-defined recurrence period of about 100 years in the region, and a marked tendency for ruptures to occur successively southward (KELLEHER, 1972). The 1943 and 1971 events may indicate a new sequence of rupture propagation, though the 1971 event only ruptured about 1/3 of the failure zone of the 1906 event. MALGRANGE *et al.* (1981) have analyzed the body and surface waves of the 1971 earthquake, finding comparable seismic moments at short and long periods. The rupture area of this event is rather small, and they infer a rapid, simple stress drop at the source. There was a large aftershock series for this event. These authors also studied the 1965 normal-fault event which was located near the 1971 epicenter but at a greater depth of 72 km. They found four other incidents of pairing of thrust faulting and deeper normal faulting in the northern Chile region.

The great 1960 event has been the most completely studied event in the region. KELLEHER and SAVINO (1975) observed that the rupture zone was quiescent (Fig. 15(c)) for at least ten years except for a large burst of foreshock activity commencing 33 hours before the main event. This foreshock sequence included two $M=7.3$ events near the mainshock

epicenter in the northern portion of the rupture zone. The nucleation of this rupture along the boundary nearest the previous activity has been discussed by MOGI (1969b) and KELLEHER (1972), who describe numerous examples where earthquakes nucleate along the margins of preceding events presumably due to stress concentration at the end of the dislocation, though SYKES (1971) points out numerous cases where this did not occur.

NAGAMUNE (1971) analyzed crustal Rayleigh waves of the 1960 Chile event to examine the rupture complexisty. He identified three source regions and a southward propagation of the rupture. A later analysis of the regional structure indicated to NAGAMUNE (1978) that each source was in a distinct tectonic block, of which there are three along the rupture zone, and one more lying south of the Chile ridge. KANAMORI and CIPAR (1974) studied the Pasadena strain record of the 1960 event and obtained a seismic moment for the event of 2.7×10^{30} dyne-cm and an average dislocation of 24 m, in agreement with the results of PLAFKER and SAVAGE (1970).

KANAMORI (1977) determined that the convergence rate along Chile is approximately equal to the seismic slip rate, and associated the low-angle, broad contact zone, and compressive state of stress in the trench with strong coupling on the fault plane. The absence of deep earthquakes along this part of the zone was attributed to absence of deep continuous plate, perhaps caused by it breaking off due to the strong coupling at shallow depth. The inferred stress state is associated with large earthquake ruptures, as may be true for the 1964 Alaska earthquake also. The southern boundary of this zone of strong coupling is indicated by the location of the June 6, 1960 aftershock of the 1960 event, which occurred on the South Chile ridge. This event, a remarkable slow earthquake, was examined by KANAMORI and STEWART (1979). The rupture process consisted of a complex sequence of 15 distinct events. The small $M_s = 6.9$ is in obvious disharmony with the seismic moment of 5.6×10^{27} dyne-cm. Such a slow event is associated with weak coupling on the fault, perhaps the result of the subducting ridge modifying the strong coupling on the fault zone that exists further north.

2.11 New Hebrides

The New Hebrides subduction zone is similar to Middle America in having typical rupture lengths of only 75 to 150 km (Fig. 16) and a narrow zone of shallow activity of uniform width of 100 km along the trench (KELLEHER *et al.*, 1974). The high subduction rate in the trench of 11 cm/yr, and the tendency for large events to be temporally and spatially linked is similar to the Solomon Islands region. MCCANN (1981) found that 1/2 of the large shocks ($M_s \geqslant 7$) occur within a few weeks and within 100 km of another large shock in the trench. This clustering occurs at a slightly different magnitude range than in the Solomon Islands, for only two $M_s > 7.8$ events in the New Hebrides have occurred this century. The significant clustering of $M_s = 7$ activity includes events in 1946 ($M_s = 7.3$, 7.0 separated by 15 days), seven shocks in 1943–1945 near Aneityum island, triplets in 1965 and 1973 near the D'Entrecasteaux fracture zone, and two clusters of activity in 1980 near Santa Cruz Island ($M_w = 7.5$, 6.8, 7.8) and near the Loyalty Islands ($M_s = 6.7$, 6.7, 7.2, 6.5). MCCANN (1981) has reviewed the historic record of large earthquakes in this region, though the rupture characteristics of events prior to 1965 are poorly constrained.

The regional seismicity shows no resolvable contortions or changes in convergence direction along the entire extent of the seismic zone (PASCAL *et al.*, 1978; ISACKS *et al.*,

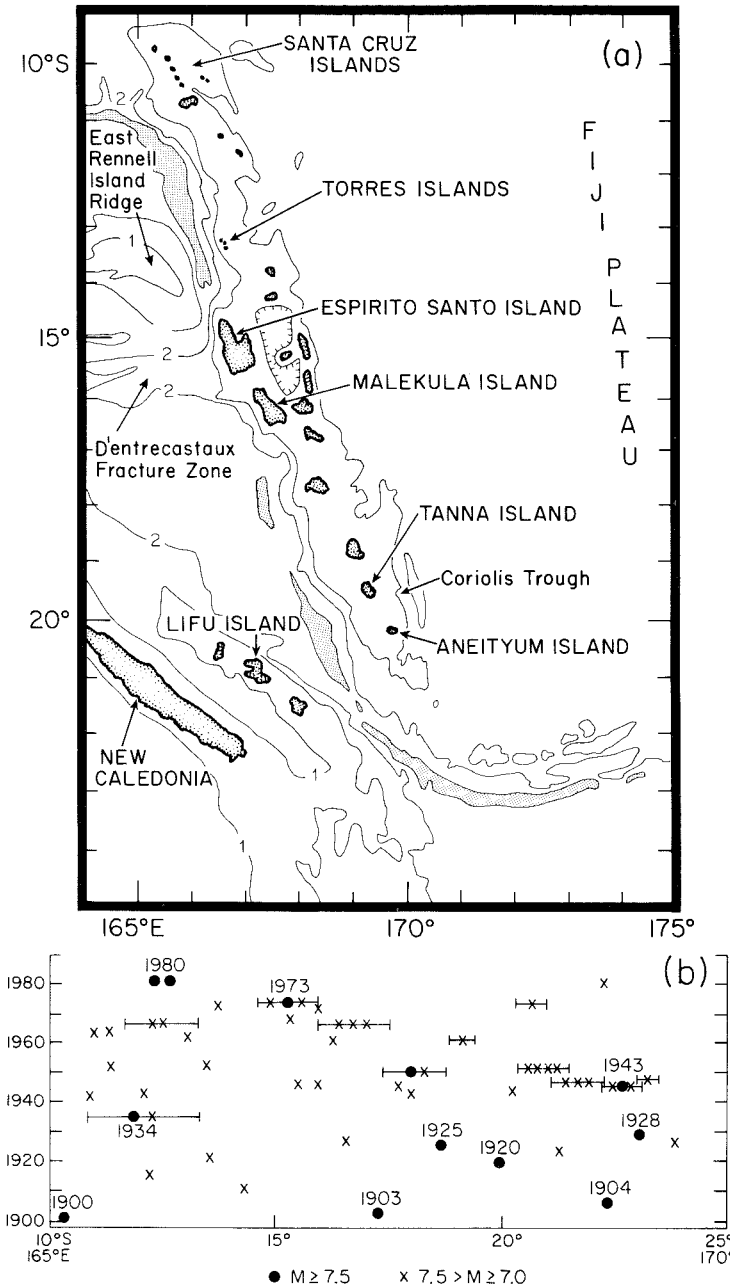

Fig. 16. (a) Map of the New Hebrides Island arc. (b) Historical record of M_s 7.0 events along the New Hebrides Island arc, (from McCann, 1981).

1981). There is abundant evidence of segmentation of the converging plates, particularly in the extensively studied region of the D'Entrecasteaux fracture zone (PASCAL *et al.*, 1978; TAYLOR *et al.*, 1980; ISACKS *et al.*, 1981). These segments tend to be about 100 km in length, and the bathymetric and structural boundaries by which they are defined appear to delimit the rupture zones of large events. A clear example of this is the separation of the 1965 and 1973 sequences along the eastward extension of the northern scarp of the D'Entrecasteaux fracture zone (PASCAL *et al.*, 1978; EBEL, 1980; ISACKS *et al.*, 1981).

Detailed seismicity studies along the New Hebrides trench reveal regional variations in seismicity pattern (PASCAL *et al.*, 1978; COUDERT *et al.*, 1981; ISACKS *et al.*, 1981). The seismic activity is generally high and continuous along the zone despite the disappearance of the trench from 14 to 18°S, where the D'Entrecasteaux fracture zone intersects it. The region of the southern Malekula and Efate islands has not ruptured in a large event for 75 years and shows a diffuse distribution of small events (ISACKS *et al.*, 1981). This region does have numerous events with $M_s \leqslant 6.5$, of which the 1978–1979 sequence is notable for having clear foreshock migration toward the mainshock epicenters and anomalously large aftershock area expansion (ISACKS *et al.*, 1981). There may be a large component of aseismic slip in this region.

Several studies have explored the effects of subduction of the D'Entrecasteaux fracture zone (PASCAL *et al.*, 1978, TAYLOR *et al.*, 1980; ISACKS *et al.*, 1981; CHUNG and KANAMORI, 1978a, b). In spite of the disappearance of the trench, the overall configuration of the subducted plate is not significantly disturbed in this region. There may be a slightly decreased dip to the seismic activity above 50 km depth, and there is a gap in activity from 50 to 120 km depths. This intersection is associated with clustering of large activity. CHUNG and KANAMORI (1978a, b) studied a very complex rupture that occurred in 1969 at a depth of 107 km, which excited anomalously large long period waves similar to those of the June 6, 1960 Chile event. This event indicates that subduction of the ridge is occurring, rather than being locked up, and there may be a substantial component of creep at depth associated with this. CHUNG and KANAMORI (1978b) review the bathymetry and seismicity of the fracture zone-trench intersection and conclude that there is strong interaction at shallow depths between the subducting and overriding plate, yielding subsidence landward of the plate contact and uplift seaward. Such a strong interaction, resulting from the buoyancy of the fracture zone, is reflected in the large events near the junction.

The recent shallow trench activity near the D'Entrecasteaux fracture zone has included the 1965 and 1973–1974 sequences. The 1965 sequence was preceded by 4.5 years of quiescence (PASCAL *et al.*, 1978), and ruptured progressively southward. The 1973 events propagated southward from the northern end of Santo Island, and were followed within a month by the 1974 sequence which extended the 1973 failure zone northward (ISACKS *et al.*, 1981).

EBEL (1980) has studied the 1965 sequence of events, including body and surface wave analysis of two foreshocks, the main shock, and two aftershocks, all of which were thrust events except the second aftershock. The events occurred near the islands Santo and Malekula, bounded on the north end by a transform fault and on the south by east-west tectonic trends. The body waves indicate a complicated stress state in this region. The first foreshock and first aftershock studied are simple events, but the mainshock was a double event with the second source being larger than the first. The surface wave moments proved

consistently greater than the body wave moments by a factor of three. The foreshocks also appear to have had longer rise times than the following events suggesting failure of weaker asperities concentrating stress on large asperities which failed in the main event. Ebel suggested that the sequence was similar to the 1968 Tokachi-Oki event with small events migrating in one direction, followed by a smoother dislocation rupturing in the opposite direction.

VIDALE and KANAMORI (1981) have studied the 1980 sequence near the Loyalty Islands, using body and surface wave data. The four moderate size events in this sequence were pure thrust events on a plane dipping 20° and striking parallel to the trench. These events had clustering of activity near the mainshock epicenters and low activity in the rest of the zone prior to rupture. The aftershock zones show significant expansion, with the area covered after one day increasing by a factor of from 5 to 10 in the next two days.

2.12 Ryukyu-Taiwan-Philippines

The convergent zone between the Philippine Sea and the Asian mainland has a wide range of large earthquake behavior. A long term zone of infrequent large earthquakes extends from Kyushu to the Miyako depression. Seaward of the Ryukyu trench lies a series of aseismic ridges, including the Kyushu-Palau ridge (KELLEHER and MCCANN, 1976). As the northern extreme of this zone, off eastern Kyushu and southwest of the Nankai trough, no great earthquakes are known to have occurred. However, events with $M_s \leqslant 7.5$, and short rupture zones of a few tens of kilometers in length, are fairly common (UTSU, 1974; SHIONO *et al.*, 1980). These events are predominantly thrust events, whereas the reported mechanisms along the central Ryukyu trench are shallow normal or reverse faulting events (SHIONO *et al.*, 1980; SENO and EGUCHI, 1981). Back arc spreading may be associated with normal faulting events in the Okinawa trough (EGUCHI, 1981). Near the southwestern segment of the Ryukyu trench, approximately between the Miyako depression and Taiwan, a large event occurred in 1771, and several smaller events have been reported (ROWLETT and KELLEHER, 1976). The 1771 event appears to have been the largest event along the trench since at least 1600.

Though there is significant uncertainty in estimating the convergence rate, an average along the trench of 5 cm/yr, with a northeast to southwest increase from 4.4–6.5 cm/yr is reported by SENO and EGUCHI (1981). CARR *et al.* (1973) correlated surface geology and intermediate and deep seismic activity with segmentation of the slab subducting in the Ryukyu arc. SHIONO *et al.* (1980) found that the northern and southern sections of the arc are distinctive. The segment north of the Tokara channel has high volcanic activity, intermediate depth focal mechanisms with downdip tension, and a steep dip (70°) of the Benioff zone below 100 km. South of the Tokara channel, the volcanism is low, intermediate depth activity has down dip compression, and the Benioff zone has a dip angle of 40–50°.

Taiwan has had numerous events with $M_s = 7$–7.5, and an event in 1920 is reported to have $M_s = 8.0$ (MOGI, 1969a; SENO and EGUCHI, 1981). The rupture zones of these events extend only a few tens of kilometers. The focal mechanisms in this region have scattered orientations, suggesting a complex stress distribution (KATSUMATA and SYKES, 1969; WU, 1970, 1979; SENO and KURITA, 1979). The tectonic environment is complicated by the collision of the remnant Luzon arc and the Asian continental mainland, which may have

produced an arc reversal (ROWLETT and KELLEHER, 1976). There are few details known about the rupture process of events in Taiwan, but SENO and EGUCHI (1981) review the available focal mechanism determinations. To the south of Taiwan, the Luzon strait region has not produced a large earthquake in the historic record. This region is a wide zone of *en echelon* faulting and complex tectonics (ROWLETT and KELLEHER, 1976; SENO and EGUCHI, 1981).

Along the Philippine islands, two major tectonic features are the Philippine trench and the Philippine fault, though the region is very complex tectonically. The general tectonics and seismicity in the Philippines have recently been reviewed by ACHARYA and AGGARWAL (1980) and SENO and EGUCHI (1981). Large subduction events in the Philippine trench have occurred in 1929, 1969, 1970, and 1972. These events have ruptured between 3 and 13°N, and typically have 100 km fault lengths (MCCANN *et al.*, 1979). A large event in 1952 ($M_s = 7.7$) may have involved normal faulting seaward of the trench. Our knowledge of the seismic record in this region is rather incomplete, and it is difficult to characterize the subduction process in the Philippine trench. SENO and EGUCHI (1981) have estimated that 10–19% of the convergence in this zone is accommodated by seismic slip, based on a 8 cm/yr convergence rate. There is substantial uncertainty associated with this estimate.

STEWART and COHN (1979) investigated a large event in 1976 in the Moro Gulf. This event is associated with a northeast dipping subduction zone in which the North Celebes Sea underthrusts Mindanao. There is some indication that this zone has had previous large events in 1897 and 1918.

2.13 Tonga-Kermadec-New Zealand

The rupture mode of Tonga-Kermadec region is one of the poorest known in the Pacific. Figure 17 shows the historic occurrence of large earthquakes along the zone. From this record it is difficult to detect systematic regional behavior. Several studies have cited the disruption of the subduction process by the Louisville ridge in the central portion of the trench (e.g., KELLEHER and MCCANN, 1976). Large earthquakes near this intersection are rare, and most events tend to occur in the central Kermadec region or along the northern Tonga Islands. The seafloor subducting beneath Tonga is quite smooth, and there is a continuous volcanic line along the trench, but the historic record does not clearly indicate whether all of this region tends to fail in large earthquakes (MCCANN, 1981). Much of the variation in seismic behavior along the trench is associated with the presence of aseismic slip. The subduction rate varies from 5–6 cm/yr along Kermadec to 10 cm/yr along Tonga (MINSTER *et al.*, 1974). At the northern edge of Tonga hinge faulting may be occurring, producing the concentration of activity observed there (ISACKS *et al.*, 1968; MCCANN, 1981).

There appears to be a zone along the Kermadec zone, from 31–34°S which has large earthquakes every 30 years, and may be converging without aseismic slip. South of this active area, the presence of back-arc spreading and subduction of old seafloor may indicate weak coupling on the fault plane and aseismic slip along the Harve trough (UYEDA and KANAMORI, 1979). The fact that aftershock zones in Kermadec often overlap indicates the distinctive behavior of this region, and this observation may be related to the presence of aseismic slip (MCCANN, 1981).

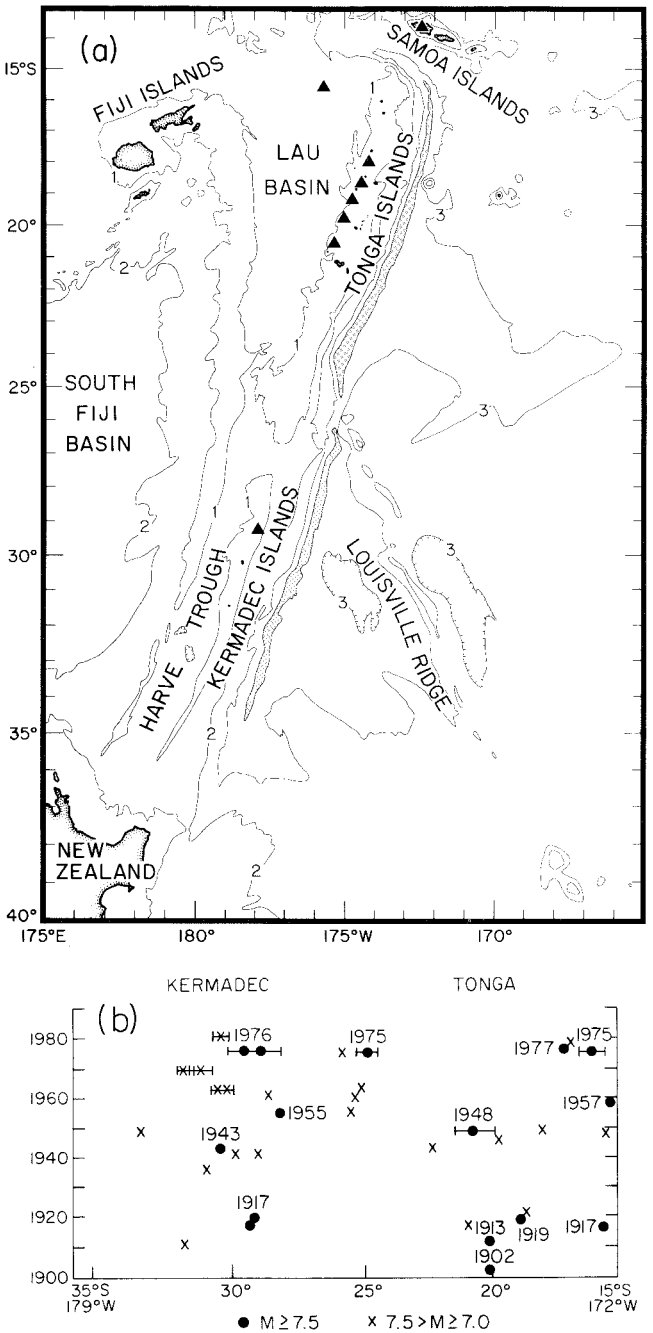

Fig. 17. (a) Map of the Tonga-Kermadec Island arc. (b) Historical record of $M_s > 7.0$ events along the Tonga-Kermadec Island arc, (from McCann, 1981).

The convergent zone between the Indian and Pacific plates along New Zealand is rather complex, and little detail is known about the large earthquakes in this region. There appear to have been large events in the subduction zone along North Island in 1855 ($M=8$) (EIBY, 1968), 1904, and 1931 ($M_s=8$). The rupture zones inferred from isoseismals do not overlap and have lengths of 100–200 km (NISHENKO and MCCANN, 1981). The trench along North Island is bounded to the south by the Chatham rise, and to the north by the aseimic Havre trough. The repeat time of large events along this zone is greater than 100–150 years, since no great shocks prior to 1855 are known to have occurred in the trench (EIBY, 1968). The recurrence interval for the 1855 rupture is estimated at 500–900 years based on offset river terraces (LENSEN and VELLA, 1971), and geodetic evidence indicates a shorter recurrence of 100–200 years (WALCOTT, 1978). The relative plate motion along this zone is somewhat uncertain due to the proximity of the Indian-Pacific pole of rotation, but is estimated at 4–5 cm/yr (MINSTER and JORDAN, 1978). Subduction also occurs at the southern end of South Island, where the Indian plate underthrusts the Fiordland region. This was the site of a large event in 1826, but previous activity is uncertain (NISHENKO and MCCANN, 1981).

2.14 Izu-Bonin, Marianas, Java, Sumatra

The Izu-Bonin and Marianas trenches are even more distinctive in the complete absence of great shallow earthquakes in this century. Many topographic features on the subducting plate are found, including the Marcus-Necker ridge, Magellare Seamounts, Caroline ridge, and many seamounts (KELLEHER and MCCANN, 1976). The absence of large earthquakes has been attributed to either a freeze-up of the subduction process due to density contrast as suggested by KELLEHER and MCCANN (1976), or due to weakening and decoupling of the interface contact and ongoing aseismic subduction as proposed by KANAMORI (1977). The southward increase in aseismic slip along the Kurile-Japan subduction region is consistent with the latter idea, as is the presence of back-arc spreading, assuming the model of evolutionary subduction proposed by KANAMORI (1971a, 1977) and UYEDA and KANAMORI (1979). The degree of aseismic convergence in the Marianas is important for determining how much of the total slip between the Pacific and Asia, which converge at 9 cm/yr, is accommodated in the Nankai trough (KANAMORI, 1977). SENO and EGUCHI (1981) suggest that 5.0 cm/yr of convergence is accommodated by the Izu-Bonin, Marianas zone. This estimate would indicate that the region has predominantly ongoing aseismic subduction.

The subduction zones along Indonesia are not well understood either. Along Java no great thrust earthquakes are recorded from 106° to 122°E (MCCANN *et al.*, 1979). The convergent zone is disrupted by the Australian continent to the east and by subduction of the Christmas Island ridge along the Java trench. The lack of large earthquakes, and the occurrence of large normal faulting events such as the 1977 event ($M_w=8.3$) which appears to have ruptured through the lithosphere (STEWART, 1978; GIVEN and KANAMORI, 1980; FITCH *et al.*, 1981) may suggest aseismic convergence, but the completeness of the historic record is uncertain. Further to the west, along Sumatra, the character of subduction appears to change, perhaps reflecting the change in orientation of the zone relative to the plate convergence direction or the difference in age of the subducting lithosphere along the zone. Large tsunamigenic events appear to have ruptured 600–700 km of this zone in

1883 and 1861 (McCann *et al.*, 1979). The average convergence rate along the trench is 7 cm/yr (Minster *et al.*, 1974). A better understanding of the indicated change in coupling along the zone must await more detailed studies of the region. The available focal mechanisms and seismicity of the region have recently been reviewed by Seno and Eguchi (1981).

3. Subduction Zone Categories

The general characteristics of each subduction zone discussed in the preceding section are summarized in Table 2. The comments and numerical values in the table are representative of the gross features of each particular zone; substantial deviations from the given values are possible within that zone. The order in which the zones in Table 2 are listed is loosely based on maximum rupture length observed for each region. This characteristic is perhaps the most reliable with which to begin to categorize the various regions. Not only does rupture length reflect the moment of the large earthquakes, but it is indicative of the degree of coupling on the fault plane. Rupture length is also one of the only parameters which can be fairly reliably determined for older events.

Several correlations in regional characteristics are apparent in Table 2. Clearly, the occurrence of large rupture zones is associated with an absence of large transverse structures segmenting the zone and smooth seafloor topography, as indicated by Kelleher and McCann (1976, 1977). There is also a tendency for regions with larger rupture zones to have a greater percentage of seismic slip. Contact widths are usually greater for regions with long rupture zones as suggested by Kelleher *et al.* (1974). Characteristic source time function durations appear to be longer for zones with large ruptures as well. Back arc spreading is associated only with regions having maximum rupture zone dimensions $\leqslant 150$ km long.

Given the great variation in mode of subduction around the Pacific, it is useful to establish general categories of behavior. We rely principally on maximum rupture length (or maximum moment) to define these categories. On this basis, it is possible to divide the subduction zones into four basic groups, as shown in Table 3. Category 1 is exemplified by Southern Chile (Fig. 14), in which great earthquakes with rupture lengths longer than 500 km tend to occur regularly in time over approximately the same large rupture zone. The 1964 Alaska and 1952 Kamchatka earthquakes occurred in regions which appear to have this mode of failure, though the historic record is limited. The behavior of the Central Aleutians, which failed in the 1957 event, is unclear (Sykes *et al.*, 1981), but this region may behave similarly. It is apparent in Table 2 that these regions have other characteristics in common than just large rupture lengths. The repeat times for these great earthquakes is around 100 years for Kamchatka and Chile, and there is a corresponding large percentage of seismic slip. The contact widths in these zones are large, from 120–300 km, and the angle of dip is known to be small in Southern Chile and Alaska. Each region has demonstrated extensive foreshock or preseismic activity in the forthcoming rupture zone prior to large events. While only the 1964 Alaska earthquake has been analyzed for its body wave time function characteristics, which is a very long duration (>120 sec) rupture, it is reasonable to expect that it is representative of these great events. The margins of the large rupture zones tend to abut large transverse features, such as the Chile Rise, the

Table 2. Characteristics of Circum-Pacific subduction zones inferred from large earthquakes.

Zone	Rupture lengths (km)	Repeat times (yrs)	Seismic slip %	Sea floor topography	Transverse structures	Anomalous foreshock-aftershock behavior	Large normal fault trench and tsunami events	Contact width (km) dip (°)	Body wave complexity and time function character	Back-arc Spreading
Southern Chile	~1000	~100	75–100%	Smooth	Bounded by Chile Rise, Some canyons	Extensive foreshock activity just before the main shock	Slow events on Chile Rise	200 km 10–20°	No record	No
Alaska	100–200 along peninsula, 800 in east	Uncertain 50–75 in west, >100 in east	Uncertain 30–100%	Gulf of Alaska Seamounts	Geologic provinces in overthrust slab	Increased preseismic activity near focus	No record	150–360 km ~9°	Complex ruptures Duration >120 sec	No
Kamchatka	500 in south 150 in north	~108[2]	~60%	Smooth	Little evidence	Increased preseismic activity near focus	No record	200 km Uncertain ~30°	No record	No
Aleutians	Temporal variation 100–1000	Uncertain ~60 in west	Uncertain May be 100%	Smooth	Changes in strike, canyons ridges	Increased preseismic activity near focus	Large tsunami & normal fault events	120 km 15–20°	Complex multiple ruptures Durations ~60 sec	No
Colombia	Temporal variation 150–600	Variable 36–>73	Uncertain 30–55%	Smooth	Little evidence	Little foreshock activity for 1979 event	No record	150 km Uncertain	Long Duration rupture ~60 sec	No
Nankai Trough	Temporal variation 150–300	170±70	Uncertain 70%[3]	Smooth	Canyons, terraces 75–100 km	No record	No record	100 km 10°–30°	Uncertain	No
Solomon Islands	Doublets 100–300	25–40	50%	Smooth	Woodlark Ridge and New Britain Trench	Few foreshocks, aftershocks	No record	<100 km 30–40°	Simple events, Doublets Durations 10–15 sec	No

New Hebrides	100–200	25–40	Uncertain 50%	Islands Ridges	D'Entrecasteaux and East Rennel Ridges	Foreshocks for some events	Slow events on D'Entrecasteaux Ridge	<100 km 35–40°	Sequences of simple events 10–15 sec Durations	Yes
Middle America	100–200	~35	Variable 10–100%	Fracture zones & Ridges	Large ridges & changes in strike	Moderate foreshock activity	No record	<100 km	Sequences of simple events 10–15 sec Durations	No
Kurile Islands	200–300	Variable 79–>140	25%	Smooth	Some canyons, 200 km	Many foreshocks	Frequent tsunami events	150 km 20°	Complex multiple ruptures 15–30 sec Durations	No
Japan Trench	150 decreasing southward	100 (north) 800 (south)	40% (north) 5% (south)	Seamounts in south	Canyons, terraces 100–200 km	Few foreshocks	Large tsunami & normal fault events	~150 km 10–25°	Sequences & multiple source events	No
Central Chile	400 (north) 200 (south)	100	Uncertain >50%	Juan Fernandez Ridge	Ridges, changes in strike & dip of Benioff zone	No record	Moderate size normal fault events at ~>70 km depth	100–150 km 10–15°	No record	No
Peru	150	Variable 100	Uncertain	Nazca Ridge	Nazca Ridge	Split aftershock zones	Intermediate depth normal fault events	150 km 10–30°	Complex multiple events	No
Philippine	100–150	Uncertain	<20%[4]	Smooth	Transition to Philippine fault	No record	Possible normal fault event in 1952	Uncertain ~30°	No record	No
New Zealand	100–250	150–500	Uncertain	Smooth	Chatham Rise	No record	No record	~150 km Uncertain	No record	Yes
Izu-Bonin Marianas	No events $M_s > 7.4$	—	Uncertain probably very small	Many ridges, seamounts	Many ridges, seamounts	No record	No record	<100 km Uncertain ~20°	No record	Yes

Table 2. (continued)

Zone	Rupture lengths (km)	Repeat times (yrs)	Seismic slip %	Sea floor topography	Transverse structures	Anomalous foreshock-aftershock behavior	Large normal fault trench and tsunami events	Contact width (km) dip (°)	Body wave complexity and time function character	Back-arc Spread-ing
Tonga Kermadec	<150	Highly variable	Uncertain 0–90%	Smooth along Tonga, Louisville Ridge	Louisville Ridge	Overlap-ping after-shock zones	Large nor-mal fault events	100–150km	Uncertain	Yes
Ryukyu	<100, may be larger in southwest	Uncertain	Uncertain probably very small	Many seamounts	Kyushu-Palau Ridge	No record	No record	~100 km Uncertain	No record	Yes
Java	No record of events $M_s > 7.0$	Uncertain	Uncertain probably very small	Many Seamounts	Christmas Island Ridge	No record	Large nor-mal fault events	Uncertain ~20°	No record	Yes
Sumatra	~300 (?)	Uncertain	Uncertain	Some seamounts	Investigator Fracture Zone	No record	No record	Uncertain	No record	No

[1]As observed on WWSSN Long Period Seismograms.
[2]Based on tsunami records.
[3]Based on convergence rate of SENO (1977).
[4]Estimated by SENO and EGUCHI (1981).

Table 3. Subduction zone categories.

Category	Zones	Large earthquake characteristics
1	Southern Chile Kamchatka Alaska Central Aleutians (?)	Regular occurrence of great events with ruptures >500 km long. Large percentage of seismic slip. Increased seismicity prior to main events. Long duration (>120 sec) body wave time function.
2	Western Aleutians Colombia Nankai trough Solomon Islands Sumatra (?)	Variations in rupture extent with occasional ruptures 500 km long. Clustering of large earthquake activity and doublets. Frequent precursory quiescence prior to large events. Body wave source time functions with discrete long duration (30–60 sec) ruptures (shorter for the Solomon Islands doublets).
2–3	New Hebrides Middle America Central Kermadec (?)	Events with smaller rupture dimensions (100–150 km) and shorter source durations (10–15 sec) than category 2, but a clear tendency to occur as discrete ruptures that cluster in time. The recurrence interval is short (25–40 years).
3	Kurile Islands Japan trench Peru Central Chile New Zealand (?) Philippine trench (?)	Repeated rupture over subzones of 100–300 km length. Recurrence intervals of 100 years or greater. Multiple rupture events and complex failure zones. Occurrence of normal faulting in trench or within subducting slab.
4	Marianas Izu-Bonin Tonga Southern Kermadec Java Ryukyu	Large thrust earthquakes are infrequent or absent. Back-arc spreading is known or suspected to occur. Large amounts of aseismic slip are inferred.

narrowing of the plate interface south of Kamchatka, and the break in trend of the Aleutian trench near Amchitka Island.

Category 2 contains zones with maximum rupture lengths of up to 500 km, as typified by the Western Aleutians, which ruptured in the 1965 Rat Island event (Fig. 10). This region, as well as the subduction zones in Colombia, northern Kamchatka, and the Nankai trough have demonstrated temporal variation in rupture mode, with occasional very large ruptures spanning several segments of the trench which fail individually in other sequences. The eastern Aleutians, along the Alaskan peninsula, may also be associated with this behavior (Fig. 10). The Solomon Islands zone is placed in this category because of the frequent occurrence of large doublet events that are spatially and temporally linked. The rupture lengths of large Solomon Islands events are smaller than in the other zones of this category, but the occurrence of discrete, closely related ruptures indicates that the mode of failure is similar to the other regions. In Table 2, it is apparent that these regions have little seafloor topography on the subducted plate, and plate contact widths of 100–150 km. There is little foreshock activity for the Solomon Islands and Colombian zones, but substantial preseismic activity occurred in the Rat Island rupture zone. The contribution of seismic slip for zones in this category is intermediate, with estimates

ranging from 30–100% of the total slip. The repeat times of these zones are also intermediate, from 25–70 years, except for the Nankai trough which has a large repeat time of 170 years. The variability in rupture mode in these regions makes estimation of the seismic slip and repeat time somewhat uncertain. The characteristic body wave time function durations are about 60 sec for the 1979 Colombia and 1965 Aleutian events, which are probably comparable to the Nankai trough events, given the similarity in moment and rupture extent of these regions. The Solomon Islands time functions are shorter in duration, but since the events occur as doublets and triplets, the effective duration may be substantially larger.

Category 3 is characterized by the Kurile Islands mode of failure (Fig. 8), in which large earthquakes repeatedly rupture the same portion of the subduction zone without coalescing to produce larger ruptures. Peru, Central Chile, and the northern segment of the Japan trench demonstrate this type of rupture behavior, although the rupture zones range from 100–300 km in length. Other features which are found in common for these zones are the repeat interval of about 100 years in individual segments, the low percentage of seismic slip (25–50%), and the occurrence of complex body waves resulting from multiple sources in the individual events. The rupture durations of the subevents are 15–30 sec in the case of the 1963 Kurile event. The interface contact width is typically 100–150 km for these zones, and the dip angle ranges from 10–30°.

Several regions, including Middle America, Central Kermadec, New Zealand, the New Hebrides, Taiwan, and the Philippine trench produce somewhat smaller rupture lengths of 100–150 km. The Middle America and New Hebrides events have been more thoroughly studied, and are known to have short recurrence intervals of 25–40 years, and are usually simple events that tend to occur in sequences. Typical body wave rupture durations are 10–15 sec for these events. These are similar to the Solomon Islands events, though on a smaller scale, and the tendency for the events to cluster in time and space suggests that these zones are intermediate between categories 2 and 3. This departure from a categorization based purely on rupture length is motivated by a desire to group zones with common rupture behavior together.

Category 4 is characterized by the Marianas-type absence of great earthquakes. The Izu-Bonin, Java, Ryukyu, southern Japan trench, Tonga and southern Kermadec zones are placed in this category. These zones are inferred to have a small percentage of seismic slip, have substantial seafloor topography, and are distinctive in the presence of active back-arc spreading.

In defining the subduction zone categories above, there is a fair degree of ambiguity associated with zones for which the rupture history is not well characterized. For instance, it is not clear whether the Central Aleutians should be placed in category 2, based on the indications of variable rupture mode, rather than grouped with the very long rupture zone events of category 1. Also, for the zone along Sumatra, it is difficult to say whether category 1 or 2 is more appropriate, and further study must be conducted in this area.

An interesting feature which is associated with the categorization based on rupture length given above, is the systematic variation in characteristic body wave source time function duration. In Fig. 18, several time functions determined for large events in categories 1, 2 and 3 are shown. It is apparent that the rupture process is distinctly different between each category, which indicates that there are fundamental differences in the state of stress associated with each. The asperity model which will be discussed in the next section

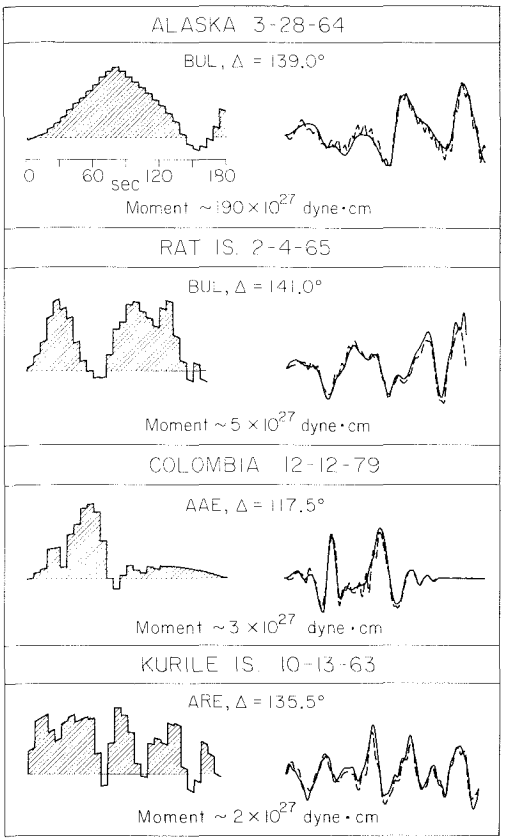

Fig. 18. Example of the variation in source time function determined from the body waves of various great earthquakes. The data (solid lines) are compared with the synthetics (dotted lines) generated for the source time functions given on the left (from RUFF and KANAMORI, 1980b).

provides an explanation for the various phenomena associated with each subduction zone category.

4. Asperity Model

Before attempting to interpret the observed variations in subduction zone characteristics, it is necessary to clarify what the asperity model is. It should be emphasized at the outset that the model is evolving as new information becomes available. The asperity model originally developed through interpretation of laboratory experiments on rock friction, and was first proposed by BYERLEE (1970) and SCHOLZ and ENGELDER (1976). These works suggested that the two sides of a fault are held together by localized areas of high strength called asperities. The stress at the asperities is large relative to the average stress on the fault plane. The nature of the stress concentrations may be variations in the geometric orientation or heterogeneities of the frictional strength or mechanical properties along the contact zone. On any given fault, there should be a distribution of various scale lengths of asperities, with local conditions favoring particular scale length distributions.

Thus, in its most general form, the asperity model is essentially a model of heterogeneous stress along a fault zone. The strength of coupling on the fault is governed by the product of the area of contact and the average breaking stress of the asperities, with strong coupling resulting from large asperity area and high friction coefficient.

The dynamic behavior of an asperity distribution was considered by KANAMORI (1978a) in an attempt to explain the complexity of large events. Localized slip occurs when the shear stress on the fault surface exceeds the local yield stress, and this slip is accompanied by an increase of stress on stronger asperities. Thus, the weak zone between strong asperities may behave aseismically, or with background earthquake activity. A small earthquake represents failure of a single asperity, for which the rupture zone is delimited by adjacent asperities. If the initial failure induces large enough stress increments on adjacent asperities, they too may fail, either in a complicated multiple source event such as the 1976 Guatemala earthquake (KANAMORI and STEWART, 1978; STEWART and KANAMORI, 1978), or in closely related events such as doublets. If the asperities are uniform in size and strength, failure of one asperity is accompanied by less slip than occurs if several asperities fail. Clearly, the details of the distribution of asperity scale lengths and strength affect the asperity interaction, and hence these govern the rupture mode of a given region.

Various aspects of seismicity patterns and foreshock activity (WESSON and ELLSWORTH, 1973; JONES and MOLNAR, 1979; ISHIDA and KANAMORI, 1978, 1980; BAKUN *et al.*, 1980) and preseismic quiescence (MOGI, 1977; KATSUMATA and YOSHIDA, 1980; KANAMORI, 1981) have been interpreted in the context of variation of asperity size within a given fault zone. In considering the large earthquake behavior of various subduction zones, we are primarily trying to characterize the relative distribution of large asperities and the interaction between them (LAY and KANAMORI, 1980, 1981; EBEL, 1980; RUFF and KANAMORI, 1980b; KANAMORI and MCNALLY, 1981). The asperity model is attractive for this purpose, since it is a simple model with substantial flexibility. The details of the failure process and triggering interaction are not well known, but qualitative behavior can be discussed using the asperity concept. Also, it is not necessary for our purpose to know the exact nature of an asperity, nor the direct causes of their variation between zones, though such a complete understanding is the ultimate objective of this line of research.

Many aspects of the source complexity of large events and associated seismicity patterns can also be interpreted in the context of the barrier model proposed by BURRIDGE and HALLIDAY (1971), DAS and AKI (1977), and AKI (1979). There is often some confusion about the differences between these two models. The basic tenet of each model is that the stress on a given fault is not uniform, and that localized zones of stress concentration govern the mode of failure. The difference between the two models is in the associated dynamic failure properties. In the asperity model rupture begins at a strongly coupled part of the fault and propagates into weaker adjacent areas, whereas in the barrier model the rupture initiates in the weaker zone and propagates into an area of high breaking stress which may or may not remain unbroken. Clearly, if a barrier ruptures, it is behaving very much like an asperity, and similarly, if adjacent asperities stop the rupture caused by failure of an asperity they are behaving as barriers. The quantitative aspects distinguishing the asperity and barrier models have been discussed by HUSSEINI *et al.* (1975), MADARIAGA (1979), RUDNICKI and KANAMORI (1981). The nature of the interaction of the rupture front

with a barrier, or of the failure of an asperity do predict somewhat different time histories of the rupture process. In particular, the asperity model indicates that rupture initiates with a high stress drop event, whereas in the barrier model rupture may terminate with a high stress drop event, but these differences are not adequately resolved in the seismic record to discriminate which model is more appropriate for the rupture process of large earthquakes.

The seismological evidence for localized stress concentrations of the type implicit in the asperity model is abundant and continues to accumulate. Preseismic clustering of events near the main shock epicenter appears to reflect the concentration of stress around a strong asperity (e.g. BRADY, 1976; MOGI, 1977; KANAMORI, 1978a; ISHIDA and KANAMORI, 1978; KATSUMATA and YOSHIDA, 1980). Aftershock zone expansion indicates that rupture in the main event occurs over a limited zone and expands into weakly coupled surrounding regions (MOGI, 1968b; ISACKS *et al.*, 1981; TAJIMA and KANAMORI, 1981; VIDALE and KANAMORI, 1981). Numerous studies of body waves of large events indicate smaller body wave source dimensions than the whole rupture zone dimensions indicated by surface waves and aftershocks, as indicated in Table 1. Additional lines of evidence include observations of reproducing earthquakes which have very similar waveforms, apparently resulting from failure at the same location under the same mechanical condition (HAMAGUCHI and HASEGAWA, 1975; ISHIDA and KANAMORI, 1980; GELLER and MUELLER, 1980), and attempts to detect an increase in stress drop of foreshocks and preshocks over that of background activity (TSUJIURA, 1977; ISHIDA and KANAMORI, 1980; BAKUN and MCEVILLY, 1979). UTSU (1980) investigated the spatial and temporal distribution of low-frequency earthquakes in Japan. He found that while the occurrence of these events varies spatially in a systematic manner associated with large earthquakes, very few are found in the focal region of upcoming major earthquakes, though they may occur in the aftershock sequence.

The regional characteristics of each earthquake category discussed in the preceding section can be modeled in terms of variation in characteristic asperity size and degree of interaction. Several fundamental aspects of the asperity model can be appreciated by considering Fig. 19, which stems from an analysis of earthquake doublets in the Solomon Islands (LAY and KANAMORI, 1980). As shown in Fig. 19(a), the individual rupture zones

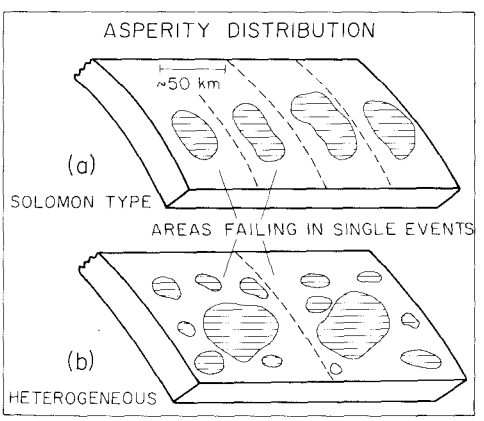

Fig. 19. Asperity model representation of coupling on the subducting slab. The Solomon Islands region has a uniform distribution of discrete, comparable size zones of strong coupling. Other regions, such as the Japan trench and Kurile Islands have a more heterogeneous stress distribution as in (b), (from LAY and KANAMORI, 1980).

of large events in the Solomon Islands are represented by a distribution of asperities. Failure of one of the asperities would produce an increase in stress on adjacent asperities. In the Solomon Islands region, the asperities are relatively large and of similar size, and the incremental stresses are large enough to trigger failure of an adjacent asperity. The scale length of the asperities, and the nature of stress release are such that failure occurs in discrete, but closely related events. The localized nature of the strong coupling is indicated by the simple, impulsive body waves and the small body wave source dimensions relative to the whole rupture zone found in the region. For regions with more complex stress distributions (Fig. 19(b)) the variation in asperity size inhibits efficient loading of adjacent large asperities, as much of the load is alleviated by failure of smaller asperities. Failure of the smaller asperities causes small stress increments to be induced on the larger asperities, and the stress on them builds up more gradually than in the Solomon Islands model. When the larger asperities fail, they may trigger failure of comparable size asperities within the same region, producing complicated multiple rupture events, but adjacent subzones are not subjected to the large sudden stress increments found when isolated large asperities fail, which inhibits the development of long ruptures.

Keeping these general characteristics in mind, it is possible to present the interpretation of the stress distribution in each subduction zone category given in Fig. 20. For the Chile-type behavior, the lithospheric plates are strongly coupled, and the asperity distribution is essentially uniform over the entire contact zone. When rupture initiates, it proceeds over the entire zone of uniform stress. One would expect that the seismic time function for such a failure would be a very long duration, single event. This is in fact the case for the 1964 Alaska earthquake. The Chile-type zones have relatively simple, geometrically uniform subduction regimes, with broad contact interfaces, which is probably favorable for developing uniform stress levels and strong coupling. The question of why the rupture stops for these zones may be answered by the presence of large lateral features which bound the rupture zones of great earthquakes in these regions, such as subducting ridges or changes in the interface geometry. These features may produce stress barriers which stop

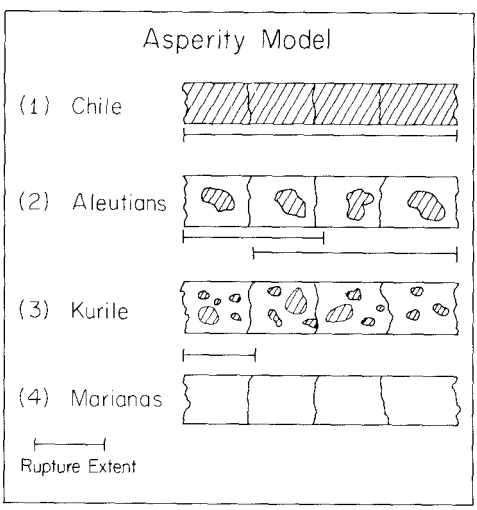

Fig. 20. An asperity model indicating the different nature of stress distribution in each subduction zone category. The hatched areas indicate the zones of strong coupling.

the rupture and fail later in separate events, or they may serve as strain energy sinks, in which the strain energy available for rupture is dissipated aseismically.

Slightly smaller, but still relatively homogeneous asperity distributions are proposed for the Aleutians-type category. The asperities are large discrete zones of strong coupling, and the failure of the larger asperities effectively loads adjacent zones causing large ruptures. The triggering of adjacent areas does depend on the previous history of the region, and if the stress on the adjacent asperity is very much lower than its strength, triggering may fail to occur. If conditions along the trench are uniform, the entire zone may fail, in a single event. If the asperities behave more or less independently, which is a function of the relative asperity size and separation, it is possible that the asperities will fail separately but may occasionally synchronize. The time functions determined by RUFF and KANAMORI (1980b) for the 1965 Aleutian and 1979 Colombian events (Fig. 18) indicate both types of behavior. The Rat Island results indicate sequential failure of two strongly coupled zones of 150–200 km in length, whereas the Colombian event appears to have involved failure of a single comparable size asperity. The previous occurrence of events to the south of the 1979 Colombian event may have caused a heterogeneous stress distribution unfavorable for triggering the adjacent zones. The asperity scale length for other zones of this category appears to be smaller, as for the Solomon Islands, and Nankai trough zones, but the general characteristic of large isolated asperities in each subzone appears to explain the observed temporal variation and triggering interaction found in these zones. If the asperity dimension relative to the overall subduction zone dimensions are further decreased, smaller rupture zones would be expected, but asperity interaction should still occur. This appears to be the case for the Middle America and New Hebrides zones which have temporally clustered, simple events with small body wave source dimensions relative to the entire rupture zone.

For the Kurile-type rupture mode, there are numerous variable size asperities within a given subzone (Fig. 20). Failure of the larger of these leads to slip throughout the subzone, but because of the relatively small size of the asperities, the stress increments communicated to adjacent zones are inadequate to cause further rupture propagation. The heterogeneous stress distribution produces complicated ruptures, with several discrete sources. This is consistent with the 1963 Kurile event shown in Fig. 18. The large percentage of aseismic slip in these regions may reflect the lack of strong asperities, and the slow rate at which they develop following an event.

The final category shown in Fig. 20 is the Marianas-type behavior for which there are no large asperities and hence no large earthquakes. The weak coupling on the fault plane produces essentially aseismic slip, regardless of the seafloor topography of the subducting slab.

The variation in rupture process along a given zone can easily be accommodated by variations in asperity size and interaction along the zone, as discussed by LAY and KANAMORI (1981). Another feature which can be related to the distribution of strength in each zone is the seismicity pattern characteristic of each subduction category. In the model presented by KANAMORI (1981), a bimodal distribution of strength in a given zone, combined with the gradual stress concentration on an asperity caused by the regional tectonic stress and the failure of smaller asperities, can qualitatively explain various seismicity patterns that are observed. Figure 21 shows the basic model presented by KANAMORI

(1981) and the effect of varying the parameters of the strength distribution. In this model, as the tectonic loading stress increases, there is scattered background activity, a precursory swarm, a period of quiescence and then foreshock-main shock activity. The most important parameters in the model are the distribution of strength within the asperity Σ_a, the ratio of the average asperity strength to surrounding fault strength $\bar{s}_a/\bar{s}$, and the distribution of surrounding fault strength Σ. These govern the foreshock activity, quiescence, and precursory swarms respectively. Comparison with the subduction zone categories in Table 3 indicates that category 1 is associated with a large asperity, which produces preshock activity. Category 2 is associated with small Σ_a, and large $\bar{s}_a/\bar{s}$, which produce a period of quiescence preceding the mainshock. Category 3 is associated with small $\bar{s}_a/\bar{s}$ and large Σ_a, which produces a characteristic swarm-quiescence-foreshock sequence. Thus, at least qualitatively, the asperity model can explain the seismicity patterns and

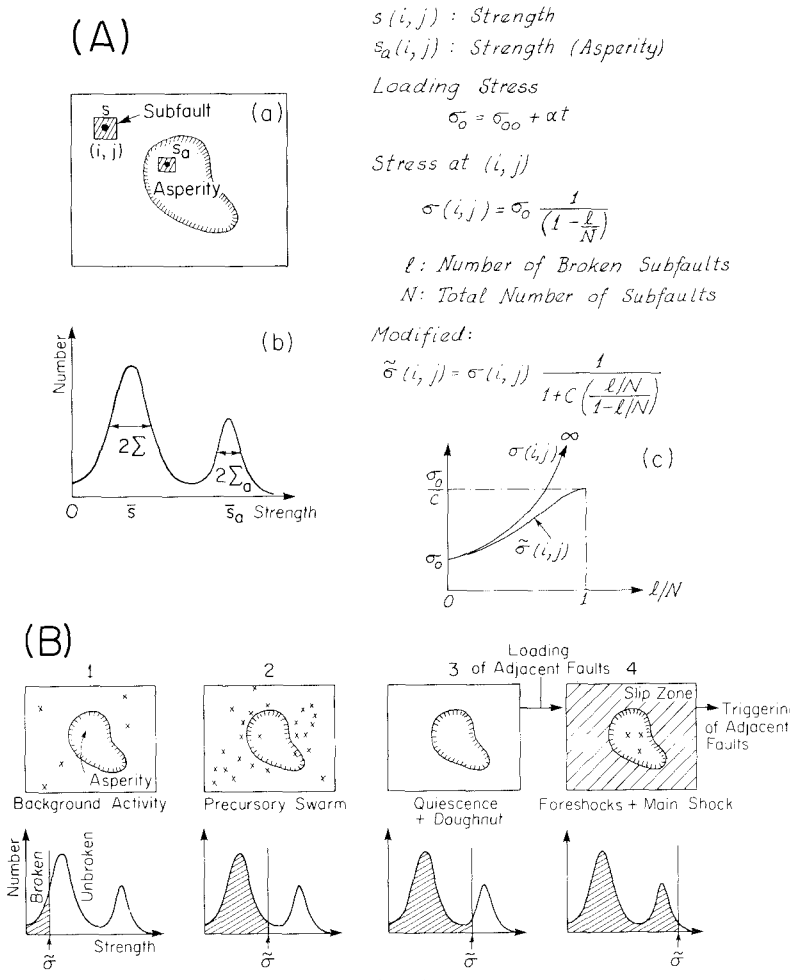

Fig. 21

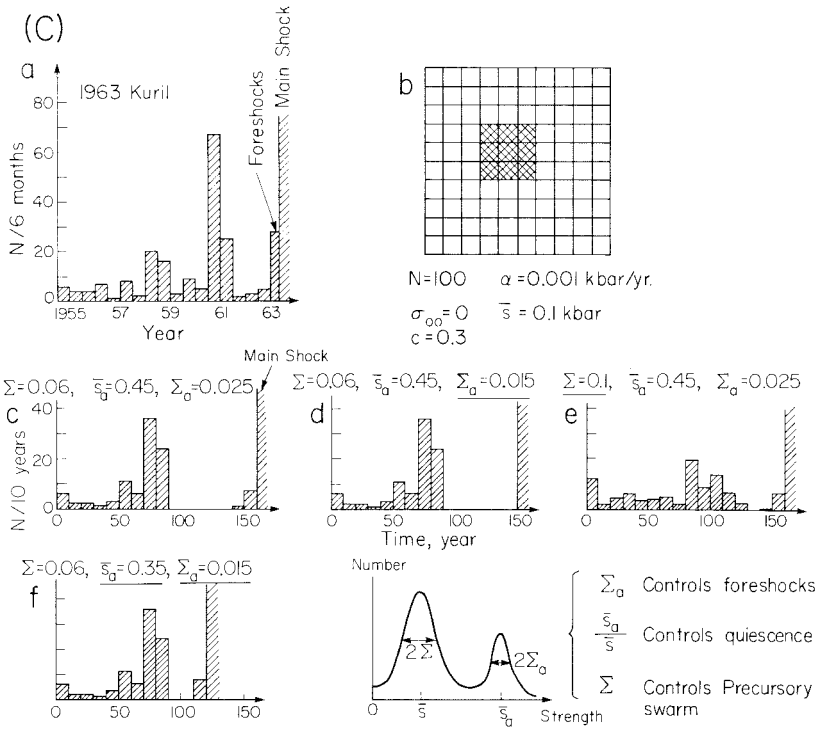

Fig. 21. (A) Schematic figure showing a unit fault containing an asperity (a), the distribution of strengths of subfaults within the asperity and the surrounding area (b), and the stress on the subfaults (c). (B) Sequence of seismicity patterns predicted by the asperity model in (A). (C) Comparison of temporal variation of number of events between the 1963 Kurile Islands sequence and the asperity models, (from KANAMORI, 1981).

rupture mode of large earthquakes in various subduction zones with a minimum of parameters.

5. Discussion

The characterization of the stress regime in each subduction zone presented above provides a useful framework in which to contrast regional behavior and in which to predict the future behavior of the large earthquake activity. As additional studies are performed, the categorization will be refined. Ultimately, it is desirable to determine the origins of stress heterogeneity and the mechanism of failure. It is of particular interest to determine whether the mode of subduction is controlled by local features such as topography on the subducted plate and orientation of the plate contact, or whether the degree of coupling is governed by more global features such as convergence rate, lithospheric age, and previous history of the zone.

The nature of coupling on the fault plane appears to be influenced by the history of subduction. Figure 22 reproduces the evolutionary subduction model proposed by

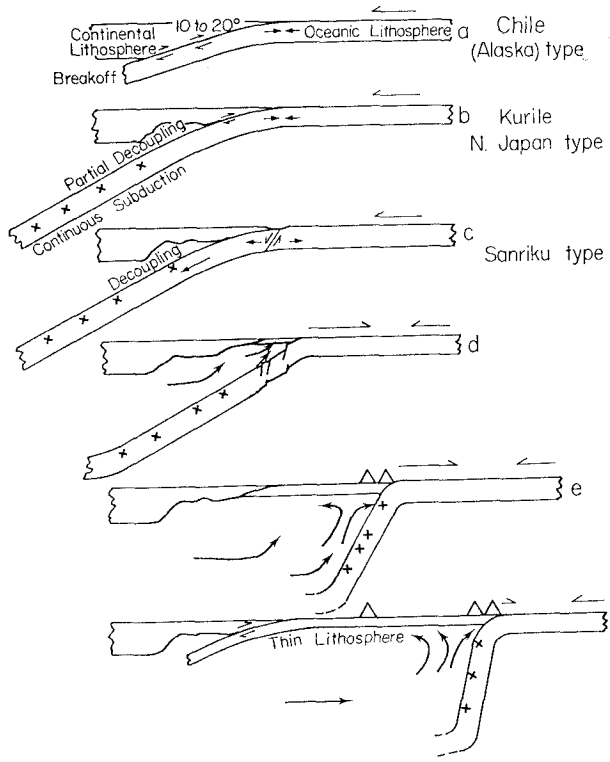

Fig. 22. The evolutionary subduction model proposed by Kanamori, (1977). (a) Strong coupling between oceanic and continental lithospheres results in great earthquakes and break off of the subducting lithosphere at shallow depths. (b) Partial decoupling results in smaller earthquakes and continuous subduction. (c) Further decoupling results in aseismic events and intraplate tensional events. (d) Sinking plate results in retreating subduction and formation of a new thin lithosphere. (e) Episodic retreat and formation of ridges. (f) Decelerated retreat and commencement of new subduction.

KANAMORI (1977). Shallow dipping, broad, strongly coupled zones such as in Chile and Alaska produce extensive ruptures. The thrust zone may be weakened and partially decoupled by repeated fracturing, yielding smaller rupture lengths as in the Kuriles and Japan trench. Large normal fault events such as the 1933 Sanriku earthquake may represent a transition to tensional stress in the slab and complete decoupling of the plate interface which may result in the development of back-arc basins by trench retreat. The variations in structure of the sedimentary wedges, upper slope basins, and terraces in the trench which correlate with maximum rupture lengths may reflect the transition between zones of strong and weak coupling on the fault plane.

UYEDA and KANAMORI (1979) associated the presence of active back-arc basin development along with subduction of old seafloor with aseismic slip. The Marianas, devoid of large earthquake ruptures, is cited as an example of a weakly coupled zone with ongoing aseismic subduction, whereas the Chilean boundary which generates large ruptures is strongly coupled and has no back-arc basin. If the subduction zone is weakly coupled,

the presence of topographic structures on the seafloor should exert less influence on the subduction process than if they intersect more strongly coupled zones. Thus, subduction of ridges and fracture zones in the New Hebrides and Chile trenches produces distinctive, slow seismic events, whereas these are not found along the Izu-Bonin and Marianas trenches. RUFF and KANAMORI (1980a) found further evidence for a correlation between active marginal sea formation with old subducting seafloor and low convergence rates.

RUFF and KANAMORI (1980a) tested correlations between seismic coupling, parameterized by regional maximum earthquake magnitudes, with other parameters of the subduction zone. They found that convergence rate and lithospheric age correlate with coupling, as shown in Fig. 23. This figure plots the results of linear multivariate regression indicating the correlation. It was also determined that penetration depth correlates with lithospheric age and horizontal contact length with convergence rate, results compatible with work done by ISACKS *et al.* (1968), VLAAR and WORTEL (1976), and WORTEL and VLAAR (1978). These correlations suggest that large earthquake generation correlates with interface geometry or with age and rate in some other manner. The dotted line in Fig. 23 encloses the subduction zones associated with back-arc spreading, showing the correlation of old seafloor and low convergence rate with these zones of low coupling. RUFF and KANAMORI (1980a) suggested a qualitative model in which convergence rate and lithospheric age determine the horizontal and sinking rates of the slab, thus controlling the dip of the Benioff zone and normal stress distribution on the fault plane. The area of coupling may change due to

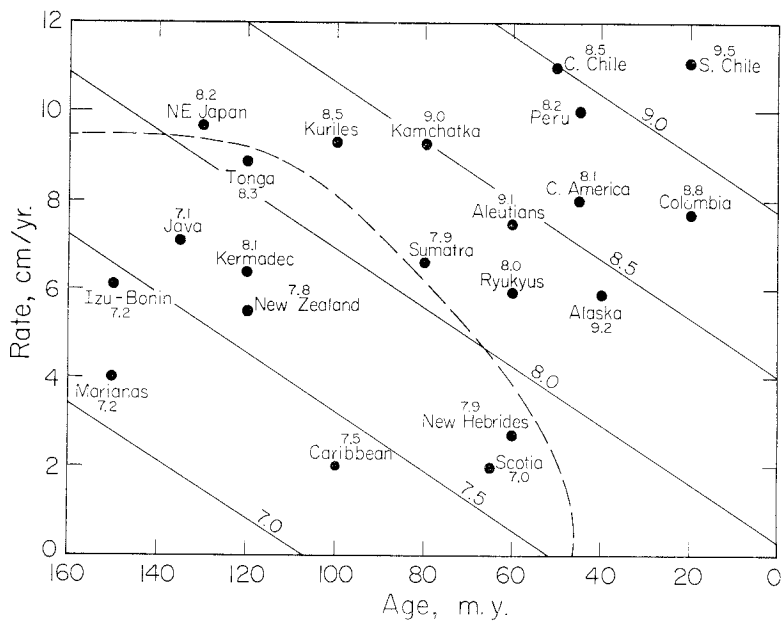

Fig. 23. The relationship of seismicity to the two variables, convergence rate and age of the subducting oceanic lithosphere. The number at each subduction zone is the associated M_w, and the contours of constant M_w define the resultant plane from the regression analysis. The broken line in the lower left corner delimits the subduction regions where there is either confirmed or suspected back-arc spreading, (from RUFF and KANAMORI, 1980a).

reducing cross section with increasing dip or degradation of the interface. This model indicates that the downgoing plate determines the stress state in the slab and on the interface, as suggested in the evolutionary subduction model of KANAMORI (1977). This is essentially similar to ISACKS and BARAZANGI's (1977) assertion that the mechanics of bending dominate in determining shallow Benioff zone geometry.

Figure 23 indicates that young lithosphere and large convergence rates give rise to great earthquakes, though Middle America and Peru obviously deviate from this trend. The relations indicated by this figure could clearly be influenced by the presence or lack of transverse features segmenting the subduction zone and inhibiting large earthquake ruptures, as well as by details of the regional asperity interaction with triggering being more efficient in some zones, accounting for the discrepancy between Middle America and the Aleutians. Further characterization of the stress regime in each region will help to understand these correlations and the scatter in the correlations.

6. Conclusions

Regional variations in the mode of rupture of large earthquakes and the degree of coupling on the fault plane have been reviewed. On the basis of maximum rupture length, seismicity patterns, percentage of aseismic slip, and source time function characteristics, it is possible to define four basic categories of subduction zone behavior. Category 1 includes the Alaska, southern Chile, southern Kamchatka and possibly the Central Aleutians zones, all of which produce great earthquakes with ruptures greater than 500 km long, pronounced preseismic activity near the epicenter, and large percentage of seismic slip. The 1964 Alaskan earthquake had very long duration source time function indicative of uniform rupture over a strongly coupled zone several hundred kilometers long, which may be representative of the failure process of the largest earthquakes. Category 2 includes the Western Aleutians, Colombia, Nankai trough, northern Kamchatka, and Solomon Islands zones, which have temporal variation in rupture mode with occasional ruptures reaching 500 km in length, and temporal clustering of large events. Events in these zones are usually preceded by seismic quiescence and the source time functions of body waves tend to consist of several long duration (30–60 sec) discrete ruptures. Several zones have similar discrete rupture events that cluster in time, but with characteristically smaller dimensions. These include the Middle America, New Hebrides and possibly central Kermadec, New Zealand, Taiwan and the Philippine trench zones. The rupture lengths in these zones are typically 100–150 km long, and body wave source rupture durations are 10–15 sec. The third category includes the Kurile Islands, Peru, central Chile and northern Japan trench zones, in which large earthquakes repeatedly rupture the same portion of the subduction zone in events with ruptures from 100 to 300 km long. These regions have a large percentage of aseismic slip, and complicated body waves resulting from multiple sources in each event. The fourth category includes the Marianas, Izu-Bonin, Java, Tonga, southern Kermadec, and Ryukyu zones, which do not produce great earthquakes. These zones subduct almost entirely aseismically. The asperity model is considered in order to explain these variations. The stress distribution in categories 1, 2, 3, and 4 are characterized by large asperities with strong coupling, large but discrete asperties, numerous smaller asperities, and an absence of significant asperities respectively.

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