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TRANSOCEANIC TSUNAMIS OBSERVED IN 1985

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ABSTRACT

This paper describes the 1985 transoceanic tsunami in the NW Pacific. In particular, it describes the 1985 Chilean tsunami which was generated by an earthquake about 18 nautical miles off Valparaiso, Chile. A strong shock occurred at 22h 6m GMT on March 3, 1985. The estimated earthquake magnitude was M=7.7 Richter scale. The generated tsunami hit the Chilean coast facing the Pacific while flooding also affected northwestern Pacific coasts. Tsunami propagation on the Chilean coast was associated with a tsunami height of 3m at Talcahuano located 440km south of the epicenter. At Valparaiso, the nearest tide station recorded the maximum tsunami height as 1.15m. North of the Chilean coast, the tsunami propagated at an average speed of 13 km/min between Coquimbo and Iquique (about 1430 km distance along the coast). This case illustrates tsunami propagation as a Kelvin wave (Nakamura, 1989).

INTRODUCTION

On coasts facing the ocean transoceanic tsunamis have repeatedly inflicted severe damage in the coastal zone. Some of the tsunamigenic earthquakes undersea have generated tsunamis close to the coast.

Nakamura (1991) described Chilean tsunamis in the northwestern Pacific and considered the dynamical effects of the planets and pole tide as well as statistical and chronological studies in tsunami research. Nevertheless, it remains difficult to predict the next hazardous tsunami event and to be ready to consider coastal protection measures.

Circum-Pacific Seismic Zone

As is well known, the seismic zone on the Earth is not distributed uniformly. Between the Far East area in Asia and north and south Americas, there is a well-defined circum-Pacific seismic zone. This zone is also characterised by tsunamigenic earthquakes. A tsunamigenic earthquake at or around the antipodes could create a tsunami hazard without any seismic signals being detectable within particular coastal areas. This problem should be noted (Nakamura, 1988, 1991) and related information should be distributed in advance.

As for the probability of tsunami occurrence, an estimate of the zones in the eastern Pacific has been given (Nakamura, 1986) although at present there is no comparable information regarding the impact of Chilean tsunamis in the northwestern Pacific.

Chronological tsunami catalogs

As for the chronology of Chilean tsunamis detected in the northwestern Pacific, one must refer to the existing data and available historical documents (eg Iida et al, 1967; Watanabe, 1985). Nakamura (1988, 1990) has discussed the probability of coastal floods in Japan caused by Chilean tsunamis and has estimated that the next hazardous Chilean tsunami may occur in 2081AD.

The 1985 Chilean tsunami on the Chilean coast

The 1985 Chilean tsunami was generated by an earthquake circa 18 nautical miles west of Valparaiso, Chile. The earthquake occurred at 22h 46m 54s GMT on 3rd March 1985. The location of its epicenter was 33° 14' 24"S and 72° 02' 24" W. The center of the earthquake was at the depth of 15km and the earthquake's magnitude was 7.7 in Richter scale. A tsunami warning was issued at 23h 02m GMT in Chile (Nakamura, 1992).

Seven tide stations on the Chilean coast recorded the tsunami generated by this earthquake. (Talcahuano, Valparaiso, Coquimbo, Antofagasta, Iquique and Arica) and this data enabled a propagation pattern of the tsunami to be determined (Nakamura, 1992).

Immediately west of the Chilean coast, the 200m bathymetric contour lies parallel to the coastline at a distance of 20km offshore, while the 300m contour is located 60km offshore. By contrast the oceanic trench deeper than 6000m is located at a distance of only about 80km offshore. The Chilean estimate of the highest tsunami height was 3m at Talcahuano, south of Valparaiso. The tsunami travelled about 440 km in 14 minutes and struck the coastal area between Valparaiso and Talcahuano. As an average estimate, the velocity of the tsunami was 0.5 km/s alongshore. The tsunami propagated also to the north along the coast at a speed of about 0.2 km/s. This means that the tsunami travelled the distance of 1430 km from Valparaiso to Arica in 100 minutes. This tsunami pattern can be explained dynamically by a qualitative tsunami model of Kelvin wave type (Nakamura, 1989).

Tsunamis in the northwestern Pacific

The 1985 Chilean tsunami in the northwestern Pacific propagated across the Pacific to strike coastlines facing SW towards the northwestern Pacific (Nakamura 1992). In a similar manner to the 1960 Chilean tsunami(Nakamura, 1987, 1990), the 1985 Chilean tsunami travelled across the Pacific to hit the Japanese Islands approximately 24 hours after the tsunami was generated in the source area.

In western Japan mareograms for this time period were analysed by courtesy of the Kobe Marine Observatory. The available records were obtained at nine tide stations (Muroto-misaki, Komatsujima, Sumoto, Kobe, Osaka, Tannowa, Wakayama, Shirahama and Kushimoto).

At Kushimoto and Muroto-misaki, the occurrence of local variations of sea level made it difficult to calculate the arrival time of the tsunami. At the other tide stations some faint sea-level variations were observed. According to Nakamura (1992), it was noted that the most likely arrival time at Shirahama was at 23h 30m GMT on 4th March 1985. The arrival time at Wakayama and Osaka is estimated to have been at 00h 50m GMT on 5th March 1985.

One of the major problems encountered has been the method of evaluation of the induced oscillations caused by the incident tsunami. An energetic study showed interesting results concerning the tsunami energy and the patterns of local oscillations (Nakamura, 1992). In Osaka Bay, an oscillation with a loop at Osaka and a node at the opening strait was detected. At the maximum amplitude in Osaka Bay, the tsunami energy was transformed into potential energy to increase the elevation of the sea level and to trigger oscillations in the bay. At Osaka, at the head of the bay, the evaluated potential energy for a unit water surface area was 1.2×10^4 erg/cm^2. A quite similar longitudinal oscillation was observed in the case of the 1960 Chilean tsunami, though the maximum tsunami height at Osaka during the latter event was about 1.5 m.

In the Kii Channel, longitudinal and lateral oscillations were induced. This can be explained by oscillations induced at Wakayama and Shirahama and at Wakayama and Komatsujima. The longitudinal oscillation between Wakayama and Shirahama was established as a result of the time that elapsed after the tsunami arrival. Similarly a lateral oscillation was also observed in the Kii Channel after the tsunami arrival. An estimated energy for the longitudinal oscillation is about 8×10^3 erg/cm^2. It is therefore concluded that the tsunami energy was transferred to local oscillations within two or three hours after the tsunami arrival (Nakamura, in press).

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EXPECTATION OF DESTRUCTIVE FAR-FIELD TSUNAMIS FROM THE ALEUTIAN-ALASKA SUBDUCTION ARC

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ABSTRACT

Available data on far-field tsunamis from the Aleutian-Alaska Subduction Arc were gathered and analyzed to detect recurrence patterns. Far-field destructive tsunamis from the Arc have been documented only during the 19-year period from 1946 to 1964. Far-field destructive tsunamis were absent during the nineteenth century; they may have occurred during the period 1788-1792, but documentation is lacking.

Previous attempts at segmenting the Aleutian-Alaska Arc into tectonic blocks by other investigators relying mostly on distribution of aftershocks of large earthquakes had resulted in blocks ranging in size from 100 km to 500 km in length. When the blocks were realigned using tsunami source area data, they turned out to have rather uniform size, roughly 300 kilometers in length.

Large earthquakes in the Aleutian-Alaska Arc occurred in four clusters during the last two centuries: in 1788-1792, 1844-1858, 1900-1917 and 1938-1965. Of these clusters only the earthquakes in 1788-1792 and 1938-1965 generated significant or noteworthy tsunamis. A cluster can consist of tsunami-generating earthquakes or of earthquakes without tsunamis.

The next cluster of large earthquakes should commence within two decades. From historical pattern, there is an even chance that the next cluster will be the tsunami-generating type or not.

In regards to predicting future tsunami hazards, the gap theory, so popular in discussing earthquake hazards, does not apply.

INTRODUCTION

The subduction zone known as the Aleutian-Alaska Arc has the reputation of being the most seismically active subduction zone in the world. In the twentieth century, the earthquakes of April 1, 1946, March 9, 1957 and March 28, 1964 generated tsunamis that have wrought death and destruction in the far-field over a wide area (Dudley and Lee, 1988). In this study an attempt is made to assess the expectation of tsunamis of comparable destructiveness from the Aleutian-Alaska Arc.

Tsunamis generated by volcanic earthquakes and explosions in the Aleutian-Alaska Arc have been excluded because such tsunamis have not been destructive in the far-field (Cox and Pararas-Carayannis, 1969).

DEFINITIONS OF TERMS.

In this study, several words and phrases that are used frequently in tsunami studies and in tsunami damage reports will be defined quantitatively to remove ambiguity and vagueness. The definitions were chosen to be useful to local and regional emergency management agencies, which have the responsibility of undertaking response measures during a tsunami alert.

Damaging Tsunamis and Destructive Tsunamis

The words "damaging" and "destructive" when used to describe tsunami effects are culture dependent and very subjective. If tsunami inundation had penetrated several hundred meters inland at a place where there was no inhabitant or man-made structure, the effect would be labelled as "no-damage". If the same horizontal extent of inundation had flooded a sea port or a fishing village however, the damage would be labelled as "severe." For this paper, a damaging tsunami is defined as one that will cause a run-up of over 1 m of vertical height, measured from sea level at the time of arrival of the tsunami. A destructive tsunami is defined as one with a run-up of over 2 m. From past experience in Japan (Okada et al., 1990), a 1-m wave will damage ponds, fish holding pens and equipment belonging to aquaculture and mariculture industry. A 1-m wave is the threshold of damage. A 2-m wave will flood and damage near-shore houses and commercial buildings, dislodge ships from moorings, and cast them adrift, and, most important, a 2-m wave can claim human lives.

Near-Field, Intermediate-Field and Far-Field

The near-field is defined as that region within 250 km of a tsunami source. The far-field is defined as those regions more than 1000 km from a tsunami source. The region between 250 km and 1000 km will be referred to as the intermediate-field. Hence, a destructive far-field tsunami is one that has run-ups of over 2 m at distances greater than 1000 km from the source.

Destructive tsunamis in the Pacific Ocean occur frequently, about once every two years (Utsu, 1990), but most of the destruction and loss of human life is limited to the near-field. Near-field destructive tsunamis do travel across the ocean, but upon arrival at distant shores, the run-ups of most tsunamis are below 2 m.

The Aleutian-Alaska Arc

The Aleutian-Alaska Arc is considered to be the subduction zone bracketed by the geographical coordinates of 55°N, 165°E and 60°N, 145°W. The subduction zone includes the shelves south of the Aleutian Islands, of the Alaska Peninsula, and of the Middleton and Montague Islands. The seismic zones of Yakataga and Yakutat are excluded from this study because they are not part of the arc structure.

Giant, Major, and Minor Tsunamis

In Furumoto (1991) tsunamis were classified according to the extent of destruction and damage in the far-field. A giant tsunami causes run-ups of over 2 m in the far field over a widespread area, for example, throughout the Hawaiian Islands. Examples of giant tsunamis are the Aleutian tsunamis of 1946 and 1957, the Alaska tsunami of 1964, and the Chile tsunami of 1960.

A major tsunami causes run-ups of over 2 m in the far field only in those bays and inlets highly receptive to tsunamis because of response characteristics of that bay or inlet, and a 1-m run-up over a wide area. An example of a major tsunami is the one from Japan in 1933 which had run-ups over 2 m in the far field only along the Kona Coast of the Hawaiian Islands.

In a minor tsunami, run-ups in the far field will be less than 1 m. A minor tsunami can cause destruction in the near field, but are usually harmless in the far field.

From the point of view of tsunami source parameters, giant tsunamis were generated by earthquakes which had moment magnitudes M_w or magnitude by tsunami data M_t (see footnote) equal to or greater than 9 and which had rupture areas with the major axes 550 km or longer. Major tsunamis were generated by earthquakes with M_w or M_t between 8.4 and 9 with source areas of 240 km to 500 km in length. Minor tsunamis occurred when moment magnitudes were less than 8.4 and rupture lengths were less than 240 km (Furumoto 1991).

FAR-FIELD TSUNAMIS FROM THE ALEUTIAN-ALASKA ARC

Table 1 lists tsunamis from the Aleutian-Alaska Arc which have been observed, documented, or recorded in the far field, irrespective of the size of the tsunami. The list has been compiled from catalogs of earthquakes and tsunamis, namely those published by Iida et al. (1967), Pararas-Carayannis (1967), Cox and Pararas-Carayannis (1969), Coffman et al. (1982), Ganse and Nelson (1981), Lander and Lockridge (1989), Soloviev and Go (1984 a, b), and Utsu (1989). Data on M_w and M_t were compiled from publications by Abe (1979), Kanamori (1977), and Pacheco and Sykes (1992). Although documentation of earthquakes and tsunamis has been very poor in the Aleutian Islands and in Alaska during the eighteenth and nineteenth centuries, far-field tsunamis are the exception because of consistent records kept in the Hawaiian Islands since 1830. The islands played a prominent observation point as they are

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Magnitude by tsunami data M_t was defined by Abe (1979) and is numerically equivalent to moment magnitude M_w (Abe 1981). It should not be confused with tsunami magnitude m defined by Iida (1957).

TABLE 1

FAR-FIELD TSUNAMIS FROM THE ALEUTIAN-ALASKA ARC

YEAR	MO-DAY	EPICENTER	Mw or Mt	RUN-UP HAWAII (m)	RUN-UP W COAST, N. AMERICA (m)	RUN-UP OTHER PLACES (m)
1854	Jan 27	Kodiak Is	?	Hilo observed		
1872	Aug 23	Fox Is	?	Hilo 1.3	Oregon, Calif. <0.2	
1929	Mar 7	51 N 170 W Fox Is	8.1	Hilo 0.2		
1938	Nov 10	55.5 N 158 W	8.4	Hilo 0.3	Santa Monica, CA. 0.1	
1946	Apr 1	52.8 N 163.5 W	9.1	Hawaii 17	Half Moon Bay, CA 3.5	Samoa 2.4 Japan 1.1
1957	Mar 9	51.5 N 175.7 W	9.1	Kauai 16 Hilo 4.2	Bodega Harbor CA, 1.7	Japan 3 Marshall 0.3
1964	Mar 27	61.1 N 147.7 W	9.2	Oahu 4.8 Hilo 2.1	Crescent City, CA 6.3	New Zealand 0.6 Japan 0.7
1965	Feb 4	51.3 N 178.6 E	8.7	Kauai 1.1	Crescent City, 0.3	Japan 0.4
1965	Mar 10	50.6 N 177.9 E	7.3	Hilo 0.1		
1986	May 7	51.5 N 174.8 W	8.0	Kauai 0.6	Crescent City <0.1	
1987	Nov 30	58.7 N 142.8 W	7.8	Kauai .12 Hilo .15	San Francisco 0.05	
1988	Mar 6	57.0 N 143.0 W	7.3		San Francisco <0.1	

strategically located in the middle of the North Pacific, and all tsunamis from the Arc can sweep into the islands unhindered by intervening islands, reefs, or shoals. In addition the residents of Hawaii have keenly observed any unusual agitation of the seas around them and have duly recorded them.

The earliest documentation of a far-field tsunami from the Aleutian-Alaska Arc was in 1854. The tsunami, generated by an earthquake in the region of Kodiak Island (Lander and Lockridge, 1989), was duly observed in Hilo Bay in Hawaii, although no quantitative estimate as to wave height was recorded.

By August 23, 1872, when the next Pacific-wide tsunami from the Arc occurred, tide gages had been installed in the Pacific region. From records on the tide gages of Honolulu Harbor in Hawaii, Astoria in Oregon, and San Francisco and San Diego in California, Cox (1984) was able to determine the tsunami source by tracing the waves back to the origin. Although the earthquake had occurred in the vicinity of the Fox Islands, there is no record of such an earthquake in Alaskan archives, simply because documentation was poor. Hawaiian records, however, had registered a tsunami on that day and had listed it as a tsunami from an unknown source for over a century until Cox (1984) determined the source.

The tsunamis of 1854 and 1872 are the only two far-field tsunamis definitely known to have originated from the Aleutian-Alaska Arc during the nineteenth century. These two were far from being destructive. In contrast, during the twentieth century, there have been ten far-field tsunamis and three of them, in 1946, 1957, and 1964, have been destructive.

The uneven time distribution of far-field tsunamis is real. Some seismologists have attributed the asymmetric distribution to poor documentation, but as far as tsunamis are concerned, documentation was done by observers outside of Alaska, by observers in Hawaii, the west coast of the United States and the Orient. The fact stands that there was no far-field destructive tsunami from the Aleutian-Alaska Arc during the nineteenth century. Furthermore, the only three giant tsunamis from the Aleutian-Alaska Arc occurred within the 19-year period from 1946 to 1964, a rather unusual concentration of energy and moment release within a very short time. Attempts at finding any quasi-periodicity in tsunami occurrence or in high-moment earthquake recurrence will have a very remote chance of success.

SOURCE AREAS OF LARGE TSUNAMIS

Several investigators have determined tsunami source areas by retracing initial tsunami phases from numerous tide gages back to their origins, such as the retracing done for the 1964 tsunami by Pararas-Carayannis (1967), who found the long axis of the source area to be 700 km. Table 2 lists the dimensions of the long axes of the tsunami source areas so far recovered from the Aleutian-Alaska Arc. Figure 1 shows the locations of the tsunami sources. The 1986 source area is an exception; the moment release area determined by Boyd and Nabelek (1988) from seismic P waves was assumed to be equivalent to the tsunami source.

The source areas of the 1938, 1946, and 1964 tsunamis are neatly aligned touching one another. The source area of the 1957 tsunami is so large that it includes all the source areas of the 1872 and 1986 tsunamis and the eastern half of the 1965 tsunami. In Table 2, the sixth column of "Blocks" refer to tectonic blocks which are explained in the next section.

TABLE 2.

DATA ON LENGTHS OF TSUNAMI SOURCE AREAS

YEAR	MO-DAY	WEST END	EAST END	LENGTH (km.)	BLOCKS	REFER-ENCES
1872	Aug 23	171.5 W	168.3 W	240	F	Cox 1984
1938	Nov 10	160 W	155 W	390	I	Hatori 1981
1946	Apr 1	168 W	159 W	570	G + H	Hatori 1981 Furumoto 1991
1957	Mar 9	179 E	168 W	900	D + E + F	Hatori 1981
1964	Mar 28	155 W	143 W	700	J + K	Pararas-Carayannis 1967
1965	Feb 4	175 E	177 W	600	C + D	Hatori 1981
1986	May 7	176 W	174 W	140	Part of E	Boyd and Nabelek 1988

In the above data, we have excluded rupture areas inferred from aftershock distributions. In discussions of rupture areas and tsunami sources, it is usually assumed that the rupture area of a tsunamigenic earthquake is the tsunami source area and that the aftershock area of a large earthquake represents the rupture area. For tsunamigenic earthquakes, aftershock areas have often been equated to tsunami sources. This syllogism is not valid for the Aleutian-Alaska Arc. In the earthquake of 1964, a good fraction of the aftershock area was on land and hence could not be part of the tsunami source. In the earthquake and tsunami of 1946, the aftershock area was only 120 km by 100 km (Sykes 1971), whereas the tsunami source area by retracing was 570 km by 120 km (Furumoto 1991). For the 1957 earthquake and tsunami, the long axis of the aftershock area was 1100 km (Boyd and Nabelek 1988) whereas the long axes of the tsunami source was 900 km (Hatori 1981). In general, equating aftershock area with rupture area or slip area for tsunami-generating earthquakes is not valid because much of the slip in tsunami-generating subduction earthquakes occurs in the accretionary prism whereas aftershocks occur along the contact interface between the down-going slab and continental slab (Pelayo and Wiens, 1992).

TECTONIC BLOCKS

Stauder (1972) recognized that the level of seismicity in the Aleutian-Alaska Arc varied from area to area so that the seismicity of one area was quite independent of the seismicity of another area. He called these seismically independent areas tectonic blocks. He identified a boundary between blocks at 179.9 E longitude. Nishenko and Jacob (1990) divided the Aleutian-Alaska-Queen Charlotte Island subduction zones into 17 "segments" which are conceptually equivalent to the tectonic blocks proposed by Stauder. For identification the blocks were numbered from the east, starting with Segment 1 as the eastern most block and ending with Segment 17 at the juncture of the Aleutian-Alaska Arc and the Emperor Seamount Chain. Of the 17 segments, 11 constitute the Aleutian-Alaska Arc, starting from Segment 7 at Prince William Sound. Nishenko and Jacob's partitioning was based on what they considered to be rupture zones of large earthquakes derived from aftershock distributions.

Upon comparison of their segments with tsunami source areas, we found generally good agreement. The only discrepancy was in their Segment 11, the "Unimak Island" segment, which they called a gap. When we realigned the blocks on both sides of the Unimak Island segment according to tsunami sources, the blocks fell neatly into place, as shown in Figure 2. The tectonic blocks all have nearly uniform size, with the long dimension roughly 300 km. In the segmentation by Nishenko and Jacob, the Unimak Island segment was exceptionally shrunken with an east-west length of about 100 km. The realigned blocks were identified by roman letters, starting from the western end of the arc, in tune with a convention started by Ishibashi (1981) for blocks in the subduction zone off the southern coast of Japan. Another reason for using roman letters is to avoid confusion with Nishenko and Jacob's segments.

As the tectonic blocks in the arc are roughly 300 km long, the rupture length L of a characteristic earthquake will be about 300 km. For a rupture length L of 300 km, the seismic moment M_o will be in the order of 50 x 10^{20} N-m according to the formula by Ward (1982),

$$L = (Mo^{.33})/5.684 \times 10^{4} . \qquad (1)$$

This implies that a characteristic earthquake would have a moment magnitude M_w of 8.4. This value coincides with the threshold for what we defined as major tsunamis.

The concepts of characteristic earthquake and, as a consequence, tectonic block, have recently been questioned by Kagan (1991,1993). Kagan's objections have been incorporated into this study. By characteristic earthqauake, I mean an earthquake that ruptures a geological entity called a tectonic block. However, the constraint that a characteristic earthquake is the maximum magnitude earthquake for a block has been removed, as data presented here demonstrate that a very large earthquake can rupture two or three tectonic blocks simultaneously.

SPACE-TIME DISTRIBUTION OF LARGE EARTHQUAKES ALONG THE ARC

It is quite natural that the question, "How often can we expect a far-field destructive tsunami from the Aleutian-Alaska Arc?" should occur in a study of this sort. It logically leads to a discussion of arc seismicity that has been the topic of numerous studies (e.g. Boyd and Lerner-Lam, 1988; Davies and House, 1979; Davies et al., 1981; Duda, 1965; Engdahl, 1977; House et al., 1981; House and Jacob, 1983; Kelleher, 1970; McCann et al., 1979; Nishenko and McCann, 1981; Sykes, 1971) and has spawned the discussions about seismic gaps (Sykes, 1971; McCann et al., 1979).

As the concern here is far-field destructive tsunamis which are generated by earthquakes of large seismic moments, consideration is restricted to earthquakes with moment magnitudes or surface wave magnitudes greater than 7.5.

Figure 3 is a plot of space-time distribution of tsunami sources and rupture lengths of large earthquakes. The solid lines represent known tsunami sources and dashed lines represent rupture areas of earthquakes that did not generate tsunamis. Tsunami sources after 1930 plus that for the 1872 earthquake are reliable, as they are sources derived by calculations from hard copy records, but the other sources and rupture areas are estimates from indirect data and historical accounts. The information up to 1929 was compiled from the results published by Davies et al. (1981), Boyd and Lerner-Lam (1988), Nishenko and Jacob (1990) and Estabrook and Boyd (1992).

Citing Russian historical documents and their translations, Davies et al. (1981) made a convincing case that the rupture area of the earthquake of 1788 extended from Sanak Island (54° N, 162.5° W) to the northern part of Kodiak Island (59° N, 152° W), a distance of about 800 km and a segment corresponding to the combination of tectonic blocks H, I, and J. As impressive tsunami actions from Sanak Islands to Kodiak Island were described in the documents, I judged that the tsunami source did extend over those blocks. The tsunami must have been destructive in the far-field too, but unfortunately for historians, means of observing and recording the events did not exist in the far-field. Even in Hawaii there is no record of a tsunami arrival, for the simple reason that Hawaii was then in the throes of recurring tribal wars that ended only in 1795 with the unification of Hawaii under King Kamehameha I. Along the western coast of North America, except for the Russian colonies in Kodiak, tsunamis and related marine phenomena were apparently not documented at that time. The 1788 event must have released more energy and moment than the Great Alaska Earthquake of 1964, as the 1788 event ruptured three blocks compared to the 1964 event of two blocks.

The rupture length designated as 1847-1848 is not due to one earthquake. According to Davies et al. (1981) and Sykes and Quittmeyer (1981) a series of earthquakes in 1847 and 1848 resulted in a cumulative rupture area as shown in the diagram. This accumulation amounted to the fracturing of tectonic block I.

From 1900 to 1917 a series of major and great earthquakes occurred along the length of the Aleutian-Alaska Arc. Davies et al. (1981) claimed that the entire arc had fractured piece by piece during this "turn of the century" seismic series, however, none of the earthquakes was large enough to qualify as a characteristic earthquake. Oddly, the various tsunami catalogs are silent even about local tsunamis generated by earthquakes of this series.

The rupture line designated as 1917 in Figure 3 in block H is again the cumulative fracture area of a series of earthquakes in 1917 (Boyd and Lerner-Lam 1988; Estabrook and Boyd, 1992), similar to the 1847-1848 series. The largest shock occurring on May 31, 1917 had a seismic moment of about 2×10^{20} N-m equivalent to moment magnitude of 7.5, but the rupture area was undeterminable (Estabrook and Boyd, 1992). The cumulative rupture area of shocks from May 31 to July 25 stretched from 53.8° N, 158.9° W to 54.5° N, 163.6° W (Boyd and Lerner-Lam, 1988), a distance of about 300 km. Nishenko and Jacob (1990) listed the tsunamigenic earthquake of 1929 as a moment magnitude of 7.8. By applying Equation (1) the rupture length would be about 140 km, which was plotted for 1929.

From 1938 to 1965, during a period of 28 years, one characteristic and four multiple block earthquakes occurred along the arc. An unusual phenomenon in the cluster was the overlapping of the eastern half of the tsunami source of the 1965 event and the tsunami source area of the 1957 earthquake. This overlapping implies that tectonic block D ruptured in 1957 and in 1965, twice within a period of eight years. This phenomenon appears to contradict stress accumulation and seismic gap theories. An explanation will be given in the Discussion.

The tsunamigenic earthquake of 1986 was not a characteristic earthquake because block E did not rupture entirely. Analyses of long-period P waves showed a moment release length of about 145 km (Boyd and Nabelek 1988), whereas analysis of Rayleigh waves by the Doppler method showed a rupture length in the order of 100 km (Furumoto and Yoshida, 1990). Consistent with the criterion for far-field destructive tsunamis, the resulting tsunami did not attain a run-up height of 2 m anywhere in the far-field.

Figure 4 shows a plot of moment release of large arc earthquakes against time. Seismic moments were obtained in various ways. When rupture lengths were known, seismic moments were obtained by equation 1. For the earthquakes in the cluster of 1900 to 1917, surface wave magnitudes were equated to moment magnitudes and seismic moments calculated. These methods are crude methods for estimating seismic moments, but they are justifiable because of the differences in orders of magnitude. In the final tabulation, as seen in the figure, significant moment releases occurred only in the events of 1788, 1847, 1872 and during the cluster from 1938 to 1965.

Similar to time distribution of giant tsunamis, we notice a concentration of great moment release during the 28-year period from 1938 to 1965. To find a comparable moment release prior to the cluster one must go back as far as 1788.

DISCUSSION

In Figure 3 four clusters of earthquakes can be recognized: (I) 1788-1792; (II) 1844-1858; (III) 1900 - 1917; and (IV) 1938 - 1965. Table 3 lists the properties of each cluster, namely the duration, median years, intervals between clusters, intervals between median years. For convenience we identify the clusters with roman numerals according to chronological order. Figure 3 and Table 3 agree with the proposition by Kagan and Jackson (1991) that large earthquakes cluster in time.

Large earthquakes have occurred singly between two clusters. The earthquake of 1872, which generated a far-field tsunami, occurred between Cluster II and Cluster III and the earthquake of March 7, 1929 with Ms 8.1 occurred between Cluster III and Cluster IV. Between Clusters I and II a great earthquake may have occurred and remained unrecorded. Perhaps the earthquake of May 7, 1986 with Ms 8.0 is the large earthquake between

TABLE 3.
EARTHQUAKE CLUSTER DISTRIBUTION

CLUSTER	YEARS	DURATION (years)	MEDIAN YEAR	INTERVAL BETWEEN CLUSTERS	INTERVAL BETWEEN MEDIANS
I	1788-1792	5	1790		
				42	61
II	1844-1858	15	1851		
				52	57
III	1900-1917	18	1908		
				21	44
IV	1938-1965	28	1952		
				27+	40+

Cluster IV and the next cluster.

There are differences in patterns and characteristics among the clusters. The intervals between median years have become shorter over the centuries, from 61 to 44 years. On the other hand, durations of clusters have lengthened progressively. In Cluster I and IV tsunami-generating earthquakes predominated, whereas in Cluster II and III together only the earthquake of 1854 generated a tsunami.

But the most salient trait is the relation of tsunamis to a cluster, which consists of either tsunami-generating earthquakes or non-tsunami earthquakes. In Clusters I and IV all member earthquakes were tsunami-generating, whereas in Clusters II and III together only the earthquake of 1854 generated a tsunami. None of the clusters had a mix of the two types of earthquakes. This clear dichotomy is, at first consideration, rather implausible. One naturally expects a mixed cluster to occur. Note that we have limited our examples to only large earthquakes. We have excluded volcanic earthquakes from our consideration because they do not have far-field effects.

The dichotomy can perhaps be explained in terms of what Kanamori (1972) called "tsunami earthquakes." Pelayo and Wiens (1992) have further made the distinction between "tsunami earthquakes" and "tsunamigenic earthquakes." In a tsunami earthquake, most of the rupture or slip occurs in the accretionary prism (Figure 5) of the subduction zone and less in the interface between the continental plate and the down-going oceanic plate. In a tsunamigenic earthquake, most of the rupture occurs in the interface between the plates. In a non-tsunami earthquake the rupture fails to extend into the accretionary prism. A rupture in the accretionary prism will move the ocean floor more than a rupture in the interface, and hence generate larger tsunamis. A good example of a tsunami earthquake is that of 1946, which had a surface wave magnitude of 7.3 but a tsunami moment magnitude of 9.1 to 9.3 (Abe, 1979). The surface wave magnitude indicates the amount of energy expended in the rupture along the interface whereas the tsunami moment magnitude reflects the large slip in the accretionary prism. An example of a tsunamigenic earthquake is that of 1965 with a focal depth of 40 km, in which the rupture occurred mostly along the deeper part of the interface. Although the rupture was 600 km long, the generated tsunami was not destructive in the far field.

During Clusters III and IV 80% of the Aleutian-Alaska Arc had ruptured. It is not possible to estimate the amount of rupture during Clusters I and II because of poor documentation. In Cluster IV the blocks from C to K had ruptured but blocks A and B remained untouched. This has led Davies et al. (1981) to classify blocks A and B as seismic gaps, but Jacob et al. (1977) had argued previously that these blocks have naturally low seismicity because of their proximity to the Emperor Seamount Chain, which stalls the subduction process.

From Figure 3 we draw the inference that all nine tectonic blocks from C to K were ruptured during the Cluster IV surge of moment release from 1938 to 1965. This phenomenon has been noticed and remarked by other investigators but no plausible explanation has been offered to date. Another observation on Figure 3 is that seismic gaps have disappeared. The much discussed Shumagin Gap is located in block H. We note that the block has ruptured in 1788, in the cluster of 1847-1848, in the cluster of 1917, and in the 1946 earthquake. In comparison with other blocks, the Shumagin block must have had all of its strain energy evaporated by 1946.

Kagan and Jackson (1991) have argued that identification of seismic gaps and assigning probabilities to rupture of the gaps differ little from random predictions. This set of data lends support to their thesis.

From the point of view of expectation of Pacific-wide tsunamis from the Aleutian-Alaska Arc, as 28 years (1993) have passed since 1965, it is time that we be on the alert for resumption of the next cluster activity. If we are fortunate, the new cluster may occur in bits and pieces with no danger, as was the case for the "turn of the century" cluster. On the other hand the violent clusters of 1788-1795 or 1938-1965 could be revived.

We need to emphasize that we are discussing the expected danger of Pacific-wide tsunamis which are generated by earthquakes with moment magnitudes greater than 8.4. The arguments advanced here should not be extrapolated to earthquake hazards of the Aleutian-Alaska Arc, because damaging earthquakes can have magnitudes as low as 5.5.

The present study reinforces the proposal by Kagan and Jackson (1991) that large earthquakes occur in clusters. At least for the Aleutian-Alaska Arc, any attempt at predicting an individual large earthquake, such as estimating the time of rupture of a specific seismic gap, appears to be meaningless, since a predicted earthquake will occur as a member of a cluster.

A logical approach to prediction will be to estimate when the next cluster will commence. Statistically this is not a good approach because we have documentation on only four clusters, and adequate data are available on only Clusters III and IV. However, it is reasonable to infer from the limited samples of Table 3 that the median years of the clusters are roughly half a century apart.

CONCLUSIONS.

Major and great earthquakes in the Aleutian-Alaska Subduction Arc occurred in clusters during the last two centuries. Four clusters have been recognized: Cluster I, 1778-1792; Cluster II, 1844-1858; Cluster III, 1900-1917; and Cluster IV, 1938-1965. In regards to tsunamis, member earthquakes in any one cluster have been all either the tsunami-generating type or the non-tsunami type. A mixed cluster has not been recorded. The only exception was the earthquake of 1854 which generated a minor tsunami during Cluster III. In other words a cluster is either a series of tsunami-generating earthquakes or a series of non-tsunami earthquakes.

In attempting to find quasi-periodicity in cluster occurrence, we found that the best index is the median year of a cluster. Median years are four to six decades apart, on the average 52 years. Clusters are approximately one-half of a century apart.

Using data on tsunami source areas, we were able to divide the Aleutian-Alaska Arc into 11 tectonic blocks, naming them alphabetically from west to east. During Cluster IV, the Arc had ruptured from block C to K. In geographical terms the ruptured section extended from Kiska Island in the Rat Islands Group in the west to Prince William Sound in the east. This implies that there is no preferred seismic gap in the Arc. Blocks A and B have low seismicity because of the proximity of the Emperor Seamount Chain (Jacob et al. 1977), and hence they also should not be considered potential seismic hazards at the present time.

The best tsunami prognostication that can be made about the Aleutian-Alaska Subduction Arc is that within two decades the next cluster should start occurring. If historical pattern holds, the next cluster will start with the rupture of block I or J, that is, in the block just east of Shumagin Island or the block on the western side of Kodiak Island. Although there is an even chance that the next cluster will be either a tsunami type or a non-tsunami type, the initiating rupture at block I or J should set the pattern for the coming cluster.

ACKNOWLEDGMENTS

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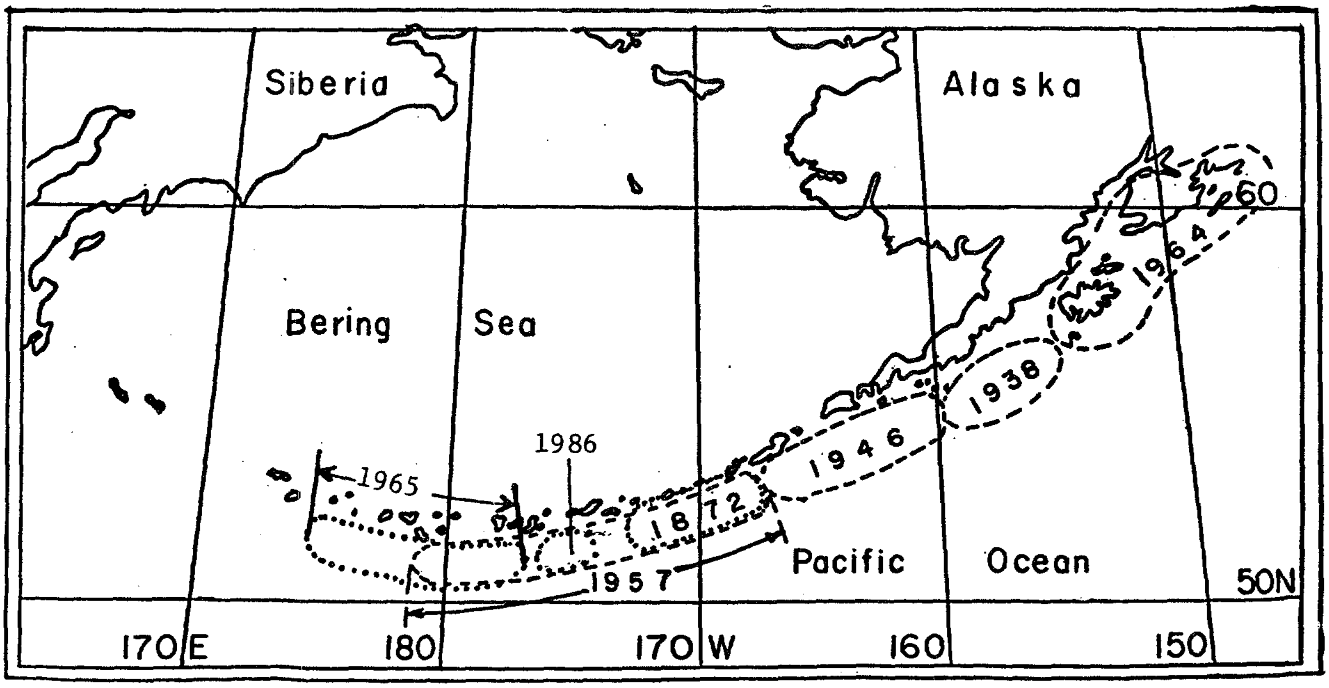


Figure 1. Tsunami source areas along the Aleutian-Alaska Arc.

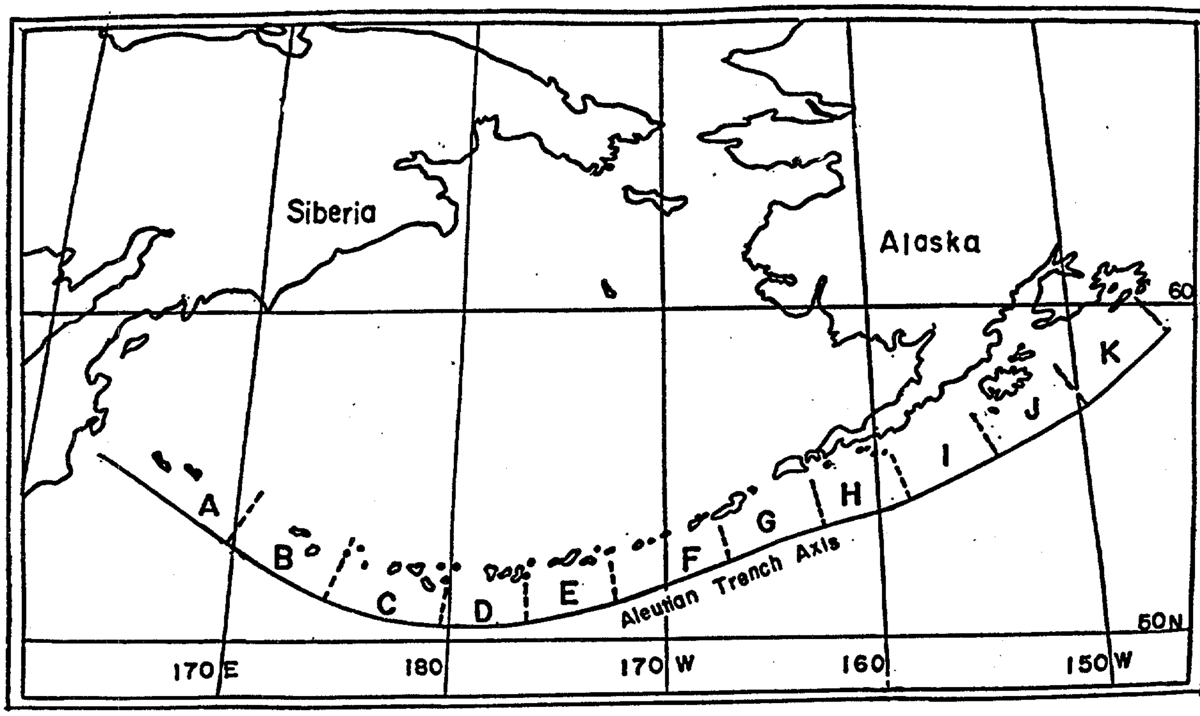


Figure 2. Tectonic blocks A to K along the Aleutian-Alaska Arc.

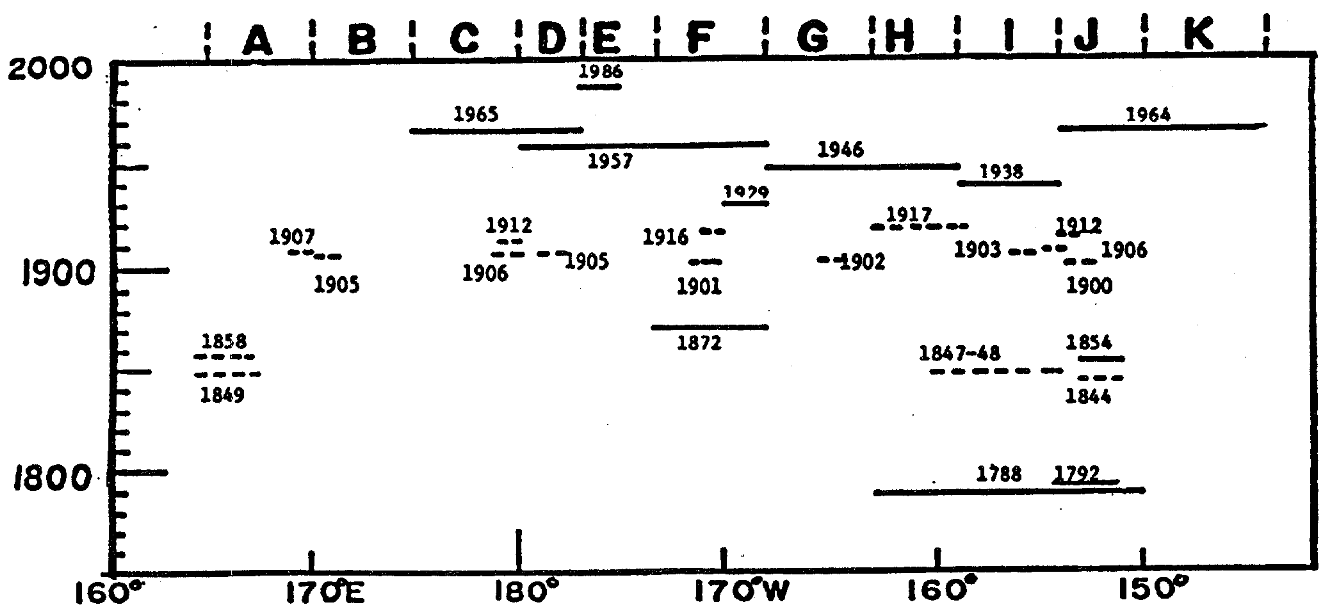


Figure 3. Tsunami sources and rupture lengths of large earthquakes along the Aleutian-Alaska Arc. Solid lines represent tsunami sources; dashed lines represent ruptures that did not generate tsunamis.

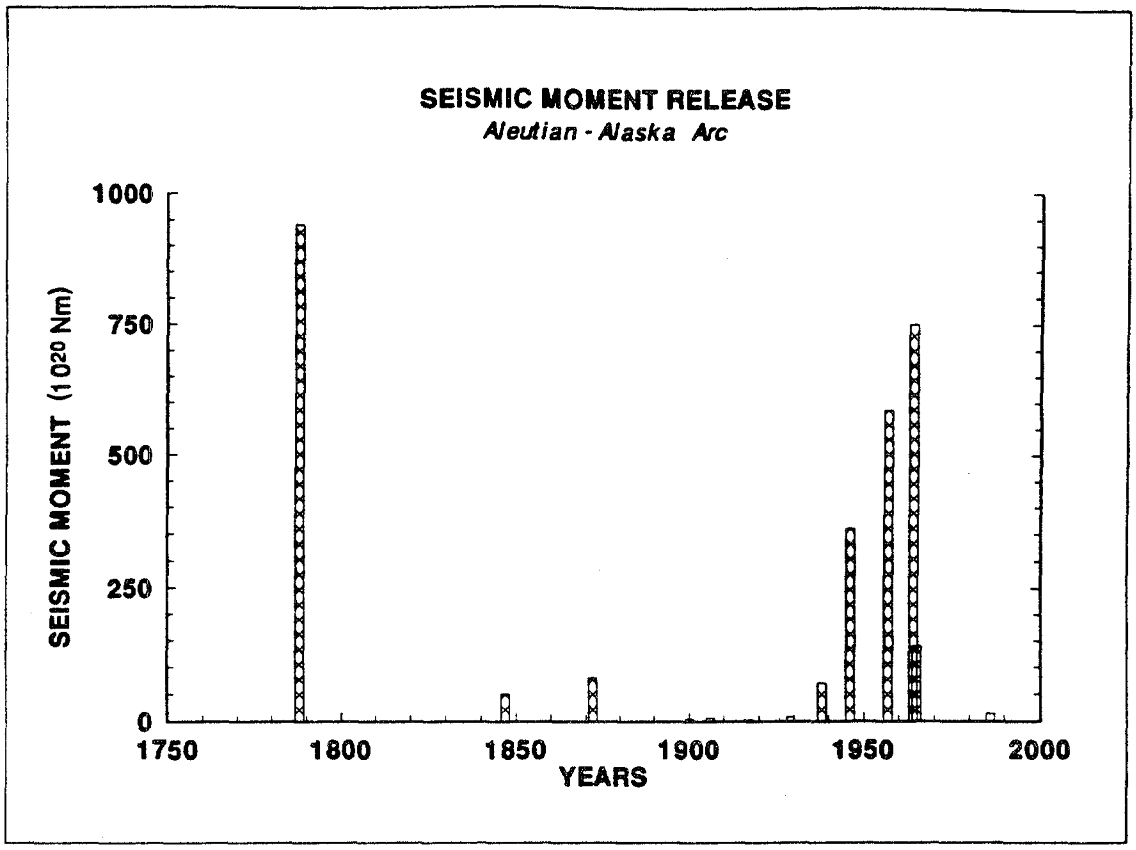


Figure 4. Seismic moment release with respect to time along the Arc. Note the concentration of moment release from 1938 to 1965.

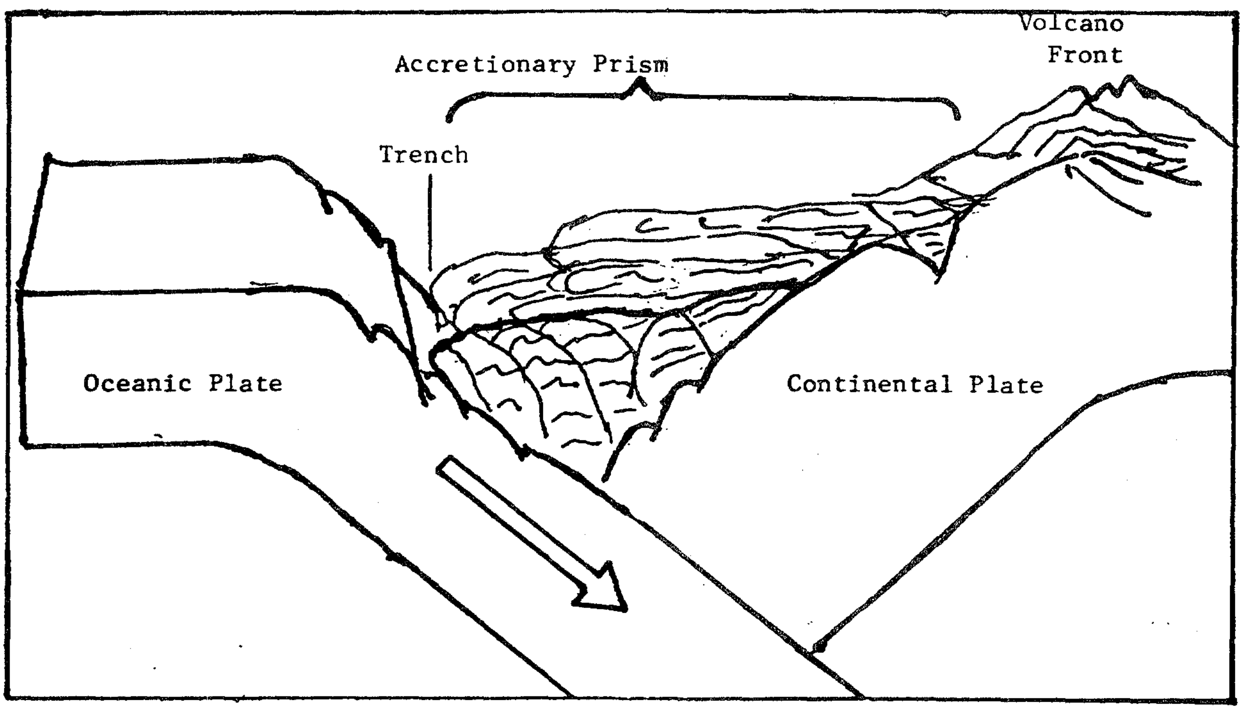
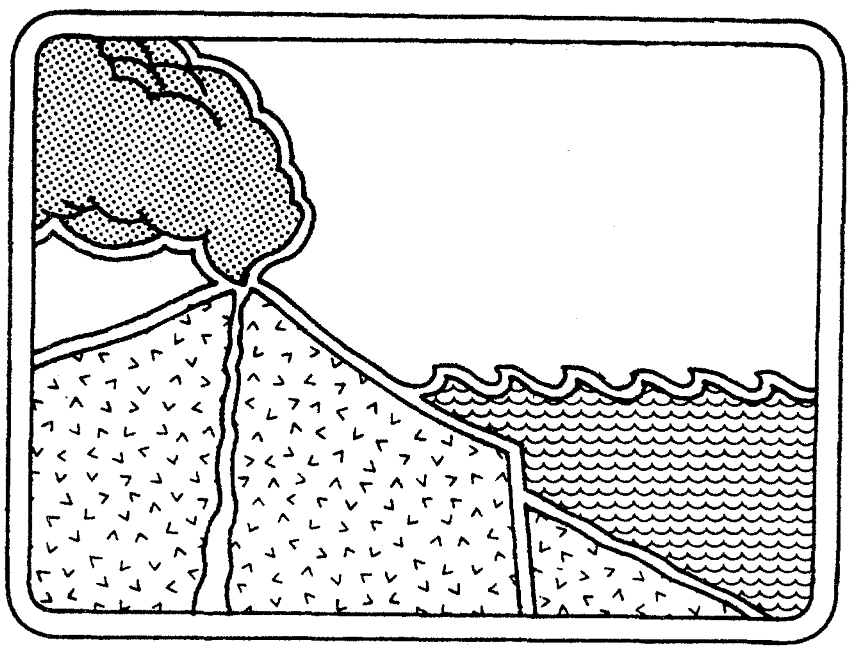


Figure 5. Cross section of a subduction zone showing relative positions of the continental plate, the down-going oceanic plate, accretionary prism and volcanic front. Original figure by W. Coulbourn and published by Uyeda (1989).



ON SOME EXCEPTIONAL SEISMIC (?) SEA - WAVES IN THE GREEK ARCHIPELAGO

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ABSTRACT

On 7 May 1991 an exceptional sea-wave was observed in the Laki Bay, Leros Island, SE Aegean Sea, Greece. The oscillatory nature of the wave and its period (12 mins), as indicated from tide-gauge records, are consistent with a seiche disturbance expected from the dimensions of the Laki Bay water volume. The lack of any tectonic movement implies that this seiche probably was a storm surge. On the basis of this evidence earlier Aegean sea-waves not correlated with tectonic movements may also be interpreted as storm surges. This includes the sea-waves observed on 23-25 April 1928, 23 February 1959 and 6 June 1961.

1. INTRODUCTION

The Greek Archipelago, that is the Aegean Sea and adjacent seas, is one of the most tsunamigenic regions of the non-Pacific Ocean area of the world. Many tsunamis have been reported in historical times (e.g Galanopoulos 1960, Ambraseys 1962). During the present century at least eighteen seismic sea-waves of intensity ranging between II and VI have been reported and/or recorded (Papadopoulos and Chalkis 1984).

Seismic activity is the most frequent generation cause of these tsunamis. Volcanic activity and submarine landslides are factors as well. However, in tsunami catalogues for this region there are cases of exceptional sea-waves of questionable or even unknown causes. Two such waves were observed in the Aegean Sea in 1991. This paper is devoted to these Aegean exceptional sea-waves. Information is given on the main features of the 1991 waves while existing information for earlier waves is summarised and critically reviewed. The waves studied seem not to be associated with tectonic displacements on the sea-floor. However, no conclusive result on the problems of their generation mechanism has been obtained, and proposals for further research are suggested.

2. THE PRE - 1991 WAVES

At least three cases of exceptional sea-waves during the present century have been listed in tsunami catalogues of the Greek area; the 1928, 1959 and 1961 ones. Table 1 and Figure 1 summarize existing information about the main parameters and the generation causes of these waves.

2.1 The 1928 wave

This event has been reported by several investigators and Ambraseys (1962) has compiled the relevant information (Table 1 and Fig. 1) Antonopoulos (1980) and Papadopoulos and Chalkis (1984) have reproduced this information in their tsunami lists. However, Ambraseys (1962) did not correlate this exceptional sea-wave with earthquake movement. On the contrary, Antonopoulos (1980) correlated the wave with a small earthquake which took place in Bulgaria some 200km inland from the nearest north Aegean coastline (Fig. 1). A sequence of strong earthquakes which occurred in the eastern part of the Gulf of Corinth (Fig 1) has been correlated with the 1928 sea-wave by Papadopoulos and Chalkis (1984).

The predominant feature of this wave is that the sea-water disturbance was spread almost everywhere along the north, west and south coasts of the Aegean Sea. It is not certain whether the wave has not been reported from the eastern part of the Aegean Sea because of lack of instrumentation and/or dissemination of information or because of a real absence of sea-water disturbance. The maximum wave intensity was observed along the north coastline of Crete in the South Aegean.

2.2 The 1959 Wave

This was a sea-wave similar to that of 1928. This wave has been tentatively reported by Ambraseys (1962) after personal communication with the Hydrographic Service of the Hellenic Navy, Athens. The disturbance was spread along north, west and southeast coasts of the Aegean Sea (Fig 1) with parameters as shown in Table 1. Antonopoulos (1980), Papadopoulos and Chalkis (1984) and Soloviev (1990) have reproduced this information. As for the generation mechanism Ambraseys (1962) noted that no earthquake shock was recorded around the time of the wave and he assumed that it was most probably a seiche.

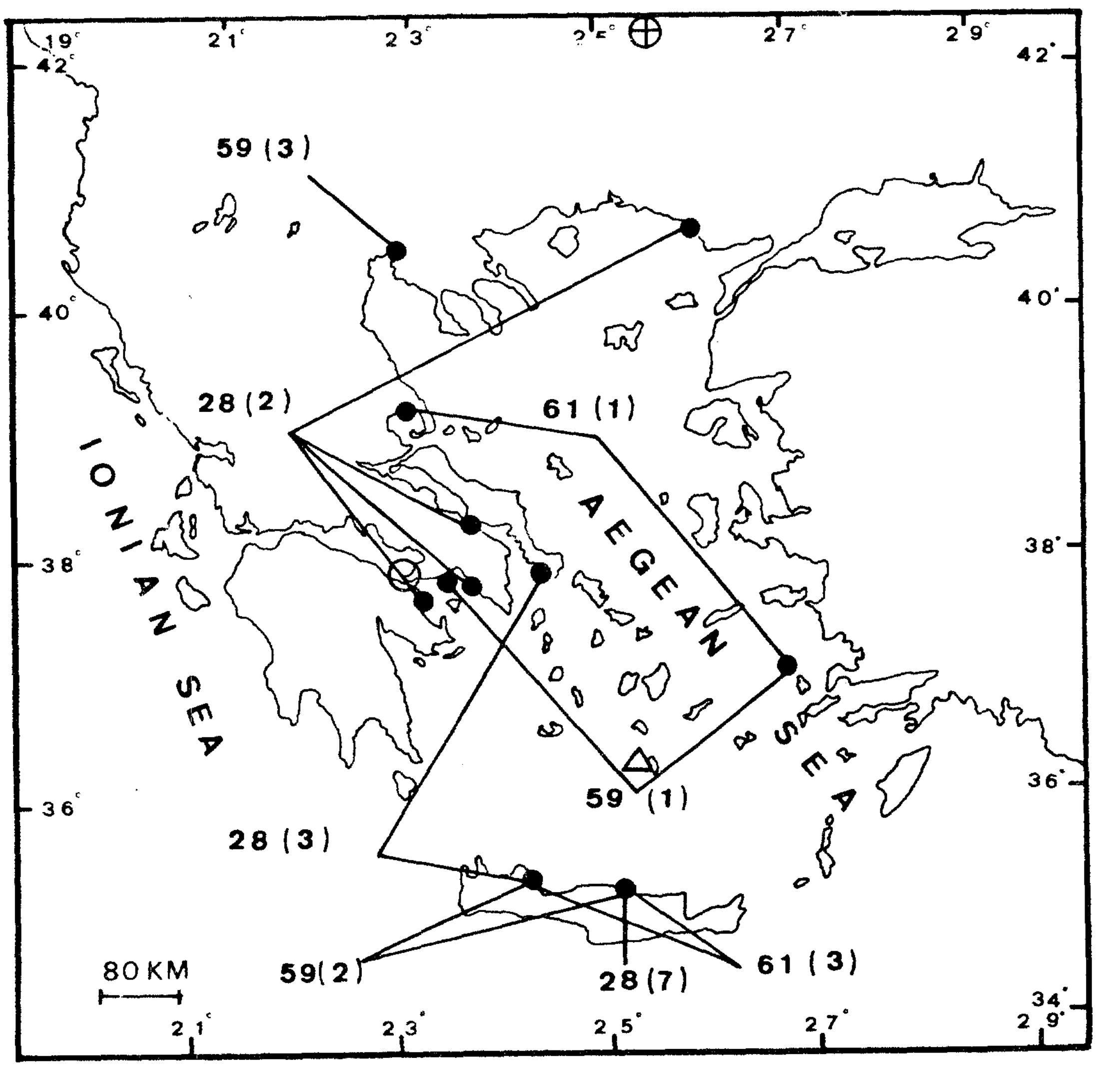

Fig. 1 Map showing the 1928, 1959 and 1961 sea-wave run-up in feet (in parenthesis) as listed in Table 1. Triangle, open circle and circle with a cross show the positions of the Thera volcano, the 1928 earthquake sequence in the eastern side of the Corinth Gulf, and the 1928 earthquake sequence in South Bulgaria respectively.

Antonopoulos (1980) adopted this assumption while Papadopoulos and Chalkis (1984) and Soloviev (1990) reported unknown causes of generation. It is not certain why the wave has not been reported from the east coast of the Aegean Sea. The maximum wave intensity was observed in the northwestern Aegean Sea.

2.3 The 1961 Wave

This is one more case of exceptional sea-wave not associated with earthquake activity. This wave was tentatively reported by Ambraseys (1962) after his personal communication with the Hydrographic Service of Athens. The disturbance was spread in the west, south and southeast coasts of the Aegean Sea (Fig 1) with parameters as shown in Table 1. Antonopoulos (1980) and Soloviev (1990) have reproduced this information. Ambraseys (1962) and Antonopoulos (1980) suggested that it was probably a strong seiche while Soloviev (1990) reported unknown generation causes. As in the cases of 1928 and 1959 it is not certain why the 1961 wave was not reported from the eastern Aegean Sea coast.

3. THE 1991 WAVES

3.1 The January 1991 case

According to press reports of 4 January 1991, a strong tidal wave was observed in Evdilos along the north coastline of the island of Ikaria, east Aegean Sea, Greece (Fig 2). From personal communication of the author with the Ikaria port authorities it is clear that the wave fundamental period was equal to about circa ten minutes, the sea water penetrating inland for about five minutes and withdrawing to the open sea for another five minutes. The sea water disturbance lasted for about three days ie between 2 and 4 January 1991. There is no information of similar disturbances in other coastal areas at the same time. Earthquake activity has not been observed in the vicinity of Ikaria Island. The largest earthquake which occurred in the Aegean Sea in the time interval between 1 and 4 January was $M_L = 3.8$, where M_L is the local magnitude (BSINOA 1991a).

3.2 The May 1991 case

This is the most recent and well-documented exceptional sea-wave in the Greek Archipelago. Some preliminary information has been given by Papadopoulos et al. (1992). Here the wave is studied by utilizing the available field and instrumental observations.

Local authorities on the island of Leros (Fig. 2), based on field observations, reported that a sudden rise of the mean sea water level of about 0.5m occurred in the Laki Bay, Leros, on 7 May 1991. The wave caused small fishing boats to move ashore.

Examination of tide-gauge records of the Hydrographic Service of Athens revealed that the wave was recorded in several ports of the Aegean Sea and in at least one port of the Ionian Sea. Their positions and names are shown in Figure 2. Analysis of the mareograms, particularly that of Laki Bay tide-gauge, showed that the wave consisted of a gradual rise of the mean sea water level followed by an amplification of high frequency oscillations. The sea water rise culminated to a maximum run-up of about 0.5m which is compatible with the field observations. The fundamental period was about T=12min with the wave superimposed on the normal semidiurnal tide. The disturbance lasted for about 13 hours with a gradually decreasing wave amplitude. Figure 3 is a modelized representation of the Laki mareogram.

No considerable earthquake activity occurred at the time of the sea wave. Between 1 and 7 May 1991 the largest shock recorded in the Aegean Sea was M_D=3.8, where M_D is duration magnitude (BSINOA, 1991b)

4. DISCUSSION

It is possible that the 7 May 1991 wave in Laki Bay was a seiche (ie a free oscillation of the water whose period depends on the dimensions of the water volume). If the equations for a standing wave are solved with appropriate boundary conditions then an approximation of the wave fundamental period, T, is (eg Nakano and Fujimoto 1983)

$$T = 4L\,(gh)^{-1/2} \qquad (1)$$

where L is the length of the water surface, h is the uniform depth of the water layer and g is the acceleration due to gravity. For T=12min. and L=3250m (see Fig 4) we get h=33m which is very consistent with the sea water depth distribution in Laki Bay (Fig 4).

A seiche can be initiated by a tectonic movement, by a temporary wind stress acting on the water, or by atmospheric pressure differences across the water surface. As there is no evidence that a tectonic movement triggered the 7 May 1991 sea-wave one may suggest that the wave originated from a disturbance of sea level arising from atmospheric causes, (ie a storm surge). Sea water disturbances due to the occurrence of submarine landslides is not a reasonable explanation because only local sea-water effects may have been triggered by such an event. In addition, it does not explain the oscillatory nature of the disturbance.

The January 1991, June 1961, and February 1959 sea waves were not associated with any apparent tectonic movement. For reasons explained above, submarine sediment sliding due to gravity does not offer a reasonable explanation given that these sea-waves are similar in their general features with that of May 1991 with the exception of the probable local character of the January 1991 tidal wave. In these cases the hypothesis that storm surges have been the generation causes of the waves appears to be applicable. As for the 23-25 April 1928 sea-wave the suggestion of Antonopoulos (1980) that it was associated with an earthquake of small size which occurred in South Bulgarian appears unreasonable. However, seismic activity was very high in Bulgaria and Greece during the days that preceded the sea water disturbance. More precisely according to Comninakis and Papazachos (1986) on 14 and 18 of April two earthquakes of $M_s=6.8$ and $M_s=7.0$, respectively occurred in South Bulgaria while a large number of smaller earthquakes ($M_s=5.0$-5.5) took place until 28 April (Fig 1). In the Gulf of Corinth, Central Greece, another sequence of strong earthquakes occurred between 22 and 25 April 1928 (Table 1, Fig. 1). In addition, it is of interest that activity in the Thera volcano, South Aegean (Fig. 1), had entered an eruptive phase between 4 and 14 March (Ktenas and Kokkoros 1928). No causal connection between tectonic displacements associated with the earthquake of volcanic activity and the sea water disturbance can be suggested. A reasonable hypothesis is that earth shaking triggered creeping of unstable submarine sediments and that this process accelerated some days later leading to sediment collapse. Such an interpretation is consistent with the suggestion of Zore-Armanda (1988) that long-waves may have been created by earth shaking associated with earthquakes occurring not only in the Adriatic but also in the Aegean Sea and adjacent areas. It is questionable, however, if a regional sea wave disturbance, such as that of 23-25 April 1928, could be triggered by submarine sediments sliding. An alternative, more reasonable hypothesis is that the sea-wave was caused by atmospheric causes similar to those suggested for other waves studied here. The wave source, no matter what the generation mechanism is likely to have been located somewhere in the South Aegean as the wave intensity distribution indicates. (Fig. 1).

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TABLE 1

Information of exceptional sea-waves reported in the Greek Archipelago during the present century. In the Table of sea-waves H is the wave run-up while ii, ii+, and iii+ indicate intensities in the Sieberg six-grade scale. In the Table of earthquakes $\phi^o{}_N$, $\lambda^o{}_E$, h, M, I and n indicate geographic latitude, geographic longitude, focal depth, surface-wave magnitude, maximum intensity in MM, and normal focal depth, respectively.

SEA WAVES

No	Date	Region	Authors
1	1928 Apr 23-25	Piraeus Chalkis Nafplio Alexandroupolis (H=2') (ii) Crete Chania Karystos (H=3') (ii+) Crete (H=7') (iii+)	AM, AN, PC
2	1959 Feb 23	Crete (H=2') Leros (H=1') Salamina (H=1') Thessaloniki (H=3')	AM, AN, PC, S
3	1961 June 6	Crete (H=3') Volos (H=1') Leros (H=1')	AM, AN, S

EARTHQUAKE OR OTHER GENERATION CAUSES

No	Date	$\phi^o{}_N$	$\lambda^o{}_E$	h	M	I	Authors
1	??.4.1928	42.4	25.7	n	4.6	VI	AN
1	22.4.1928	37.9	23.0	n	5.2	VI	PC
1	22.4.1928	37.9	23.0	n	6.3	IX	PC
1	25.4.1928	38.0	23.0	n	5.2	V	PC

No	Description	Authors
2	No earthquake shock was recorded. Most probably a seiche.	AM, AN
2	Unknown generation causes	PC, S
3	Indicated that a strong seiche has taken place	AM, AN
3	Observed that the generation cause was unknown	S

AUTHORS KEY:

- AM - AMBRASEYS (1962)
- AN - ANTONOPOULOS (1980)
- PC - PAPADOPOULOS AND CHALKIS (1984)
- S - SOLOVIEV (1990)

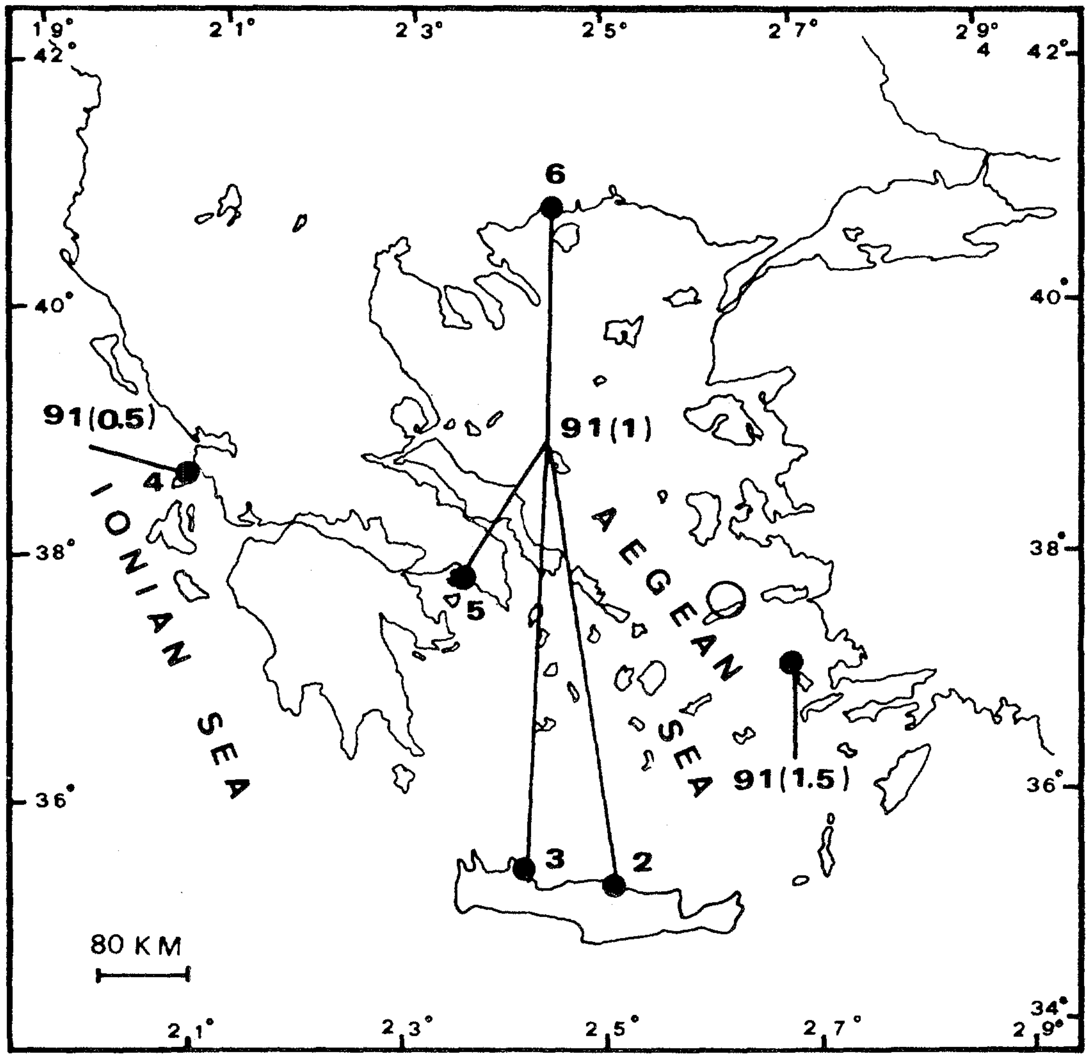

Fig. 2 Map showing the 7 May 1991 wave run-up in feet (in parenthesis) as recorded by tide-gauges in Laki (1), Iraklio (2), Chania (3), Lefkada (4), Piraeus (5) and Kavala (6). Open circle represents the position where 4 January 1991 sea-wave was observed in Evdilos, Ikaria Island.

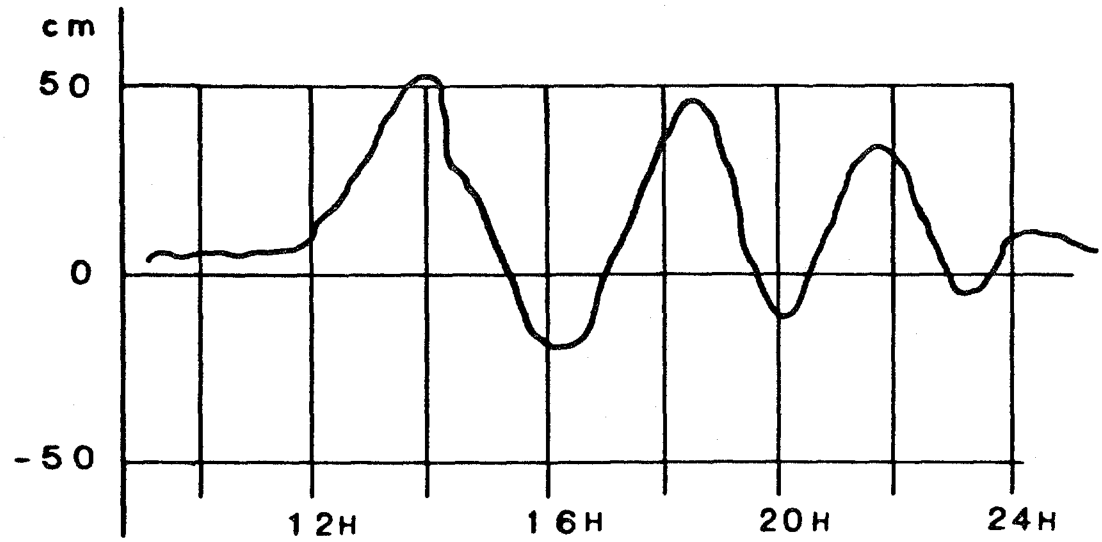

Fig. 3 Modelled record of the 7 May 1991 seiche at Laki Bay, Leros. Maximum amplitude has been adopted by assuming that the normal semidiurinal tidal recorder fully responds to the seiche frequency.

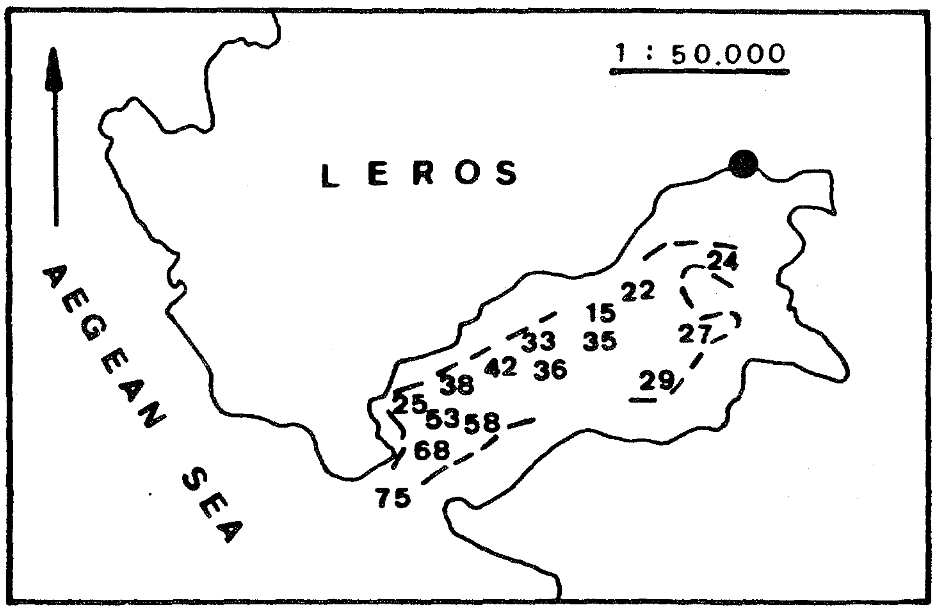

Fig. 4 The Laki Bay, Leros, and its port position (solid circle). Dashed line is the 20-m isobath. Numbers show sea water depths (in m) larger that 20m (after Topographic Map of Greece, Kalymnos Sheet, Hellenic Army Geographic Service, 1972).

TSUNAMI SEDIMENTATION SEQUENCES IN THE SCILLY ISLES, SOUTH-WEST ENGLAND

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ABSTRACT

Sediments deposited by both storm surges and tsunamis are well documented. Discrimination between these two types of deposit is frequently complicated by the existence of both processes in the same geographical region. In an earlier paper, Foster et al. (1991) analysed historical records and traced both storm surge and tsunami activity to the early 18th century in south-west England. Investigation of sedimentary deposits in five coastal lagoons on the Scilly Isles provided evidence for a complex Holocene lithostratigraphic sequence with peats interspersed with sand layers of varying particle size and thickness. It was earlier suggested that a 30 to 50cm thick sand layer in Big Pool, St Agnes may have been deposited by tsunami waves generated by the Lisbon earthquake of November 1st, 1755. The present paper presents lithostratigraphical evidence for the area immediately surrounding Big Pool based on site surveys conducted in July 1992. The results identify a thickening of the deposit on the seaward side and over a metre of sand is observed in parts of the sedimentary basin. This suggests a high energy depositional environment of a type commensurate with patterns of tsunami deposition observed elsewhere in the world.

INTRODUCTION

It has long been recognised that tsunami waves deposit marine sediments on land, frequently deriving material from the nearshore zone as the wave height increases in shallow water. Some of these deposited sediments comprise boulder size materials (Moore & Moore, 1988) but the majority consist of sand layers, often interbedded with peats and/or coastal silts and clays that range in thickness from a few cm to over 2m (eg Atwater, 1989; Dawson et al, 1989; Minoura and Nakaya, 1991; Plafker & Mayo, 1965; Wright & Mella; 1963).

Although northwestern Europe is generally considered to be geologically stable a major area of instability is associated with the intraoceanic subduction zone located immediately south of the Gorringe Bank (Andrade, This Volume). Indeed, Murty (1977) lists 29 tsunamis for which written records exist in the eastern Atlantic south of the Bay of Biscay between 1531 and 1960 and 22 tsunamis in the European and Mediterranean seas between 1902 and 1957. One of the highest magnitude earthquakes in history with an estimated magnitude of 8.5 was centred immediately south of the Gorringe Bank and caused extensive shock wave damage in Portugal. With the additional damage caused by the subsequent tsunami, an estimated 60 000 people lost their lives in Portugal alone. Elsewhere, large waves struck Gibraltar, Morocco, Madeira and the south-west coasts of England and Ireland. A detailed review of the historical literature relating to the nature and extent of shock waves, seiches and the tsunami wave itself is reported by Reid (1919); much of the contemporary evidence being reported in letters written from various parts of the world and published in Volume 49 of The Philosophical Transactions of The Royal Society of London, 1755.

BACKGROUND TO PRESENT RESEARCH

In a recent investigation of coastal flooding in southwest England, Foster et al. (1991) identified two distinctive types of marine inundation documented for the region. These included six major storm surges reported between 1703 and 1989 and written evidence for the arrival and magnitude of the Lisbon tsunami (Borlase, 1755; 1758; Huxham, 1755). It was demonstrated that the Lisbon tsunami arrived about one hour after high tide and Borlase (1758) also reported maximum amplitudes of the waves in excess of 10ft for Mounts Bay, West Cornwall. These marine inundations were superimposed upon a region characterised by rapid relative sea level rise reconstructed from archaelogical and beach level evidence (Hawkins, 1971; Thomas, 1985).

Coring of five shallow coastal lagoons on the Isles of Scilly, located some 40km off the south-west coast of Cornwall (Figure 1), revealed a complex lithostratigraphy with both wind blown sand and marine deposits present in the sedimentary sequences. Of particular significance was the sedimentary record of Big Pool, St Agnes. St Agnes is the most southerly inhabited island of the group and is located on the margin of a Carbo-Permian age granitic intrusion less than 0.5 km from the 50 m marine isobath. This island is one of the most exposed of the island group. The lake is located in a low lying marshy area (c 2.5 m above local datum) on a northern promontory of the island in close proximity to three coastal bays, Periglis, Porth Coose and Porth Killier (Figure 2). In consequence, this lake is exposed to marine incursion from the west, north-west and north-east. Today, the lake is protected by artificially raised coastal embankments which were originally constructed in the 19th century (Tonkin and Tonkin, 1887). These embankments rise to a height of

5.25 m above local ordnance datum (ALOD). The upper limit of the high water mark of ordinary spring tides (HWMOST), estimated by levelling the upper limit of the seaweed Fucus vesiculosus (Bladder Wrack) in the three embayments, is 2.41m ALOD. A second indirect measure of the altitude reached by spring tides is that of the lower limit of land based vegetation which was measured here at 3.48 m ALOD.

Six cores taken in and around the Pool (Figure 3) were described in detail by Foster et al. (1991). These cores contain organic gyttja, peats and, in one core analysed in detail, 23 separate sand layers ranging in thickness from a few mm to 50 cm (the main sand unit). A peat layer is identified above (Peat I) and below (Peat II) the main sand unit. A coarse sand/fine gravel layer forms the basal stratigraphic unit below Peat II. The thin sand layers increase in frequency and thickness upcore from the base of Peat II to the main sand unit. Particle size analysis of the sand layers was performed for comparison with contemporary marine and subaerial deposits taken in close proximity (Figure 3). Results from these analyses suggested three dominant groups of sediments. Fine sands in the sedimentary sequence are of similar particle size distribution to Holocene sand dunes exposed in nearby coastal sections, whereas medium sands are similar in particle size distribution to contemporary beach sediments. The coarsest sediments are contained in the thick sand layer which are much coarser than any of the contemporary sediments collected in the area. Detailed particle size analysis of 1 cm slices of this sedimentary unit revealed median particle sizes ranging from 400 to 800 μm which appear to represent a sequence of several fining upward sedimentary units. Although rather variable below 40cm depth, three distinctive fining upward sequences are apparent above 40cm depth in the core. The general trend is for particle size to increase upcore from the underlying basal peat for the first 10 cm of deposition and subsequently decrease towards the overlying peat (Figure 4).

Radiocarbon dating was performed on the basal peat sequence immediately below the main sand layer. At the junction with the basal coarse sand/fine gravel deposit, the calibrated radiocarbon ages for two standard errors range from 1019 to 1216 AD (100% of the relative area under the probability distribution). At the juntion with the overlying sand layer, again for two standard errors, the calibrated date was was 1474-1672 (91% of the relative area under the probability distribution) and 1751-1795 (7% of the relative area under the probability distribution). This provides a minimum date only for the upper peat, since peat inclusions in the sand layer would indicate some erosion of the original underlying surface and the onset of the Suess effect (an apparent increase in radiocarbon age upcore caused by burning of fossil fuels) from the mid 18th century onwards may have had some impact on radiocarbon assay (Lowe and Walker, 1984). A mid 18th century date for the upper layer of the peat surface from the radiocarbon dating is therefore not unreasonable in the context of the arguments provided above.

In summary, the sedimentary sequence contains a major peat unit overlying and underlying a sand layer of up to 50 cm thickness. The Peat II unit contains approximately 20 thin sand layers which increase in thickness and frequency up-core. The Peat I unit contains only 2-3 thin sand layers immediately above the junction with the main sand unit.

THE 1992 SURVEY

In July 1992, a further field excursion was undertaken in order to provide detailed lithostratigraphic information on the distribution of the coarse sand layer and sub-layer basin configuration. In total, 44 Gouge Auger bore holes were sunk along five major transect lines across the pool and its surrounding marsh (Labelled T1 to T5 on Figure 3). Due to compaction and the presence of large boulders, no coring was undertaken through the artificially raised coastal barriers on the margins of Periglis and Porth Coose. Coring through the contemporary beach in Porth Coose revealed a thin beach deposit (30cm) underlain by a coarse angular slopewash, possibly of periglacial origin, which could not be penetrated with manual coring equipment.

Transects 1, 3 and 5 trend west to east and south-west to north-east. Figure 5A illustrates the stratigraphy relative to local ordnance datum. Transect 1 flanks Big Pool, and comprises a relatively simple stratigraphy with a peat layer overlying (Peat I) and underlying (Peat II) a wedge of sand which at its thickest point in core 2 is almost 1 m thick. The sand thins rapidly from core 7 towards the eastern side of the basin and is only 2 cm thick in core 13. Transect 3 exhibits similar characteristics to transect 1 in that the thickest sand layer is recorded on the Porth Coose/Periglis side of the basin, wedging towards Porth Killier where the underlying Peat I is not present. The main sand unit here is underlain by angular slopewash deposits, possibly of periglacial origin. The lake sediments in Big Pool, as determined in the 1988 survey by Foster et al. (1991), are generally less than 15 cm in thickness and are underlain by peat before the sand layer is encountered. This information is recorded on transect 3 although no cores were retrieved from Big Pool during the 1992 survey. The thickest part of the main sand unit is recorded in Transect 5 at coring sites 36 and 37 where 1.3 m of sand were penetrated before encountering the underlying Peat II. Again, at the north-eastern end of the basin, the sand layer rests unconformably upon an angular slope deposit and the Peat II layer is absent from the sequence.

Transects 2 and 4, trending north-south contain a similar stratigraphic record to the other three transects, with the sand layer thickening towards a maximum of c 1.3 m towards the Porth Coose and Periglis embayments (Figure 5B). However, in cores 15 to 18 (Transect 2) a fine wind blown sand and/or slopewash deposit was penetrated to reveal an inorganic laminated fine clay and silt which appears to be a lake sediment. This sediment extended to 0.85 m ALOD.

A further core taken from Little Pool (Figure 3) revealed a simple sedimentary sequence with 90 cm of peat overlying a basal coarse sand/fine gravel unit. The peat contains no identifiable sand units in the stratigraphic sequence.

DISCUSSION AND INTERPRETATION

The sedimentary basin of Big Pool contains a variety of stratigraphic units ranging from rhythymically laminated lake sediments to peat. Overlying the varved lake sediment in the east and north of the basin is a coarse angular slopewash deposit and a fine sand. The slopewash is probably of periglacial origin related to the last major ice advance in northern Europe which left extensive periglacial deposits now found in coastal sections in the Scilly Isles (Scourse, 1986). The fine sand is probably of wind blown origin and may be part of extensive Holocene coastal dune sequences found on Scilly. The exact date of their deposition is as yet unknown. Dimbleby (1976) records extensive sand deposition from an analysis of a buried soil on the island of St. Mary's (Figure 1). The lower layers of the soil contain pollen of an oak-hazel woodland but the profile appears to have been truncated, perhaps as a result of early clearance for agriculture. Immediately below the dune sand the soil contains pollen indicatve of a grass heath. Later work by Scaife (1984) suggests that the change from arboreal to herbaceous flora took place at around the period of the Bronze age/iron age transition although two of the radiocarbon dates in this study are inverted which makes the exact determination of the date of this transition problematic.

The earlier study by Foster et al. (1991) suggested that accumulation of the Peat II layer began around a thousand years ago. It was suggested that this accumulation resulted from impeded drainage in a low lying marsh, perhaps as a result of rapid relative sea level rise constructing beach ridges; a pattern repeated in other coastal areas in south-west England (cf Borlase, 1758 for Loe Pool, Cornwall and Morey, 1976 for Slapton Ley, Devon). At this time, it was suggested that sea levels may have been as much as 2.4 m below that of the present day. With a subsequent rise in sea level, and an increase in storminess during the Little Ice Age (Lamb, 1984), it seems reasonable to suggest that the increase in frequency and thickness of the sand layers in Peat II is directly related to more freqent marine incursions up to the 18th century, although some of the finer sand layers may well have been wind blown. The decrease in the frequency of thin sand layers in the upper Peat I sequence was ascribed to a reduction in the frequency of Atlantic storms at the end of the Little Ice Age and the artificial raising of all coastal barriers around Big Pool to reduce inundation.

The presence of over 1 m of sand between the Peat I and Peat II units as indicated by the 1992 survey is undoubtedly associated with the most significant marine depositional event in the stratigraphic record of Big Pool and its surrounding marsh. Indeed, the particle size characteristics of the sand layer are similar to those described in north-west Scotland by Dawson et al. (1991) and appear to fit a recent conceptual model of tsunami-generated sedimentation described by Dawson et al. (1991). The thinning of this main sand unit to the north and east of the basin coupled with the absence of sand from the lithostratigraphic record of Little Pool would suggest that the main direction of penetration to the pool was from the west and north-west; through the Periglis and Porth Coose embayments. The greatest direct exposure to Atlantic waves, however, is more likely to come from the north-west since some protection is afforded to the Periglis embayment by the island of Annet and the granite reefs of the Western Isles (Figure 1). Furthermore, the current offshore zone from Periglis is dominated by a rocky rather than sandy substrate whereas an offshore sand bar (as yet unsampled) was observed some 200-300 m north-west of the Porth Coose beach.

The exact date and mode of deposition of this sand layer is indeterminate. Radicarbon dating does suggests that a mid 18th century date is not unreasonable, but two major storm surges of 1703 and 1744 are also encompassed by the range of dates provided. However, the weight of evidence presented here and by Foster et al. (1991) would make a tsunami origin for this layer most probable. The layer is much thicker than that normally associated with storm surges recorded in historical and contemporary inundations and more closely mirrors descriptions of tsunami sediments reviewed in papers referenced in the introduction to this paper. The particle size analysis of the main sand unit suggests a high magnitude low frequency depositional sequence, with a number of large waves spread over a period of several hours in order to produce the fining upward sequences observed. This sequence closely mirrors the descriptions provided by Borlase (1755); "The first and second refluxes were not so violent as the third and fourth, at which time the sea was as rapid as that of a mill-stream descending to an undershot wheel, and the rebounds of the sea continued in their full fury for fully two hours....alternatively rising and falling, each retreat and advance nearly of the space of ten minutes, till five and a half hours after it began."

If such a major sedimentary deposit is indeed associated with the Lisbon tsunami, other sedimentary units of similar type in the coastal zones of south-west England and elsewhere will undoubtedly come to light in the future and provide evidence of the exact mode of origin of this high energy stratigraphic unit.

The Holocene lithostratigraphy indicates that the tsunami was not only significant in relation to depositional processes but has also had an impact on the geomorphology of the area surrounding Big Pool. The evidence suggests that the present location of Big Pool itself is related to a depression left in the post-depositional period, eventually forming an area of shallow open water.

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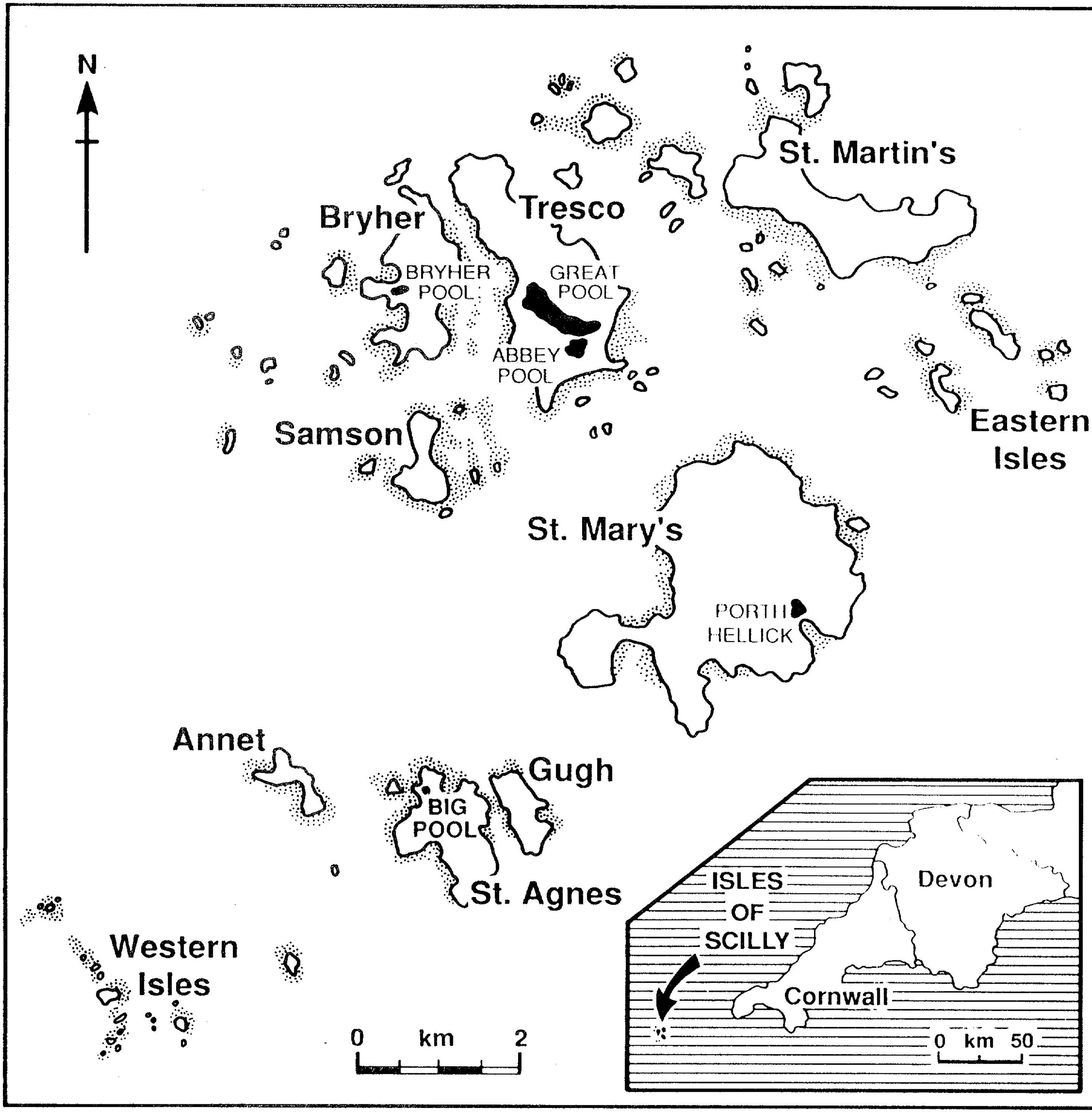


Figure 1. The Isles of Scilly. Location of Islands in relation to south-west England (inset) and location of the five shallow coastal lagoons on the Islands of Bryher, Tresco, St. Mary's and St. Agnes.

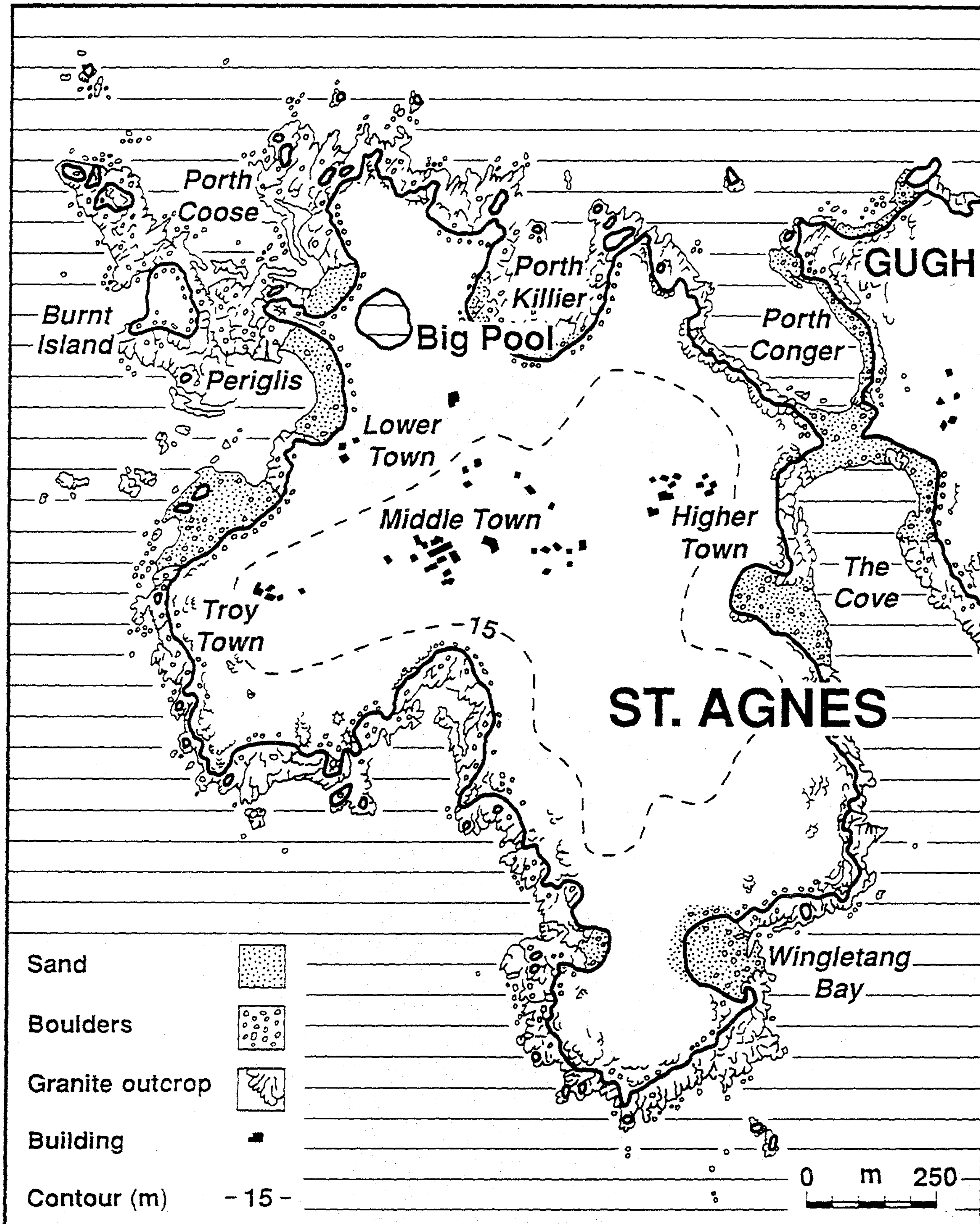


Figure 2. Big Pool St. Agnes in relation to the main island mass and three embayments of Periglis, Porth Coose and Porth Killier.

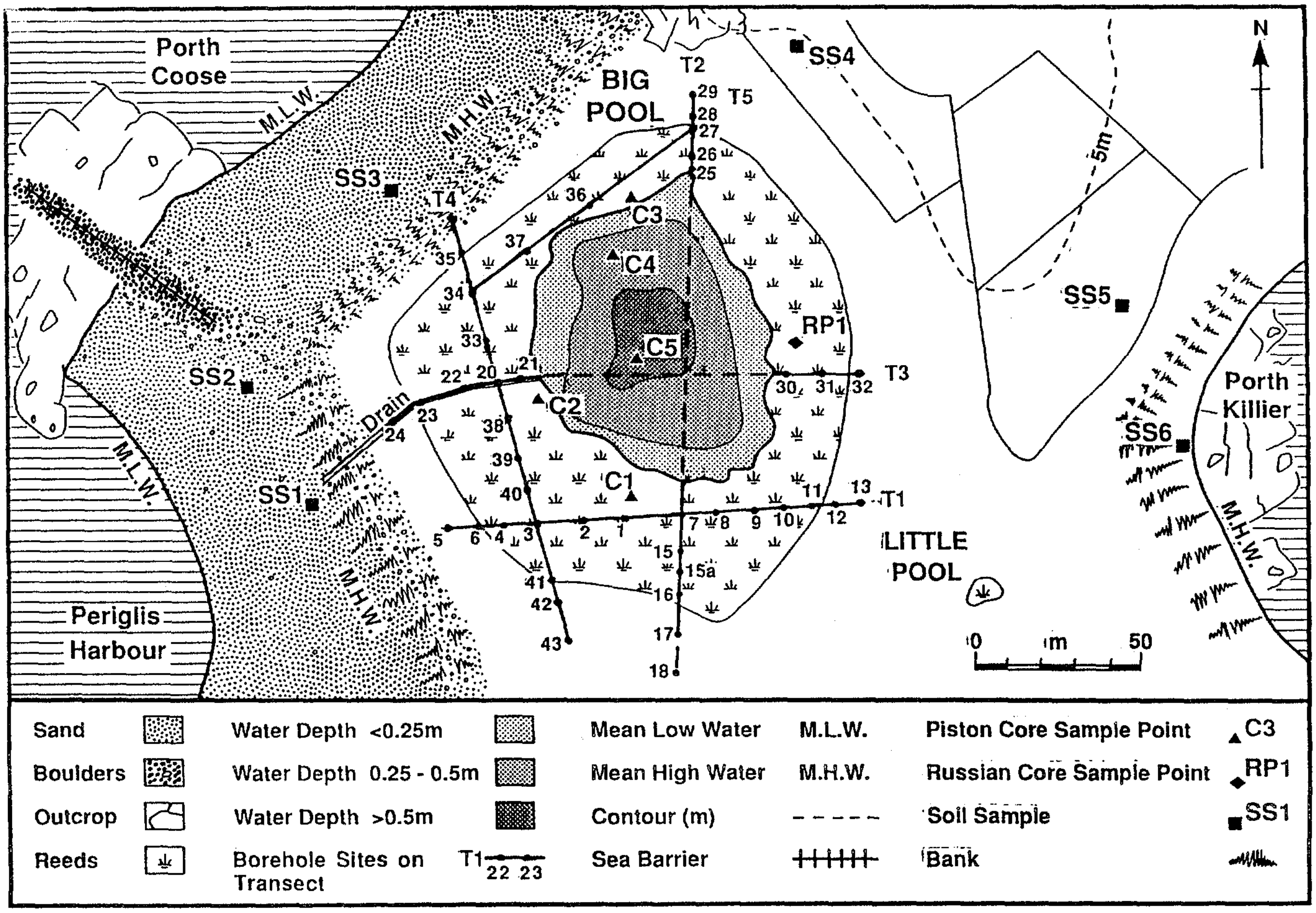


Figure 3. Large scale map showing the position of the 1988 and 1992 coring locations. Coring on transects T1 to T5 relate to the 1992 survey. Other samples were collected during the 1988 survey.

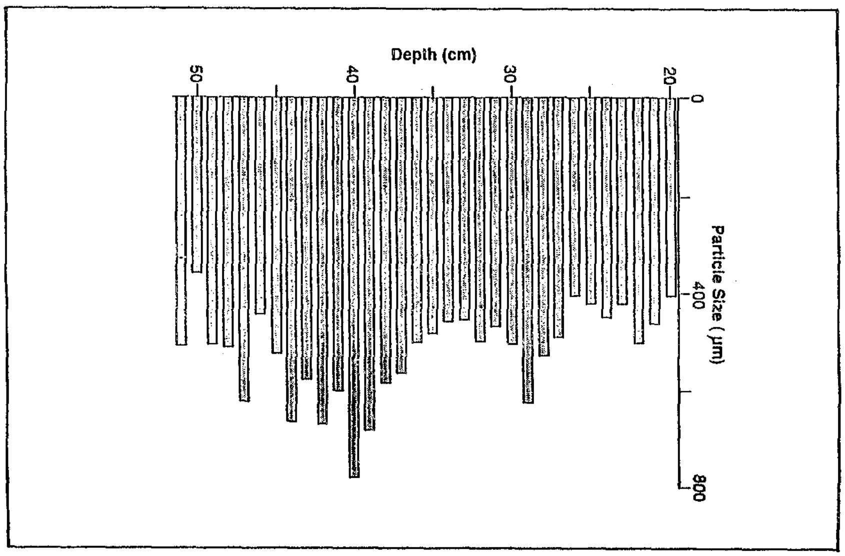
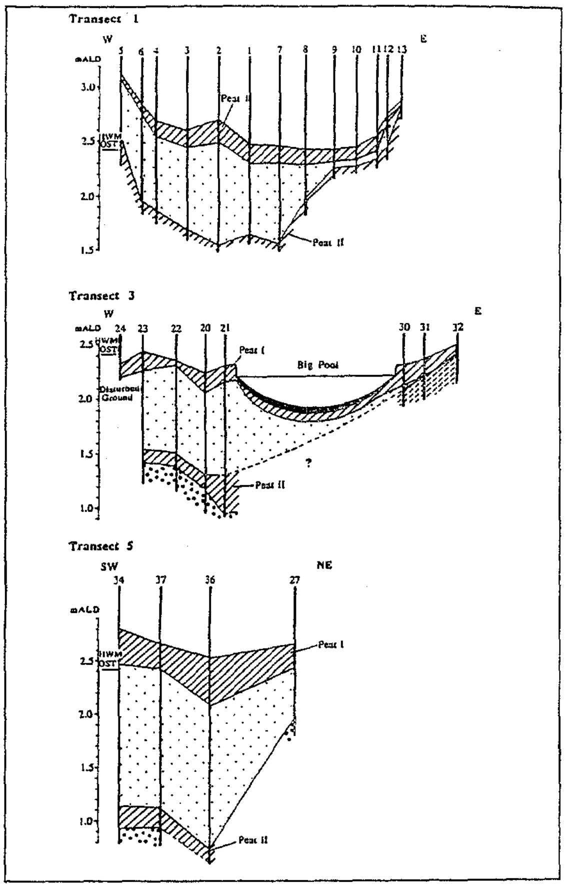


Figure 4. Downcore variations in modal particle size of the thick sand layer (Core C1; Figure 3). The Peat I and Peat II layers immediately overlie and underlie the sand layer.



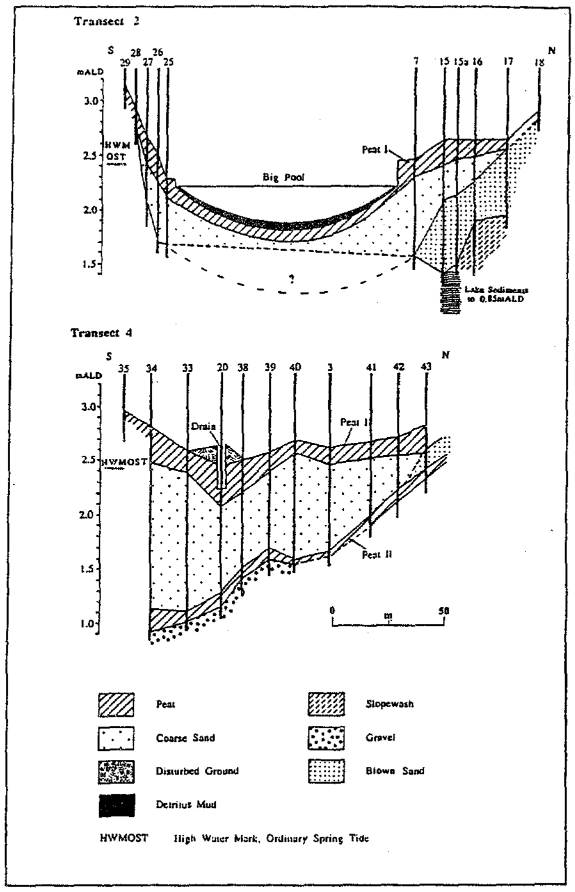
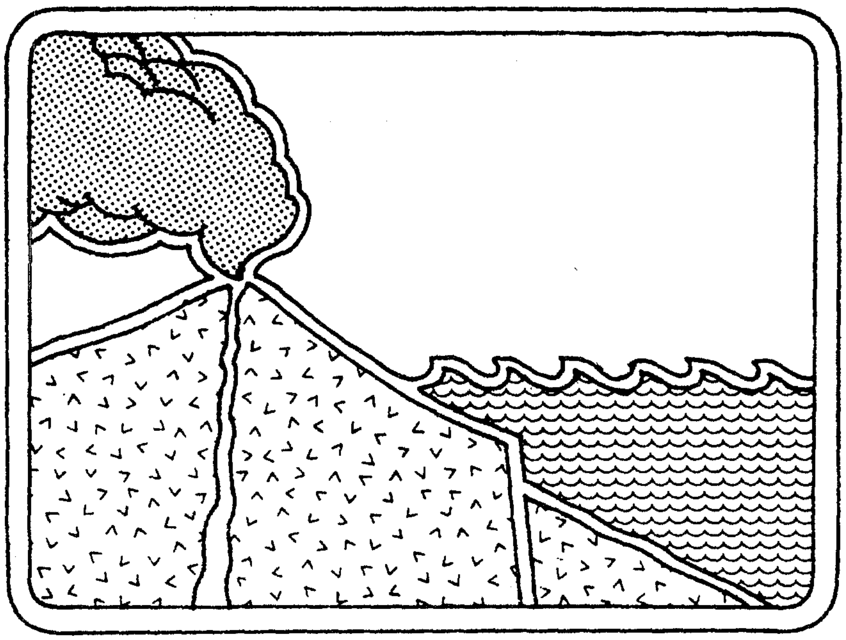


Figure 5. Stratigraphic sections drawn from the 1992 field survey of the five transects given on Figure 3. Figure 5A; transects 1, 3 and 5. Figure 5B; transects 2 and 4. Numbers and vertical lines relate to core position numbers given in Figure 3.



ANALYSIS OF OCEAN LEVEL OSCILLATIONS IN MALOKURIL'SKAYA BAY CAUSED BY TSUNAMI ON THE 16 FEBRUARY 1991

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ABSTRACT

On the 16th February 1991 at 01h 23m 43.3s GMT an earthquake occurred SE of Shikotan (Kuril Island) of magnitude Ms = 5.9. Analysis of seismic data has shown that the aftershocks of this quake were located along faults transverse to the island arc. About 1½ hours later in Malokuril'skaya Bay (Shikotan), intensification of seiche oscillations of 18 minutes period was observed. Synchronous measurements of ocean-bottom hydrostatic pressure (ocean level) inside and outside the bay were made. An attempt was made to measure the response of the bay as a resonator to the tsunami. The period of the first harmonics of seiche oscillations (Helmholz mode) in the bay and Q-factor of the oscillating system were estimated. The empirical frequency-amplitude response function relating the input signal outside the bay to the ocean level oscillations record inside the bay was calculated. The maximum value was 4 for an 18 min. period. A new method of estimation of the period and exponential decrements of oscillators based on an autoregression model was developed.

1.1 INTRODUCTION

Despite the significant progress achieved in the development of sea measuring techniques the registration of open ocean level oscillations is performed rather rarely in modern oceanographic practice. Technical complications and the high cost of facilities are limiting the wide application of deep tide gauges.

Such experiments are usually associated with the development of marine measuring systems for the detection of tsunami waves in seismically active oceanic zones.

Measurements of ocean level oscillations are based on the applications of high precision ocean-bottom hydrostatic pressure sensors installed at the ocean bottom. In the range of long wave frequencies, of tidal and tsunami type, fluctuations of hydrostatic pressure at the bottom are caused mainly by displacement of the free ocean surface.

In the USSR measurements of open ocean levels are carried out in order to create an effective system of detection and operative forecasting of tsunamis. On the basis of the hydrophysical observatory on the island Shikotan (the Malye Kuril Islands), long-term ocean bottom systems to record hydrostatic pressure have been installed. By means of such a system tsunami waves in the open ocean were measured in 1980 for the first time in the USSR. (Dychan, Jaque and Kulikov et al, 1981).

Measurements of open ocean level oscillations also provide data on the propagation and transformation of tsunami waves on the shelf, their interaction with the ocean bottom relief, and their reflection from the shore etc. It is known that local manifestations of a tsunami may be significantly different close to shore. Local effects are more typical for straits, bays and estuaries, where the resonance properties of partly enclosed water volumes as a result of the impact of a running wave may be displayed.

In the present work two synchronous records of hydrostatic pressure obtained by sea floor sensors inside and outside of Malokuril'skaya Bay after the earthquake of February 16th 1991 are investigated.

2.1 RECORDS OF THE TSUNAMI OF FEBRUARY 16th 1991, IN MALOKURIL'SKAYA BAY

The Kuril-Kamchatka zone is one of the main tsunamigenic zones in the Pacific Ocean ring (Soloviev and Go, 1974,1977). In general, sources of strong earthquakes and of tsunamis extend along the island arc which is divided into blocks by general transverse faults. (Lobkovskiy and Baranov 1983). Earthquakes occur along these faults also, but are of less energy (but sufficient to generate weak tsunami waves). Such seismotectonic characteristics are typical of other island arcs in the Pacific Ocean (Soloviev 1965).

The occurrence of tsunamis and strong earthquakes in the Kuril-Kamchatka zone is irregular. In general they cluster in time with alternate high and low periods of earthquake and tsunami recurrence (Ivatshencko 1972). An active period has been observed between 1945-1965. During the last decade there were no strong earthquakes and tsunamis in this region.

The earthquake of 16th February 1991 was characterised by a hypocentre depth of circa 40 km. The location of the epicenter was between 48'2" - 48'28" N and 154'5" - 154'26"E.

The data on epicenters and other parameters of the aftershocks in first hours after the

Shikotan earthquake are given in Table 1. From Figure 1 it is evident that the earthquake was caused by displacement along of one of the faults lying orthogonal to the Kuril Islands arc. This conclusion is proved also by the fact that the strongest aftershock on the 20th February occurred in the Okhosk Sea (latitude 48.9° N, longitude 153.4° E, h = 33km) along the continuation of the line of the epicenters location immediately after the earthquake. The fault reactivated during the earthquake coincides with one of 14 transverse faults in Kuril-Kamchatka seismoactive zone determined by Baranov (personal communication).

TABLE 1

Data on aftershocks immediately after the Shikotan earthquake of February 16th 1991.

Date	Origin Time GMT		Epicenter		Energy Class
	h	m	Lat	Long	
16	01	41	48.2°	154.4°	9.5
16	01	47	48.2°	154.4°	10.0
16	01	59	48.1°	154.9°	10.5
16	02	02	48.1°	154.7°	9.5
16	02	10	48.0°	154.9°	9.0
16	05	07	48.0°	154.8°	9.0
16	15	03	48.0°	154.8°	9.0

(After N Poplavskaya, Yuzhno-Sakhalinsk, pers.comm.)

It is known that the maximum tsunami energy radiates orthogonally to the long axis of the wave source. In the case considered here this means that the main energy of the tsunami should have been propagated along the Kuril continental slope. On the 16th February 1991 approximately 1.5 hours after the earthquake, seiche oscillations of period about 18 minutes were recorded by sea-floor sensors installed inside the Malokuril'skaya Bay and by standard tide-gauges installed along the coast. (Fig.3). Calculations of the arrival time of the tsunami waves generated by the earthquake were made. In Fig.1 a map of times of tsunami propagation calculated for the Shikotan earthquake is given. It was estimated that the time of propagation of a wave from the source to Shikotan was about 90 minutes.

It is known that in closed bays tsunami waves respond to seiche oscillations the frequency of which is determined by the shape and the size of the basin, the bottom relief, the width and depth of the mouth etc.(Rabinovich and Levyuant 1992). The lowest (zero) Helmholz mode is related to the availability of a channel connecting the bay with the open ocean. Sea level oscillations reach their maximum usually on the side opposite the mouth. On the contrary, the minimum variations of sea level are observed directly in the channel where the seiche node of the zero mode is located, together with the maximum currents velocities. It is interesting to compare these characteristics with the description of the tsunami in Malokuril'skaya Bay on the 13th October 1963. "The maximum water rise equal to 1.5 m was inside the bay just opposite its entrance. Near the throat the height of the tide was less and on the rocky coast in the strait, the observation post of the hydrometeorological station showed that water did not rise even to the first reference mark (0.5m). Thus, the ocean surface during maximum tide was increasing from the mouth to the top of the bay, the average angle of inclination of water surface was 2 seconds".

On the 16th February 1991 the observed increase of the amplitude of seiche oscillations inside the bay led to an increase of the mean square root of deviations of sea level relative to the position of its equilibrium; about 3 times for the filtered record in the frequency range from 0.5 - 30 cp/h (Fig.2). The duration of the "seiche storm" was about 5-6 hours. At the same time, the record obtained outside the bay showed that the arrival of the tsunami was scarcely noticed.

2.2 ANALYSIS OF TIDE-GAUGE RECORDS

In 1991 two synchronous records of ocean level oscillations at 1 minute intervals were obtained near the entrance into Malokuril'skaya Bay. The duration of the time series used for the analysis of records was from 7.2.1991 to 18.2.1991. In addition to the mixed diurnal and semidiurnal tides with amplitude 40-50 cm, seiches with periods of circa 18 minutes and with amplitudes not exceeding 8-10 cm were clearly recorded.

The records were specially processed including one of main tide harmonics, precalculation of the tide, subtraction of the tide from the records and finally high frequency filtration. In the initial records the mean square root deviations were about 90 cm and after extraction of the tidal factor became equal to 14-15 cm and after high frequency filtration reached only 5-6 cm.

The spectra of the background oscillations of the ocean level out and inside the bay were calculated (Fig. 3). In the spectrum calculated for inside the bay the peak is distinguished at a frequency of 0.05 cp/min. In the spectrum derived from the record outside of the bay the peak at the same frequency is one of an order of magnitude less. In Fig.3 the spectra of the ocean level oscillations related to the period when they were supposedly saturated by tsunami are shown. Spectra were calculated for a time interval of 8 hours.

The increase of energy (the spectral density) of ocean level oscillations related to the tsunami is in the frequency range 0.01 - 0.1 cp/min. The existence of the peak to 0.05 cp/min is caused by seiche oscillations in the bay. Small maxima distinguished in the spectrum of the ocean level oscillations outside the bay seem to be related to oscillations diffracted from the neck of the basin to the open sea.

The frequency-amplitude response factor was calculated to estimate the relation between the records of ocean level oscillations outside and inside the bay. This response factor depicts the relation between input (outside the bay) and output (inside the bay) signals and background oscillations of the ocean level (Fig.4). It shows that in the range of low frequencies the changes of ocean level in both points are synchronous with a relation factor equal to 1. In the region of the frequency 0.05 cp/min a sharp peak is observed caused by resonance excitation of the zero Helmholz mode in the bay.

Maximum value of the amplification factor is about 4. By passing through the resonance frequency the jump is equal to π in the phase observed. At higher frequencies the value of the frequency-amplitude response factor decreases, the waves of frequency .08 - 0.2 do not "penetrate" inside the bay. Some increase of the factor not exceeding 1 is observed at frequencies of 0.2 - 0.4 cp/min. These seem to be associated with seiche modes of higher orders. Thus the shape of Malokuril'skaya Bay may be considered as a resonator that effectively amplifies waves coming from the open ocean in a period range of 15-30 minutes.

3.1 ESTIMATION OF PERIODS AND Q-FACTORS OF SEICHES IN THE MALOKURIL'SKAYA BAY

The record of ocean level oscillations in the bay is the cumulative effect of decaying sinusoids of a period of about 18 minutes. It is possible to compare such time series with the model of "noise" of a random process generated by the Poisson process. In this case oscillations can be presented as the sum of a large number of short time impulses of different amplitudes described as decaying sinusoids. These oscillations may be expressed by the non-uniform linear equation:

$$xx_{tt} + 2\delta x_t + W^2 x = \epsilon(t) \tag{1}$$

where $\epsilon(t)$ - is the random forcing influence of the white noise type,

W - frequency
δ - decay decrement

It is known that the linear system of (1) type may be brought into line with the statistical autoregression model of the 2nd order (6)

$$x_1 + a_1 x_{1-1} + a_2 x_{1-2} = \epsilon_1 , \tag{2}$$

where ϵ - is the random value such as $< \epsilon_i \epsilon_j > = \delta^2 \delta_{ij}$,

$< >$ is the sign of averaging
δ^2 - δ the dispersion of ϵ_i

The parameters of the linear system (1) may be expressed by means of the autoregression factors:

$$e^{-\delta} = J a_2 \tag{3}$$
$$\cos(\Omega t) = -1/2, a_1 / J a_2, \tag{4}$$

where $\Omega^2 = W^2 - \delta^2$.

In more general cases the linear oscillating system may be of a higher order. It includes the combination of different oscillators:

$$x^{(n)} + (b_{n-1})^{n-1} + \ldots + b_1 x + b_o = \epsilon(t), \tag{5}$$

In order to make the random process related to this linear system stationary equation (5) should have only decaying solutions.

The autoregression model (AR) of m-order (m>n) may be the discrete analogue of the linear system (5) particularly if (5) is written as a finite-difference equation.

Stochastic equation (5) may be brought into line with the model of discrete random process of autoregression of m-order:

$$\sum_{j=0}^{m} a_j X_{i-j} = \epsilon_i. \qquad (6)$$

Accordingly the condition of stationary character of the model (5) is the requirement imposed on the roots of the characteristic equation:

$$\sum_{j=0}^{m} a_j Z^{-j} = 0. \qquad (7)$$

All the roots of equation (7) should lie inside the circle with radius equal to 1 on the complex plane.

The real roots of equation (7) correspond to solutions (5) and the complex conjugated pairs of roots (7) correspond to solutions of type

$$e^{-st} \cos (wt).$$

Thus, autoregression filtering of the high order may be substituted for a succession of simple filters of the first and second orders. Knowing the factors of these components of the filters it is possible to calculate the frequency and decay decrement of each elemental oscillator in the respective linear system (5).

The programme to estimate the parameters of the linear system (6) was worked out on the basis of estimation of the AR model of the random process. An algorithm provides for the estimation of the order of AR-model in accordance with the Akaike criterion (Akaike 1974). The calculated autoregression filter is separated into the components of the first and second order and then the parameters of the elemental oscillators - the frequency Ω and Q-factor ($Q = 1.2w/\delta$) are calculated for each of them.

In order to estimate the contribution of each component (of the individual oscillator) the dispersion of the input of the AR-filter was calculated on condition that at output the signal is converted into the white noise with dispersion equal to 1:

$$\delta^2 = 1 + a_1^2 - a_2^2 - 2 a_1^2 / (1 + a_2). \qquad (8)$$

In Table 2 the results of the estimation of the periods (T and $2\pi/\Omega$) and Q-factors of the main four components of the linear model for the records of ocean level oscillations in the bay are given. The optimum order of the AR-model is 20. Two roots were actual, the respective solutions are features of decaying exponents and are not included in the Table. In total 18 of the roots conjugated in pairs of roots give 9 solutions in the form of decaying sinusoids. Only 4 of them correspond to "smooth" solutions and the remaining 5 are "high frequency" (noisy) ones and are not included in the Table. The values of the solutions however, are included. The dominating oscillations in the system are harmonic of a period 18.5 minutes with a Q-factor about 10 corresponding well with the observed seiche period (Helmolz mode) in the Malokuril'skaya Bay. This may be compared to the values of periods and Q-factors given by

Rabinovich and Levyuant (1991). They are 18.6 - 18.8 and 12 and 14 respectively.

TABLE 2

Parameters of main components of oscillators in model calculated for ocean level oscillations recorded in Malokuril'skaya Bay.

Period T (min)	Q-factor	Conditional dispersion
18.54	10.3	141.0
10.88	1.2	3.5
7.32	2.8	3.5
5.04	4.9	2.8

4.1 CONCLUSION

The measurement of very small tsunami waves by means of ocean-bottom hydrostatic pressure sensors is of particular interest because they make it possible to extend the statistics of detected tsunamis and to estimate more exactly their recurrence. The standard shore tide gauges provide precision of measurements no higher than 2-4 cm. In this sense the registration of a tsunami with a small amplitude on the 16th February 1991 caused by a moderately strong earthquake (magnitude 6) was quite noteworthy. It is possible that in this case the effect of the radiated waves was displayed with the result that the seismic source was oriented transverse to the continental slope.

The pressure sensor installed inside Malokuril'skaya Bay was the most "sensitive". The selective response of this isolated basin in relation to the different waves is manifested in the amplification of the ocean level at a period of 15-30 minutes (typical for tsunamis) and in the suppression of the background high frequency noise.

The investigations of tsunami inundation on sections of the coast with clearly expressed frequency-selective properties (eg bays, gulfs, estuaries etc.) is very important for tsunami zonation. The natural oscillations in bays or harbours excited by waves coming from the open ocean may be a serious danger to moored vessels and port facilities. The method used to calculate the characteristics of natural oscillations (seiches) outlined in this work makes it possible to estimate not only the frequency of the individual modes but the quality of the resonator characterising the velocity of attenuation for the energy of the seiche oscillations at a particular frequency.

ACKNOWLEDGEMENTS

The authors would like to thank Dr A A Poplavsky for help in the calculation of the tsunami refraction diagram.

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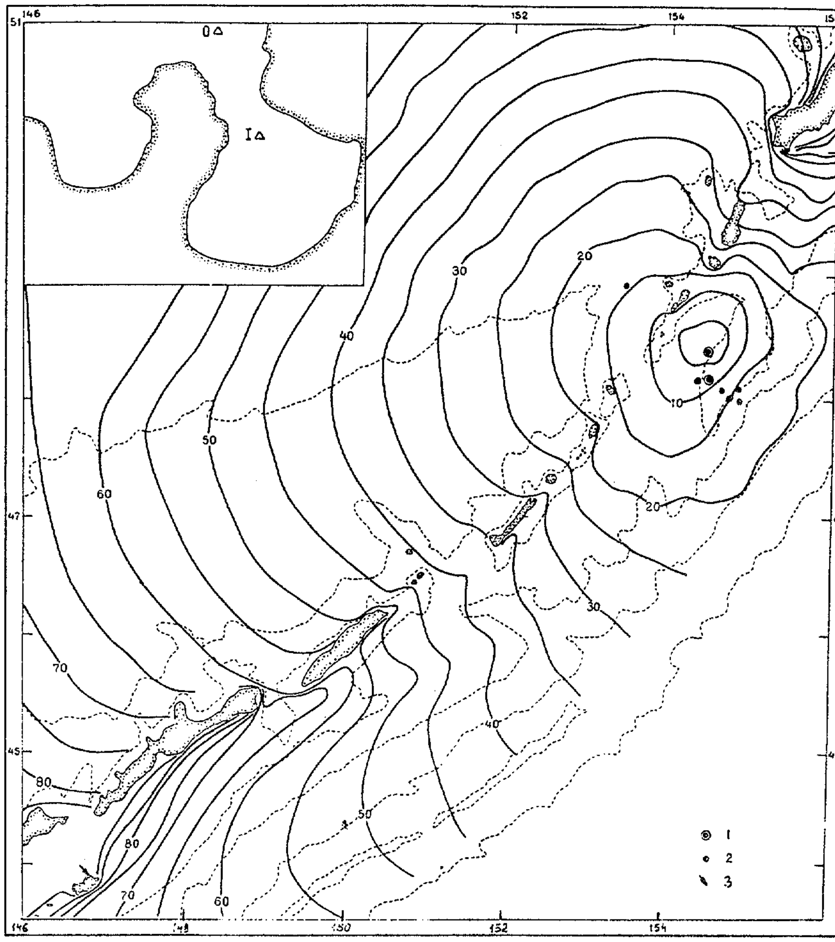

Fig. 1 Location of seismic sources near Shikotan Island calculated times of tsunami propagation (mins) and locations of outside (o) and inside ( ) ocean-bottom tide-gauges in Malokuril'skaya Bay. Possible epicenter locations showed by dots.

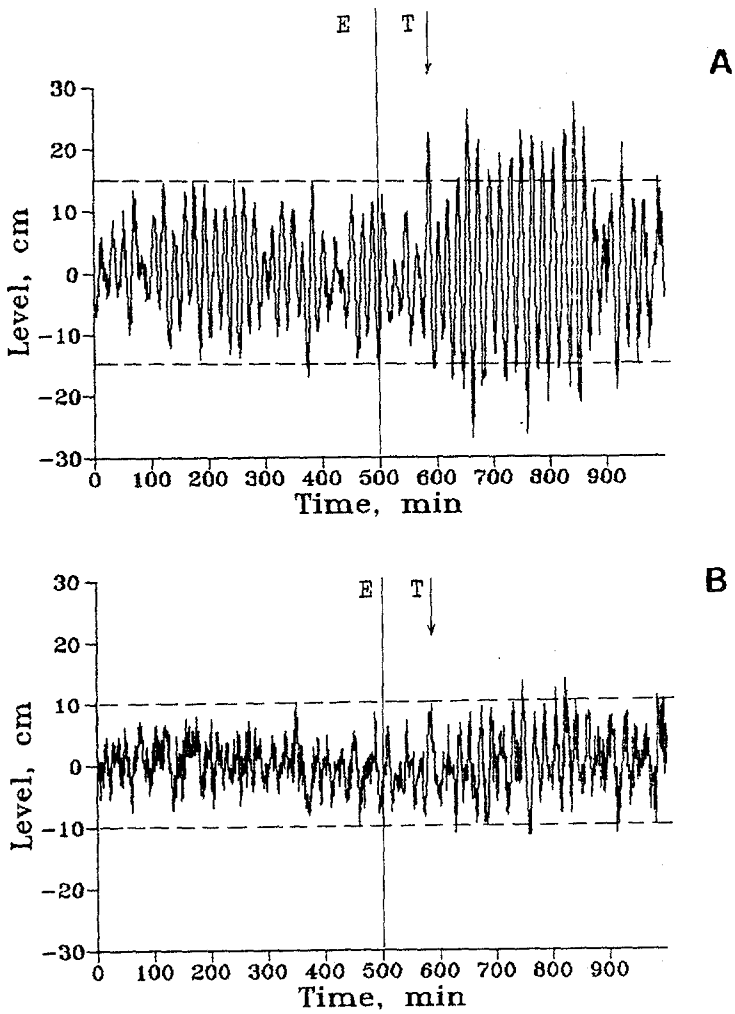

Fig. 2 Filtered records of ocean bottom tide gauges with fluctuation in the frequency pass band 0.01 - 0.05 cp/min; a) inside Malokuril'skaya Bay, b) outside the bay. The letters E and T indicate the time of the earthquake and the calculated time of the arrival of the tsunami wave. The hatched line shows the interval equal to the double mean square root deviation of the level for each record.

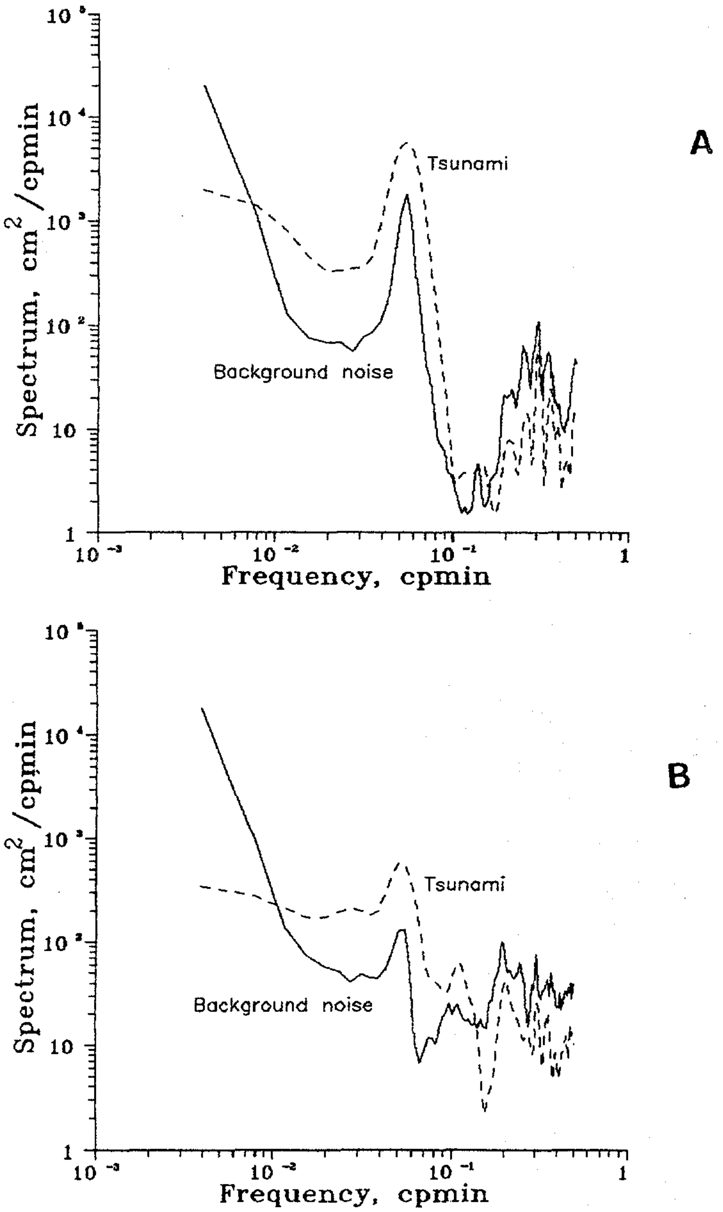

Fig. 3 The spectra of ocean level oscillations inside Malokuril'skaya Bay (a) and outside the bay (b). The solid line indicates the spectrum equivalent to the background ocean level oscillations. The dotted line shows the spectrum calculated for the period of 8 hours after tsunami arrival.

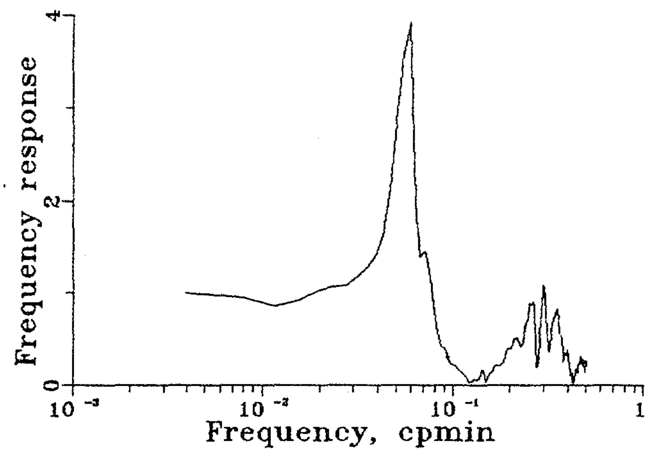

Fig. 4a Frequency-amplitude response function (frequency characteristics) relating the input signal (outside the bay) to the output signal (inside the bay).

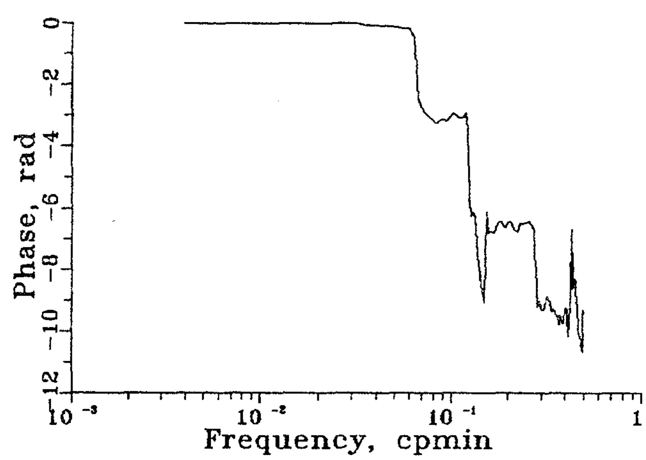

Fig. 4b Respective phase characteristics corresponding to Fig. 4a.

HAZARDS '93

DISASTER MITIGATION: SCIENTIFIC AND SOCIO-ECONOMIC ASPECTS

Disaster Mitigation: Scientific and Socio-Economic Aspects

第五届国际自然和人为灾害会议

FIFTH INTERNATIONAL CONFERENCE ON NATURAL
AND
MAN-MADE HAZARDS
29 AUGUST - 3 SEPTEMBER, 1993
QINGDAO, CHINA

Organized by: INTERNATIONAL SOCIETY FOR THE PREVENTION AND MITIGATION OF NATURAL HAZARDS (NHS)
THE IAPSO COMMISSION ON NATURAL MARINE HAZARDS
THE RESEARCH COMMITTEE OF NATURAL DISASTERS, CHINESE ACADEMY OF SCIENCES
THE INSTITUTE OF OCEANOLOGY, CHINESE ACADEMY OF SCIENCES

BACKGROUND

This international symposium is the fifth in the continuing interdisciplinary series begun in 1982, with the first being held in Honolulu, USA. The second one was held in Rimouski, Canada in 1986, the third in Ensenada, Mexico during 1988, and the fourth meeting was held in August 1991 in Perugia, Italy. The objectives of this series of symposia on natural and man-made hazards are to promote the advancement of hazard sciences, to perceive and explore those aspects that may be similar among some of the various hazards, to review the newest developments in a few selected fields, and also to outline new directions for future research.

HAZARDS–93

The United Nations declared the 1990s as the International Decade for Natural Disaster Reduction (IDNDR). Its objective is to prevent or mitigate natural disasters and the loss of life, property damage, and social and economic disruption they produce worldwide. The 1990s is also a time when for many countries coping with disasters is becoming virtually synonymous with development. The cost of rehabilitation and reconstruction in the wake of disasters is consuming available capital, significantly reducing the resources for new investment. Tackling this problem requires a sound evaluation of disaster mitigation policies and tools.

The theme for HAZARDS–93 is **DISASTER MITIGATION: SCIENTIFIC AND SOCIO-ECONOMIC ASPECTS.** The organizing committee welcomes papers on all aspects of natural and man-made disasters, but priority will be given to those emphasizing the mitigation aspects. The global impact of disasters is on the increase as can be seen from the recent storm surge impact in Bangladesh in April-May 1991, the heavy rain and flood disasters in China in May-July, 1991 and the impact of the Mount Pinatubo volcano in the Philippines, hurricane Andrew and typhoon Iniki in the United States of America in 1992. It is felt that it is time to develop ways and means of mitigating disasters and not just to do the basic science of understanding these.

All those interested in natural and man-made hazards and their mitigation will not only find much value in the formal sessions, but will also have a unique opportunity to confer personally with eminent researchers and policy makers in this important field. The Organizing Committee invites all scientists, engineers and policy makers and others involved in natural and technological hazards to participate actively in HAZARDS-93.

SPONSORSHIP

The International Society for the Prevention and Mitigation of Natural Hazards (NHS), the IAPSO Commission on Natural Marine Hazards, the Research Committee of Natural Disasters of the Chinese Academy of Sciences (CAS) and the Institute of Oceanology of CAS are the principal scientific organizations sponsoring HAZARDS-93.

To date, the following Chinese organizations have officially agreed to co-sponsor the symposium:

- National Natural Sciences Foundation of China
- Department of International Cooperation of the State Seismological Bureau
- Shandong Provincial Commission of Agriculture
- Qingdao Branch of the People's Insurance Company of China
- Chinese Academy of Meteorological Sciences
- The United Research Center for Water Problems, the Chinese Academy of Sciences
- China Association of Disaster Prevention
- Chinese National Committee for IAPSO
- Chinese Society of Oceanology and Limnology
- Chinese Research Society of Storm Surge and Tsunami
- Commission on Marine Meteorology, Chinese Meteorological Society

Non-Chinese organizations who have officially agreed to co-sponsor the symposium include:

- The Intergovernmental Oceanographic Commission (IOC) of UNESCO
- The Tsunami Society, Honolulu, U.S.A.
- International Association of Geodesy – Special Commission on Marine Position – Working Group on Practical Applications

SCIENTIFIC PROGRAM

Under the umbrella of HAZARDS−93 several symposia and workshops will be held, dealing with all aspects of natural and technological disasters, with particular emphasis on the mitigation aspects and preventive measures. Keynote speakers, special invited lectures and contributed papers on current practices and research activities will be grouped into the following themes:

— Geological hazards (earthquakes, volcanoes, landslides, snow avalanches, soil erosion)
— Meteorological hazards (cyclones, droughts, desertification, forest fire, etc.
— Hydrological and marine hazards (tsunamis, storm surges, floods, sea-level rise, ice, icebergs, and marine biological hazards)
— Technological and man-made hazards (air and water pollution, deforestation)
— Disaster prevention, mitigation and management
— Economic, social and political aspects
— Tools and techniques for disaster assessments
— Public education and preparedness
— Risk assessment problems
— The IDNDR: a perfect chance to put hazard research into practice

These themes will be organized into formal sessions for the symposium and will be described in detail in the final program. Each session will commence with a keynote address by an invited eminent scientist, to review the state of knowledge and set the stage for the contributed presentations which will be limited to 20 minutes, including the question period. Authors preferring to offer a poster presentation are requested to write to the Chairman of the Scientific Committee. The plenary session, planned for September 3, will focus on practical responses to natural and man-made hazards, the identification of preferred lines of investigation, and actions required to improve measures of hazard mitigation.

Those who have already been invited to present keynote presentations include:

J.P. Bruce, Chairman, IDNDR Scientific and Technical Committee, Geneva, Switzerland
Dr. M.L. Khandekar of Atmospheric Environment Service, Downsview, Ontario, Canada
Professor Joane M. Nigg, Director, Disaster Research Center, University of Delaware, Newark, DE, U.S.A.
Dr. Iouri Oliounine Senior Assistant Secretary, Intergovernmental Oceanographic Commission, UNESCO, Paris, France
Professor Eliot Tepper, Department of Political Science, Carleton University, Canada
Dr. G. Berz, Münich Reinsurce Company, Germany

Other topics which will be presented during the symposium include: plans and programs in support of the International Decade for Natural Disaster Reduction (IDNDR); several case studies of disaster management in developing countries; environmental impacts of the Gulf war; natural disasters in Bangladesh; recent flood hazards in China, Pakistan and European countries; the deadly storm surges in Bangladesh, India, China, and U.S. Gulf Coast; hazards from volcanic eruptions, near-shore earthquakes and landslides; soil and coastal erosion; human adjustment to snow avalanche hazards; wind storm and other severe weather systems; release of toxic chemicals in the atmosphere and other air pollution problems; desertification processes in Africa; severe drought hazards in Ethiopia and other regions; environmental and ecological impacts of human and industrial discharge to the coastal marine environment; tsunami generation, propagation, modelling, mitigation and their coastal impacts.

PROGRAM OUTLINE

Sunday, August 29	Arrival and registration
Monday, August 30	Opening Ceremonies: Plenary Session 1: Keynote speakers; evening reception given by the Institute of Oceanology, Academia Sinica
Tuesday, August 31 to Thursday, September 2	Formal scientific sessions; Excursion to Mt. Laoshan; Banquet
Friday, September 3	Meetings of societies and commissions; Plenary Session 2; Recommendations, Closing Ceremonies
Saturday, September 4	Post-Symposium Tours

International Symposium

HAZARDS – 93

Qingdao, China

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REGISTRATION FEES:

Participants	NHS Members	Everybody Else	Enter Amount
Before May 15, 1993	$330 US	$350 US	
After May 15, 1993	$380 US	$400 US	________
Accompanying Person(s) Rate:			
Before May 15, 1993		$150 US	
After May 15, 1993		$165 US	________
		TOTAL:	________

Pre-registration payment <u>MUST</u> accompany registration form.
Please make cheque payable to "HAZARDS – 93 SYMPOSIUM" in U.S. funds, drawn on a US or Canadian bank only. Cheque or bank draft are accepted. Credit card payments cannot be accepted.

ADVANCE REGISTRATION WOULD BE APPRECIATED

GENERAL INFORMATION FOR PARTICIPANTS AND GUESTS

How to reach the site of the HAZARDS−93 Symposium

The seat of the Symposium is located in Qingdao. Participants coming by air are advised to reach Beijing, Shanghai or Hong Kong first, no matter the direction of entry. Then they can easily arrive in Qingdao by either train or plane. Please contact your Travel Agent for exact time table.

NOTE

VISA Passports must be valid for at least six months subsequent to the date of the proposed visit to China. Please obtain your Entry Visa from your nearest Chinese Consulate before departure. Letters of invitation can be obtained from the Local Organizing Committee of Hazards−93 Symposium (see address on page 9). ***before*** June 1, 1993.

1. On August 29, 1993 the Local Organizing Committee (the secretariat) will provide bus service all day (24 hours) free of charge from Qingdao Airport or Qingdao Railway Station to the hotel. Those who will not arrive on August 29 but want to be met at the Qingdao Airport or Qingdao Railway Station, are kindly requested to inform the secretariat beforehand in order to arrange for a bus. But they will pay the bus fare themselves. The cost for this service between Airport-Hotel and between Railway Station-Hotel, for those participants who will not arrive on August 29, is $20 US and $5 US respectively.

2. Following the close of the Symposium, transportation from the hotel to Qingdao airport or the railway station will be available for a fee. For those taking part in any of the post-symposium tours (1−5), this fee is included in their package price.

FOR MORE INFORMATION PLEASE WRITE TO ONE OF THE FOLLOWING:

Professor Mohammed I. El-Sabh
Département d'Océanographie
Université du Québec à Rimouski
300 Allée des Ursulines
Rimouski, Québec, Canada, G5L 3A1
Tel: (418) 724-1707
Fax: (418) 724-1842, Telex: 051-31623

Professor Franco Siccardi, Vice-President, N.H.S.
Hydraulic Institute
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Genova, Italy
Tel: + 39-10-353-2401
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Dr. T.S. Murty
Institute of Ocean Sciences
P.O. Box 6000
Sidney, British Columbia, Canada, V8L 4B2
Tel: (604) 363-6311
Fax: (604) 363-6390, Telex: 0636700764

Professor LE Kentang, Secretary-General
Local Organizing Committee, Hazards−93
Institute of Oceanology, Academia Sinica
7 Nanhai Road, Qingdao, 266071, China
Tel: 279062, Ext. 378, 386
Fax: (086532) 270882, Cable: 3152
Telex: 32222 Answer Back ISS CN

IMPORTANT DEADLINES

March 15, 1993:	Submission of Abstracts
May 15, 1993	Pre-registration
May 15, 1993	Hotel Reservation
June 1, 1993	Deadline for requesting letter of invitation for Entry Visa
July 15, 1993	Deadline for cancellation
December 31, 1993	Submission of full MSS

APPLICATION FOR MEMBERSHIP

THE TSUNAMI SOCIETY
P.O. Box 8523
Honolulu, Hawaii 96815, USA

I desire admission into the Tsunami Society as: (Check appropriate box.)

☐ Student ☐ Member ☐ Institutional Member

Name ______________________ Signature ______________________

Address ______________________ Phone No. ______________________

____

Zip Code ______________________ Country ______________________

Employed by ____

Address ____

Title of your position ____

FEE: Student $5.00 Member $25.00 Institution $100.00

Fee includes a subscription to the society journal: SCIENCE OF TSUNAMI HAZARDS.

Send dues for one year with application. Membership shall date from 1 January of the year in which the applicant joins. Membership of an applicant applying on or after October 1 will begin with 1 January of the succeeding calendar year and his first dues payment will be applied to that year.