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GENERATION, WAVE FORM, AND LOCAL IMPACT OF THE SEPTEMBER 19, 1985 MEXICAN TSUNAMI

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ABSTRACT

The September 19, 1985 earthquake in México was associated with an atypical subduction, with small plate angle, small vertical ground uplift, and unusually long duration; and occurred within a narrow and deep underwater portion of the zone of faulting. Apparently, the leading tsunami wave which was generated had low edge steepnesses and slow surface rise and fall speeds. This particular form of the wave explains the low level of drag-related damage at Lázaro Cárdenas and other coastal communities adjacent to the subduction zone. Damage due to inundation was indeed noticeable, and a tsunami bore travelled 9.5 km upstream the Balsas River. Future local tsunamis in this region may have similar potential damage type characteristics. A rough estimate of tsunami run up heights which may be expected, shows that the industrial and port facilities of Lázaro Cárdenas are not risk free.

1. Introduction

The subduction thrust fault of the Cocos plate adjacent to continental western México is considered one of the most active in the western hemisphere. During this century México has had 42 earthquakes with magnitude greater than 7 associated with this subduction zone, while California has had only 5 associated with its correspondingly long San Andreas fault system (Singh et al, 1981; and 1984). At least 15 of the earthquakes in the last three centuries were the source of locally destructive tsunamis with waves from two to six and even nine meters high (Soloviev and Go, 1975). The southeastern portion of the trench, just in front of Acapulco, known as the Guerrero seismic gap, had not been the source of any large event since 1911. The likelihood of a rupture there, with one or several large events in the next decade or so, remains high (Anderson et al, 1985).

Hence, tsunamis contitute a real and permanent threat to the central west coast of México in front of the Middle America Trench. However, there is a general belief in México that tsunamis do not constitute a real hazard; and most fishermen and other coastal settlers which have been hurt in the past by tsunamis, are not presently protected (Pararas-Carayannis, 1985).

The Civil Defense Plan of the Nation does not include any instruction for tsunamis and the National Urban Development Plan does not contain any specific policy for the tsunami hazard zones planning. Nevertheless, the National Disaster Prevention System indicates as some of its priorities the following tsunami research fields: generation, propagation and coastal effects modeling, risk evaluation, and damage mitigation. There is also an urgent need to determine the extent of buildings and services in the hazard zones, to decrease the potential loss of life and property. This study deals with some of these subjects in connection with the September 19th, 1985 earthquake-associated tsunami and its coastal effects at the city and port of Lázaro Cárdenas in the state of Michoacán, México.

2. Mechanism and Bottom Displacement Features of the September 19th, 1985 Earthquake.

The earthquakes were associated with the subduction of the northwest portion of the Cocos plate, called the Michoacán seismic gap, beneath continental México. On September 19, two subevents occurred with 26 seconds time lag and 80 km source separation, with the summing effect partially responsible for the unusually long duration of the whole event (Anderson et al, 1985). Each source had an estimated 3 to 4 km radius dimension

(Houston and Kanamori, 1986). The gradual and slow beginning suggests that the faulting was not immediately large, and that there was a delay in the major faulting after the initial event (U.N.A.M. Seismology Group, 1985).

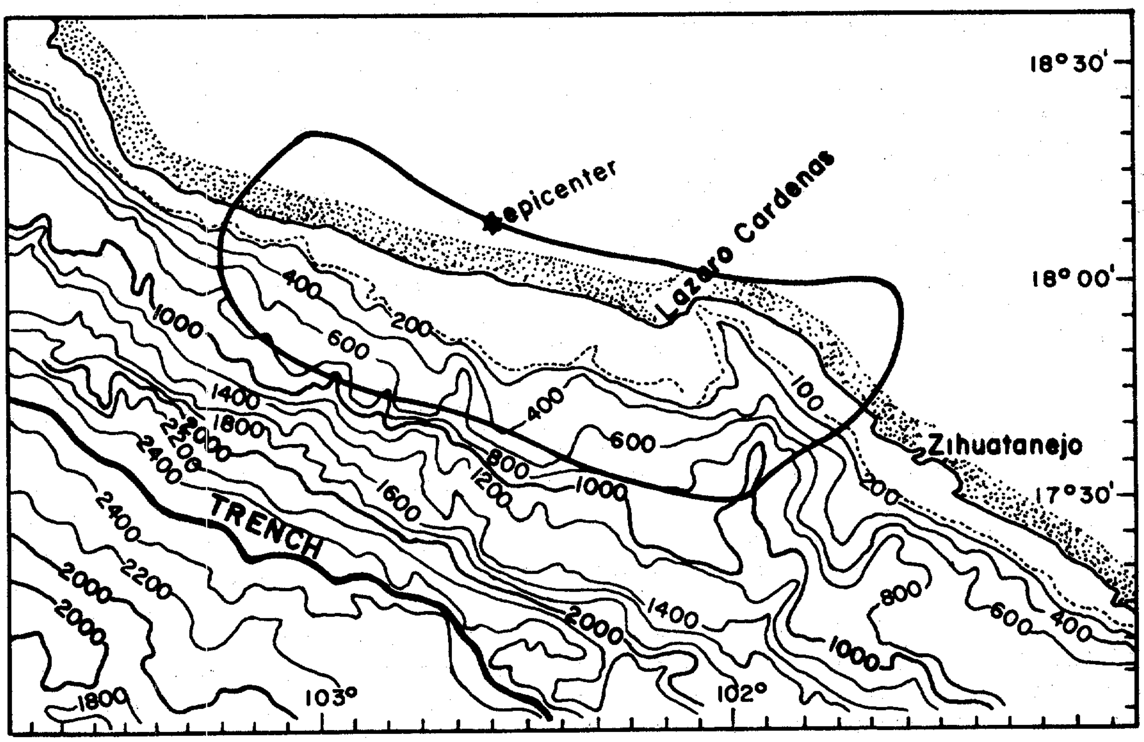

Fig. 1. Ocean bed topography at the september, 1985 earthquake rupture zone (depths in fathoms) (1 fathom = 1.83 meters).

Figure 1 shows an outline of the rupture zone based on after shock records, from which a fault area of 50 by 170 km^2 is derived. A vertical ground uplift of 0.93 m was obtained through double integration of accelerograms on the site (Bodin and Klinger, 1985), and confirmed by a 1 meter uplift observed by eyewitnesses along the coast. The plate angle of subduction at the trench was relatively small (15°), with anomalously low dynamic stress drop and energy release compared with average earthquakes and unusually low peak accelerations in the near epicentral field (González-Ruíz et al, 1985). Das and Kostrov (1986) called this a "slow" or "weak" earthquake.

In summary, the September 1985 double earthquake was a very atypical subduction thrust earthquake, with small plate angle, small vertical ground uplift, within a relatively narrow, short and deep underwater portion of the faulting zone, and of unusually long duration.

3. Tsunami Generation, Wave Form, and Related Damage Potential

The detailed form of submarine tectonic displacement is extremely difficult to determine since the deformations cannot be observed directly, and in most cases pre-earthquake and post-earthquake bathymetries are unknown with enough precision to be compared. However, it seems that the time-displacement history of a section of ocean bed deformation, during an earthquake, can be characterized by: the amplitude of the vertical displacement, halfwidth of the block section (b), and a characteristic time of the displacement (T).

Hammack (1972) shows that 3 adimensional combinations of the above parameters (including also the water depth h) are important in determining the characteristics of the generated tsunami wave: a disturbance-amplitude scale, a disturbance-size scale, and a time-size ratio $T(gh)^{1/2}/b$ (g being the acceleration of gravity).

Particularly important as an indicator of the form of the leading tsunami wave is the time-size ratio value, with two distinctive extreme regions:

a) Impulsive: Very fast bed movement, time-size ratio much less than one, displaced water surface similar in form to the deformed bed (block shape), water surface at both edges of the wave rises and falls rapidly.

b) Creeping (a word of general use in seismology to describe slow adjustments of earthquake faults): Very slow bed movement, time-size ratio much greater than one, water surface at the leading edge of the wave rises slowly, and at the back edge falls even more slowly, profile resembling a bore.

Results from experiments performed with a movable bed wave tank were consistent with the theoretical wave profiles derived (Hammack, 1972). Figure 2 shows the very distinctive wave profiles for each extreme region of generation. Liu and Earickson's (1983) finite-difference numerical solution of the long wave equations for this problem, also agrees with the experiments.

Calculations show that the 1964 Alaska tsunami generation was typically Impulsive, while the 1985 Mexican tsunami approached a case of Creeping. In fact, Hammack (1972), using data from

Plafker (1969) on a 600 km by 224 km fault area, with halfwidth of 80 km, average water depth of 180 m and a characteristic time of 2 minutes obtained an average time-size ratio of 0.06 for the Alaska tsunami (within the impulsive region). Approximate halfwidth and water depth for the underwater region of the rupture zone during the 1985 Mexican tsunami are: b = 10 km, and h = 730 m (Figure 1). An estimated characteristic time of 4 minutes gives a time-size ratio of 2.03 (approaching Creeping region). Clearly, the differences between the Mexican and the Alaskan events are the larger water depth and characteristic time, and the shorter half width associated with the Mexican tsunami.

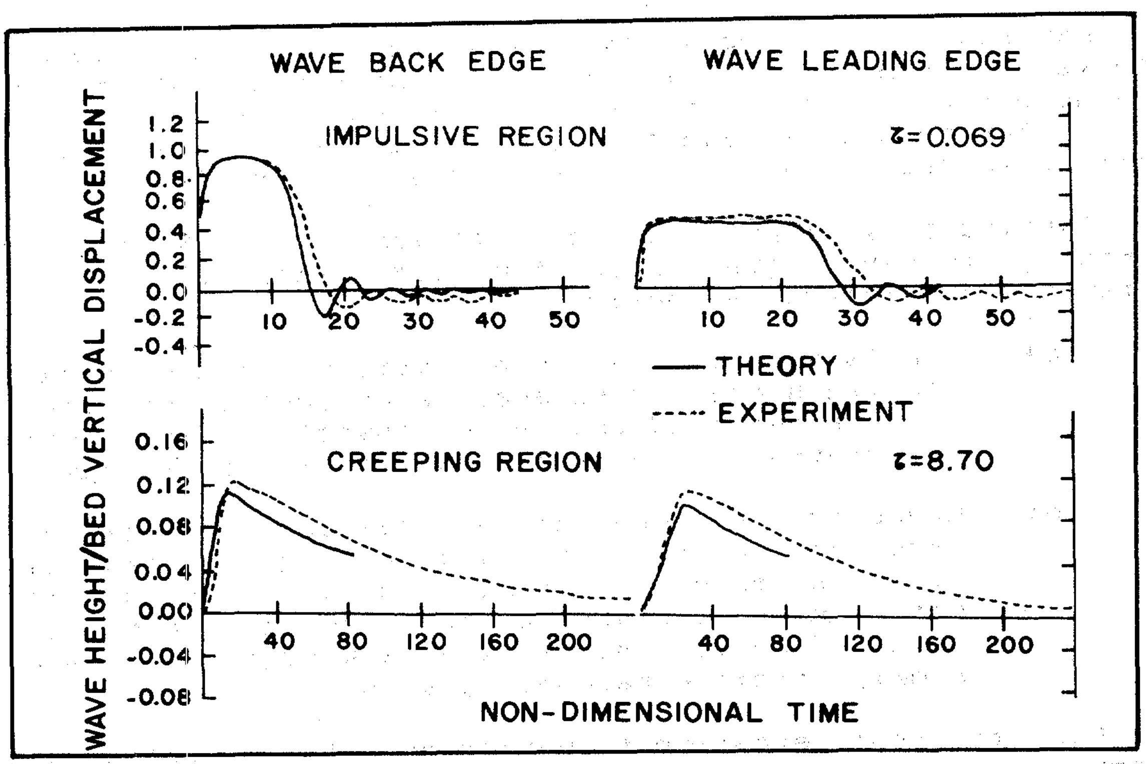

Fig. 2. Typical wave profiles in each extreme region of generation for exponential bed displacement (τ=time size ratio) (from Hammack, 1972).

Independent of the vertical bed displacement height, tsunamis generated by impulsive movement seem to have a nearly breaking wave type profile, while those generated by creeping seem to have a nearly bore wave type profile. Drag and inertia wave forces on coastal structures depend on the square of water particle velocities and the accelerations, respectively (Horikawa, 1978). For shallow water wave approximation, particle velocities and accelerations depend linearly on the water surface slope (Le Méhauté, 1976). Hence, impulsive-breaking type tsunamis should present much more potential for inertia-drag damage than creeping bore types.

For the September 19, 1985 tsunami events, the leading edge of the tsunami generating zone of faulting was almost at the shore line. Thus, the wave profile observed at the beach should have had a very similar form to the one at the generation zone. This wave both entered and receded slowly, and developed a bore wave upstream the Balsas River at the city of Lázaro Cárdenas (Figure 3). It did not show any kind of breaking wave pattern at the beaches and left little or no drag associated damage. These observations seem to be related with the form of the leading wave generated, as mentioned above.

4. Damage, Coastal Effect and Bore at Lázaro Cárdenas

Lázaro Cárdenas is a city with heavy industry located at 17° 55' N, 102° 15' W, and has an estimated population of 150,000. The main industries are 2 steel mills, a fertilizer factory, a ship yard, and an oil terminal. Port facilities include a container terminal, metal and mineral docks and grain storage silos (Figure 3). A site inspection of coastal communities affected by the September 1985 earthquakes and tsunamis, from 25 km north to 90 km south of Lázaro Cárdenas was conducted in January 1986.

A local tsunami of approximately 2.5 m wave height, and bore configuration, arrived during the earthquake tremor, flooding a horizontal extension of 500 m inland of the coastal area (Fig. 3). Major damage was attributable to inundation. Little or no impact or drag effect were observed; the water flow neither tipped over empty railroad cars on the tracks close to the beach, nor removed the many frail palapa-built restaurants along the shore. Rocks on the beach groins and river entrance jetties were also left unmoved. Results of the survey, with detailed damage patterns, are presented elsewhere by Preuss et al (1986).

A large perturbation followed by a train of small waves, resembling an undular type bore, travelled up the Balsas River with effects observed as far as the second bridge. Jetties at the entrance, as well as docks at the inner harbor, were covered by a leading wave of 2.5 m height. Upstream the observed bore height declined from 2.0 m to its point of disappearance. Neglecting bottom friction and slope, the average propagation

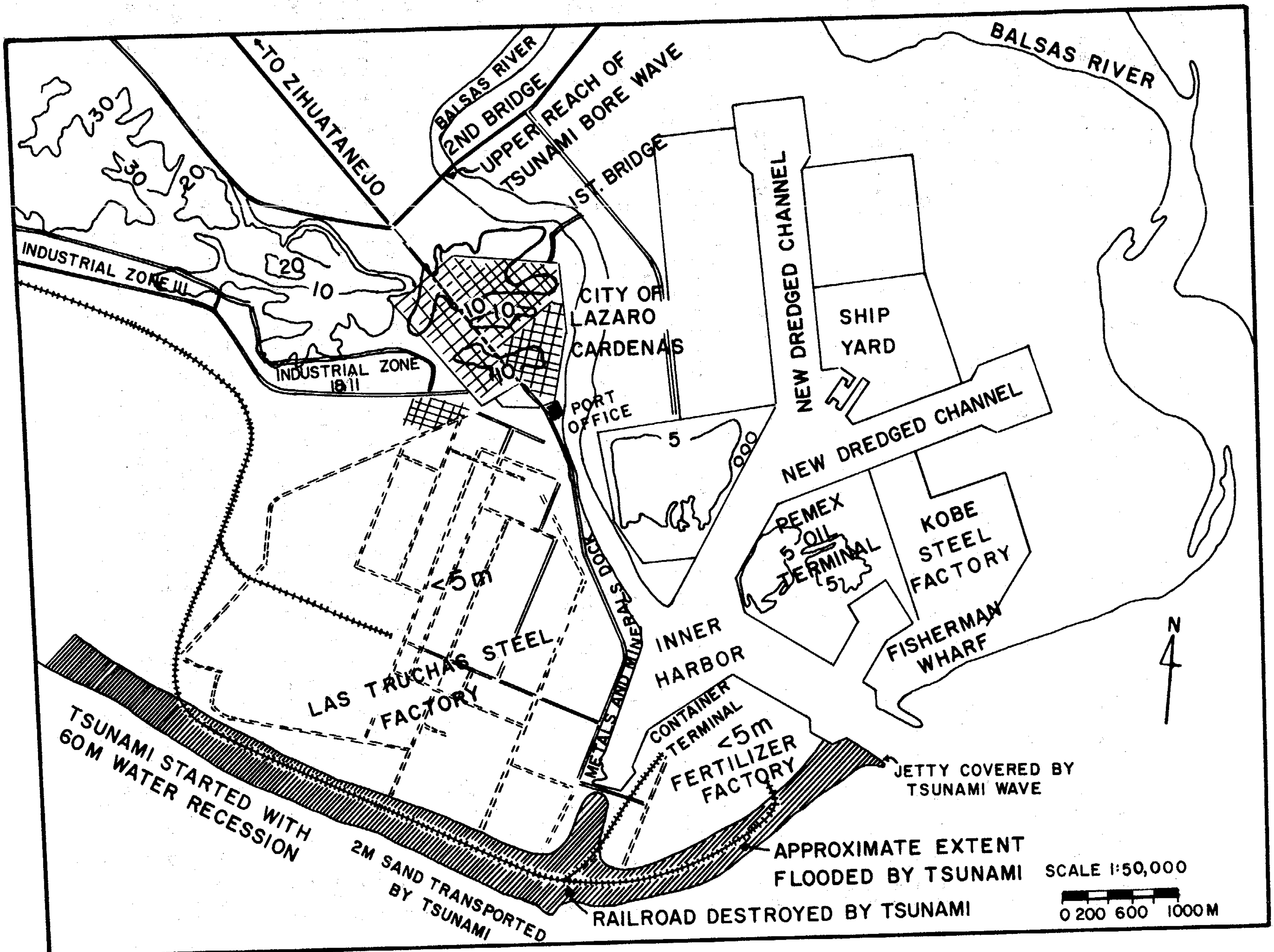

FIG. 3. LAZARO CARDENAS OVERVIEW WITH SEPTEMBER 1985 TSUNAMI EFFECTS.

velocity for a solitary wave of moderate height with respect to water depth, travelling upstream on a subcritical flow, is approximately:

$$c = (gh + 1.5gH)^{1/2} - v \qquad (1)$$

where g is the gravitational acceleration, h is the water depth, H is the wave height, an v is the stream velocity (Barré de Saint-Venant, 1870). With average water depths before the tsunami arrival of 13.9, 2.9 and 2.0 m from the entrance to the inner harbor, inner harbor to the port office, and port office to the second bridge, respectively (Fig. 3), the computed velocities give corresponding times of 2.6, 8.1 and 14.6 minutes for the arrival of the bore at the inner harbor, port office, and second bridge. Eyewitnesses observed approximate arrival times of 10 and 18 minutes at the port office and second bridge, respectively, which are in reasonable agreement with the computed values.

5. Future Tsunami Inundation Risk for Lázaro Cárdenas

It is hard to quantitatively estimate the tsunami inundation risk probabilities for Lázaro Cárdenas within an acceptable an reasonable certainty. However, the 1985 event shows that the statement by Rascón and Villarreal (1975), based on tsunami arrivals at Hawaii, that Lázaro Cárdenas is under no tsunami risk, should be considered as no more than a theoretical speculation. Although Lázaro Cárdenas is an important industrial port for the west coast of México, no tsunami records are available. With the exception of a few months of sporadic recording, no tidal gauge has ever been in operation on the site. Because the city, port and industrial facilities were built only 15 years ago and located on the uninhabited sand shoals of the Balsas River delta, not even historical observations of tsunamis are available. Thus, there is no information with which to calibrate a numerical or even an analytical model on tsunami generation, propagation, and inundation.

Nevertheless, there is evidence that repeat times of large earthquakes along any given portion of the Mexican subduction zone average about 30 to 75 years (Singh et al, 1981), although successive large earthquakes may occur at shorter intervals in the same or nearby regions. Sánchez and Farreras (1986), based on 15 years of recent records, predict a tsunami with a 2.15 m run up height once every 100 years for the neighboring port of Manzanillo, 250 km north of Lázaro Cárdenas. Historic information from Soloviev and Go (1975), and Sánchez (1980) shows that for the entire subduction zone (not necessarily in front of Lázaro Cárdenas), a tsunami with 2.0 m run up height can be expected approximately once every 26 years; a 6.0 m tsunami once in every 54 years or more; and that a 9.0 m one is only remotely possible. An event of the first kind will produce no more than the small amount of damage already observed from the September 1985 tsunami. An event of the second kind will cause minor damage to the industrial zones and port facilities with danger to life, but the urban area of Lázaro Cárdenas, built 10.0 m above mean sea level, will be in no danger. An event of the third kind would be a major disaster with loss of life and extensive property damage.

6. Conclusions

The September 19, 1985 earthquake was a "slow" one, with low rupture speed, long slip duration, and confined to a deep and short portion of the underwater zone of faulting. Consequently, the leading tsunami wave generated had low edge steepnesses with slow surface rise and fall speeds. This particular form of the leading tsunami wave seems to explain the small amount of impact or drag associated damage at Lázaro Cárdenas and other neighboring coastal communities. However, damage due to inundation was noticeable.

Hammack (1972) stated that the impulsive type 1964 Alaska earthquake was indeed a typical major tsunamigenic earthquake, with general characteristics of the generated tsunami appearing to be applicable to most tsunamis. The creeping type 1985 México earthquake seemed to be very atypical, with general characteristics of the generated tsunami applicable to very few others. The fact that most of the west central coastline of México is less than 80 km from the Middle American Trench (Fig. 1) with a short distance for the generated tsunami disturbance in the direction of wave propagation, and with an average water depth at these eventual earthquake faulting zones around 1000 m, indicates that most of the tsunamis in this region may have been or will be nonimpulsive, but rather near to the creeping-bore type, with little impact or drag associated damage. Nevertheless, major damage should be expected from inundation. The city, port and industrial zones of Lázaro Cárdenas are not exempt from this type of tsunami hazard.

Acknowledgments

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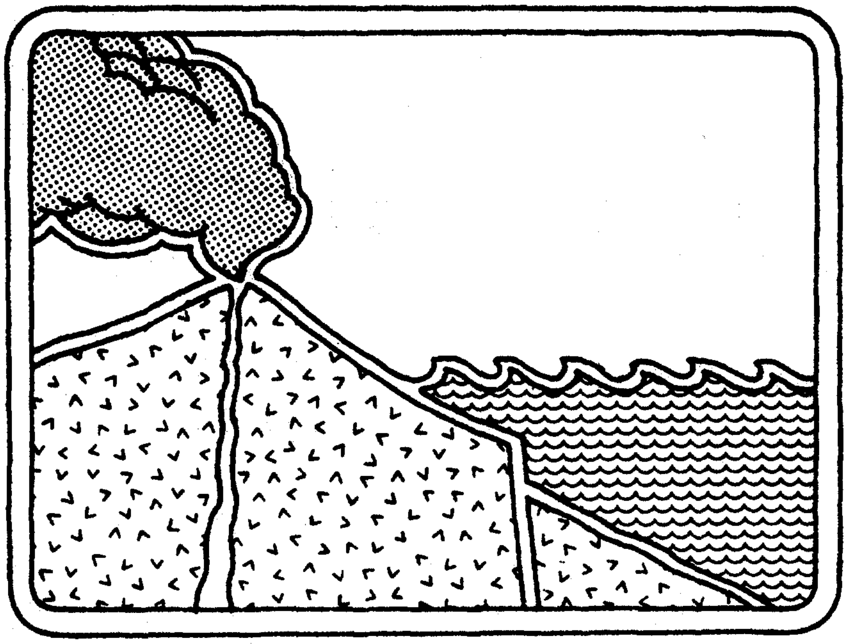

The Foreslope Hills of the Fraser Delta: Implications for Tsunamis in Georgia Strait

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(G.S.C. contribution number 30886)

Abstract

The Foreslope Hills (FSH), an 11 x 6 km area of hummocky bathymetry on the face of the Fraser Delta, is the largest of a series of youthful slump deposits in Georgia Strait. Whether the ultimate cause of this feature is seismic or sedimentological, slumps of this kind pose a direct hazard to marine engineering works. While Georgia Strait is sheltered from even the largest tsunamis originating in the open Pacific, slumps like the FSH can impart sufficient momentum to the water column to cause locally generated tsunamis. Geological evidence suggests that this region of high population density and extensive foreshore development is subject to continued risk.

Introduction

Tsunamis or seismic sea waves are long period water waves which originate from sudden vertical displacements of sizeable masses or areas of seafloor. The cause may be submarine dip slip faulting which accompanies major earthquakes such as occur in tectonically active zones, or large submarine slumps which occur in areas of high relief, high sedimentation rates or occasionally by man made disturbances (Bolt et al., 1975). Documented examples of the former kind where tsunamis are related to earthquake activity include: Honshu, 2 March 1933; Chile, 23 May 1960; Alaska, 28 March 1964 and Mexico, 19 September 1985. Examples of the latter kind include Kitimat, British Columbia, Canada, 27 April 1975 and the Var River, Nice-Cote d'Azur, France, 16 October 1979. The common factor is that large enough masses of seafloor move quickly enough to transfer significant momentum to the water column. For tsunamis which can be related to specific seismic events there is an empirical relationship between tsunami magnitude (M_T) and causal earthquake magnitude expressed by:

$M_T = 2\ M_E - 14$ (Bolt et al., 1975)

where M_E is the earthquake magnitude on the Richter scale. Furthermore one can estimate the energy released in an earthquake from its surface wave magnitude as:

$\log E = 4.8 + 1.5\ M$ (Richter, 1935)

where E is in joules. Typically the energy of associated tsunamis is 1% to 10% of this total available earthquake energy (Iida, 1963). Tsunami run up can be estimated from tsunami magnitude according to empirical formulae like:

$M_T = 3.33 \log h$ (Bolt et al., 1975) where h is in metres.

The Strait of Georgia - Puget Sound region is a fore-arc basin related to modern subduction off the west coast. Accumulations of sedimentary strata in this tectonic depression and faults which cut these strata indicate that there has been intermittent subsidence in the region since Cretaceous time (Muller, 1977). The youngest and shallowest of the faults cut a variety of unconsolidated Quaternary sediments and indicate the persistence of shallow crustal motions into Recent time.

Historically (< 100 years) the pattern of seismicity in the region is that most earthquakes, both large and small are confined to the Georgia-Puget depression. During the period of observation, the shallow earthquakes have all been small (<4.5, Richter), and are not obviously correlated with any mapped fault system, and the larger earthquakes have all been deeper, the result being that none have caused significant seafloor displacements capable of triggering tsunamis. Seismic monitoring and an understanding of the tectonic setting suggest that one can expect earthquakes in the downgoing plate, or overlying crust, as large as 6.5 to 7.0 as frequently as every 10 to 20 years, while major subduction zone quakes ($\geq$ M 8.0) are expected only once or twice in a millenia (Rogers, 1982). When the former type earthquakes are shallow enough to be tsunamigenic, either by moving the bedrock of the seafloor or by causing major slumping, one could predict tsunamis with energies on the order of 2×10^{12} to 6×10^{14} joules and run ups of 0.5 to 2.0 m.

The latter type of major subduction zone earthquake should be expected to engender a tsunami on the order of 10^{16}J with run ups greater than 4 m; which would be on par with the largest recorded tsunamis in the world. Taking both large earthquakes and seismically triggered slumps into account, one would expect internally generated tsunamis to occur in Georgia Strait at least as frequently as every few hundred years. Slumps of sedimentological and anthropogenic origin would make the tsunami recurrence more frequent.

FIG. 1

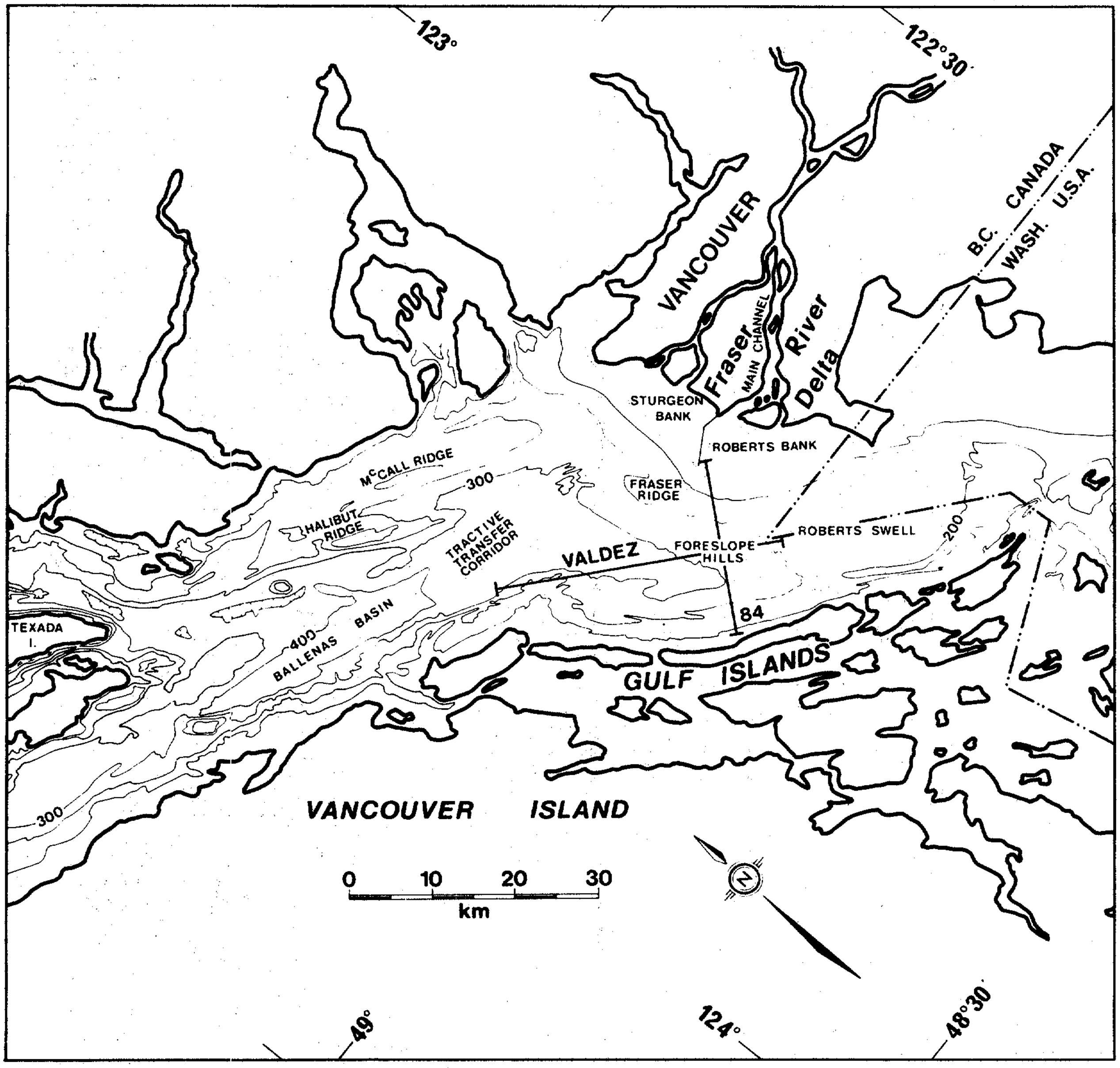


Fig. 1: Georgia Strait showing coastline and 100 m bathymetric contours. Prominent bathymetric features are labelled after Hamilton and Luternauer (1983a). Locations are given for seismic and sounding profiles of Figure 4a, b and c (Valdez and 84).

The Foreslope Hills

The Strait of Georgia, located between the B.C. mainland and Vancouver Island (Fig. 1), is a tectonic and erosional basin which is underlain by a thick sedimentary succession, reflecting a history of intermittent subsidence since Cretaceous time (Muller, 1977). This body of water is intermediate in depth between typical continental shelves and open ocean basins (up to 400 m, Fig. 2) and although it is effectively sheltered from even the largest tsunamis generated in the Pacific Ocean, it may for a variety of geological reasons be prone to internally generated tsunamis. The foremost of these is the existence of the Fraser delta, which has many of the characteristics of active, unstable, high sedimentation rate (Mathews and Shepard, 1962, Luternauer, et al. 1983) river deltas elsewhere in the world, and the tectonic setting on the west coast of Canada in an area of high seismicity (Rogers, 1982).

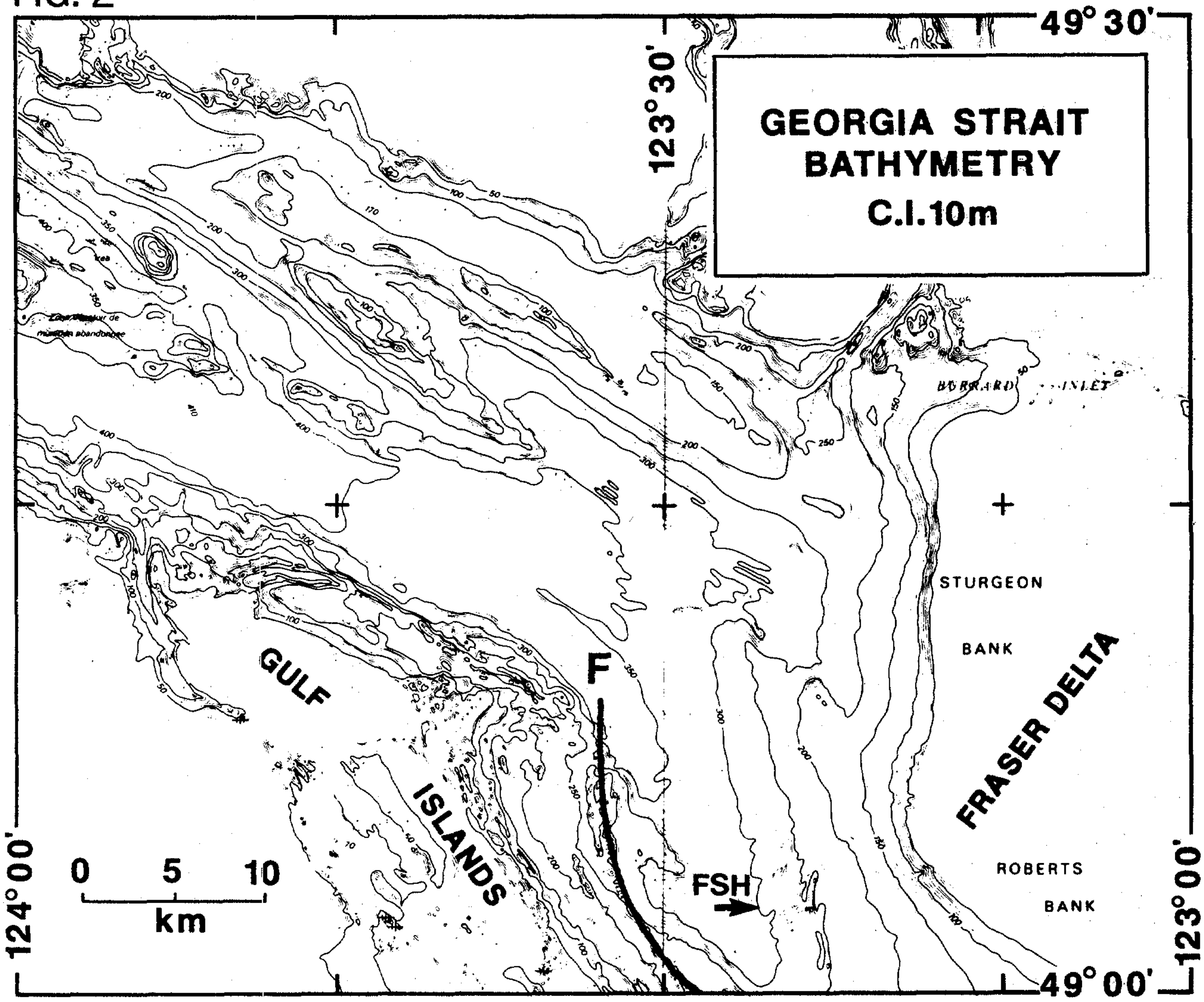
The modern morphology of the seafloor and coastlines of the Strait of Georgia originate from a long history of successive Quaternary glacial erosional and depositional cycles followed by a post glacial interval of sedimentation spanning the last 9,000 years. This latest interval of sedimentation is dominated in the central portion of Georgia Strait by rapid outbuilding of the delta of the Fraser River (Clague, et al. 1983). As an area of youthful unconsolidated sedimentary deposits there is a relatively high degree of seafloor instability including some large slumps, both in areas with oversteepened banks and in areas of high sedimentation and undercompaction on the modern delta (Hamilton and Luternauer, 1983a). The largest of the youthful slump deposits in Georgia Strait is the Foreslope Hills.

The Foreslope Hills (FSH) are located off the main channel of the Fraser River and opposite the Gulf Islands in low strength fine grained muds and silts (Pharo and Barnes) 1976 (Fig. 1). The bathymetric contours (Fig. 1) clearly show the FSH as an area of contorted topography in 230 to 330 m water depth on an otherwise unremarkable low angle (1/2° to 2°) submarine slope. Note also the Fraser Ridge, an all but buried Pleistocene bank (Tiffin, 1969) parallel with the others (McCall to Halibut) and extending in the subsurface below Roberts Bank to Point Roberts. This bank of glacial deposits provides a competent buttress in the subsurface architecture on the east while the bedrock ridges of the Gulf Islands provide a complimentary buttress on the west, with the FSH lying in between.

The FSH area comprises a series of ridges and troughs about 400 to 600 m from crest to crest (Fig. 3). Typical heights crest to trough are about 20 m. The trend of the ridges and troughs (N15°E) is slightly oblique to the local delta foreslope, (N15°W, Fig. 1, Fig. 3) and individual ridges extend as much as 4 km in length. Some of the ridges are continuous in the sub-surface and mappable with seismic methods. The thickness of the slump deposit ranges from 35 to 100 m in this region with extensions to the east and south beneath areas of more recent sediment accumulations. The total volume of the slump mass (which judging from other deposits, probably includes some reworked seafloor sediments in the path of the slump; Prior and Coleman, 1984; Prior et al., 1984) is somewhere between 5 and 14 km^3.

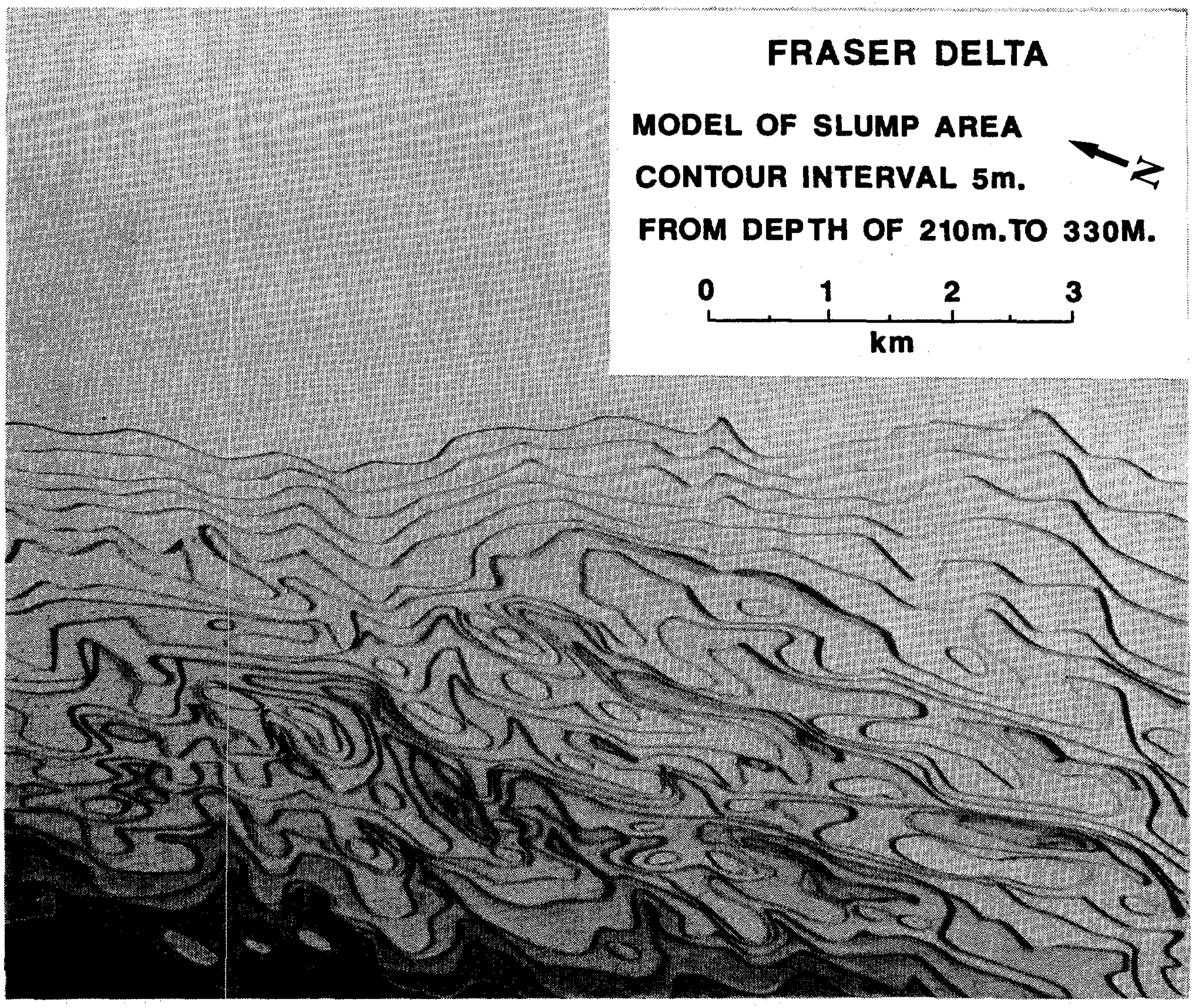
Fig. 2: Central Georgia Strait with 20 m bathymetric contours based on the Canadian Hydrographic service Natural Resource Maps 15792A and 15794A. The Foreslope Hills (FSH) are clearly visible as an area of contoured topography in 230 to 330 m water depth on the low angle (1/2° to 2°) submarine foreslope of the Fraser Delta. F denotes the Gulf Islands Fault Zone.

FIG. 2



Analyses of gravity cores taken from the FSH indicate both on the basis of faunal and textural evidence that the sediments of the FSH originate in much shallower water, specifically the crest of the subtidal platform of the Fraser delta, and that they have been translated into their present bathyal position (Hamilton and Luternauer, in prep.). The sediments of the FSH have internal deformations and rotations, compared to simple

Fig. 3: Physiographic model of the Foreslope Hills.



horizontal laminations in adjacent undisturbed areas. The sediments that comprise the FSH are remolded, having been partially dewatered and degassed during the slumping event so that they are now denser by (≥ 1.2X) and display greater undrained compressive strengths by (≥3X) than nearby seafloor muds and silts in comparable water depths. The simplest interpretation is that these sediments have been remolded and transported to their present location by a single slumping event.

In attempting to determine the age of the Foreslope Hills there are a few constraints. The FSH are clearly visible on the original sounding rolls of the Canadian Hydrographic Service Survey of Georgia Strait performed in 1938 and were recognized as a slump feature on a later geological survey of the region in 1959, published by Mathews and Shepard (1962). Mayers (1968) and Tiffin et al. (1971) estimated that they were no older than a few hundred years on the basis of their position in the sediment column and the regional sedimentation rate (which is poorly known). Although cores of the hills and sediment ponds between them have been found to contain both shell and wood material, the individual fragments are too small for conventional ^{14}C radiometric dating.

Repeated surveys of the FSH (1938, 1959, 1968 and 1983) permit local accurate estimates of sedimentation rates. Repeated detailed bathymetric surveys of the FSH area in 1968 and 1983 match ridge positions and depths to within $\pm$ 1 m attesting to the accuracy of the surveys and demonstrating no ongoing (creep type) deformation in the area. However the troughs show considerable infilling in this 15 year span and the foreslope shows considerable sedimentation and encroachment from the east. Sediment accumulations ranging from 1 to 13 m (E to W respectively) indicate local sedimentation rates from 7 to 87 cm yr^{-1} across the area. These rates are extremely high and essentially imply that the FSH will be buried by trough infilling and foreslope advance within a few tens of years, if these rates of accumulation persist. Extrapolating these rates backwards for the sediment accumulated in the troughs implies them to be not much older than about 100 years.

The structure and stratigraphy of the FSH slump area are presented in Figures 4a, b and c. Continuous seismic profiles (Fig. 4a and 4b) were acquired with a single channel hydrophone array and a small (16.4 cc, 1.38×10^7 Pa) dual airgun source. A 3.5 kHz high resolution near subbottom profile (Fig. 4c) was acquired at the same time as CSP 4b. The location of these lines is given in Figure 1. The vertical scale on 4a and 4b is given in 2-way reflection time and reduced to an equivalent water depth (1500 m/s, water velocity). Both seismic profiles show a similar stratigraphic sequence interpreted to be Tertiary and Cretaceous bedrock overlain by Pleistocene tills and diamictons overlain by Lower Post Glacial (LPG) and finally Upper Post Glacial (UPG) sediments (Hamilton and Luternauer, 1983a, b). The FSH extend from the modern delta foreslope, above the buried "Fraser Ridge" (Pleistocene) to the foredeep above and against a buttress of (up thrown?) basement rock. All of the postglacial sediments, in this natural box bounded by more competent formations, show evidence of folding, faulting and deformation. Some seismic profiles, like Figure 4a, show offsets of both UPG and LPG sedimentary layers, suggesting an underlying tectonic cause for these large scale slumps. Note the disharmonic nature of the folds and structural complexities: peaks below troughs, synclines below anticlines, slope reversals (down west slope beneath down east slope), deformed layers beneath modern foreslope, etc. This subsurface structure can most easily be interpreted in terms of a succession of similar large magnitude slumps, extending for a few hundred meters down into the postglacial record.

On the 3.5 kHz record (Fig. 4c) the depth of acoustic penetration is more restricted but the near surface structure is better resolved. Upon examination, this profile displays all of the diagnostic morphological characteristics of major submarine slumps (Dingle, 1977) including: 1) an upslope zone of fissures and small slumps, 2) a slide plane scar, 3) a tensional depression at the upslope edge of the slumped mass and 4) the main mass of detached, rotated and slumped material. The $\leq$ 30 m of relief of the main ridges and troughs is clearly visible in the middle portion of the

FIG.4a

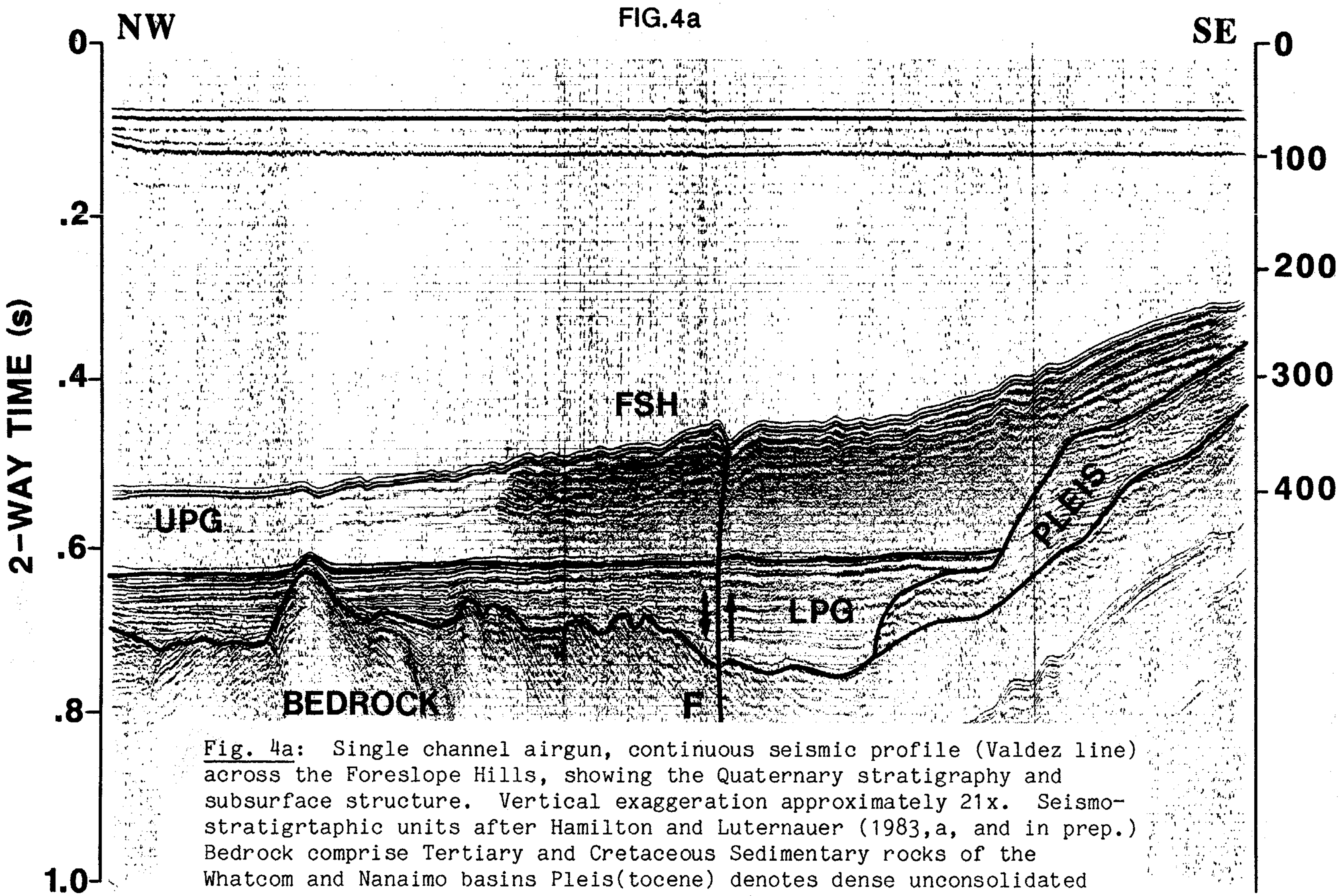
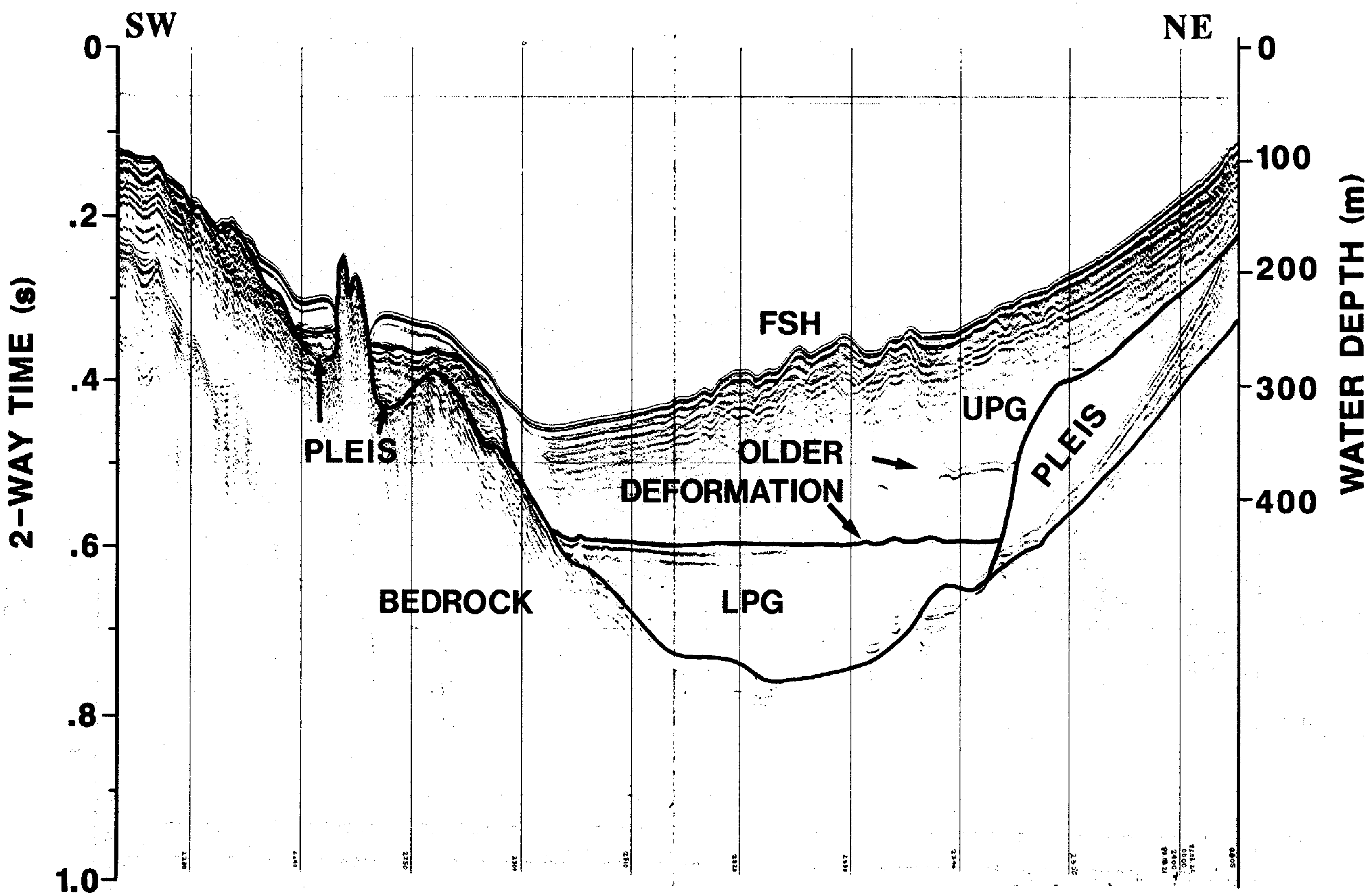


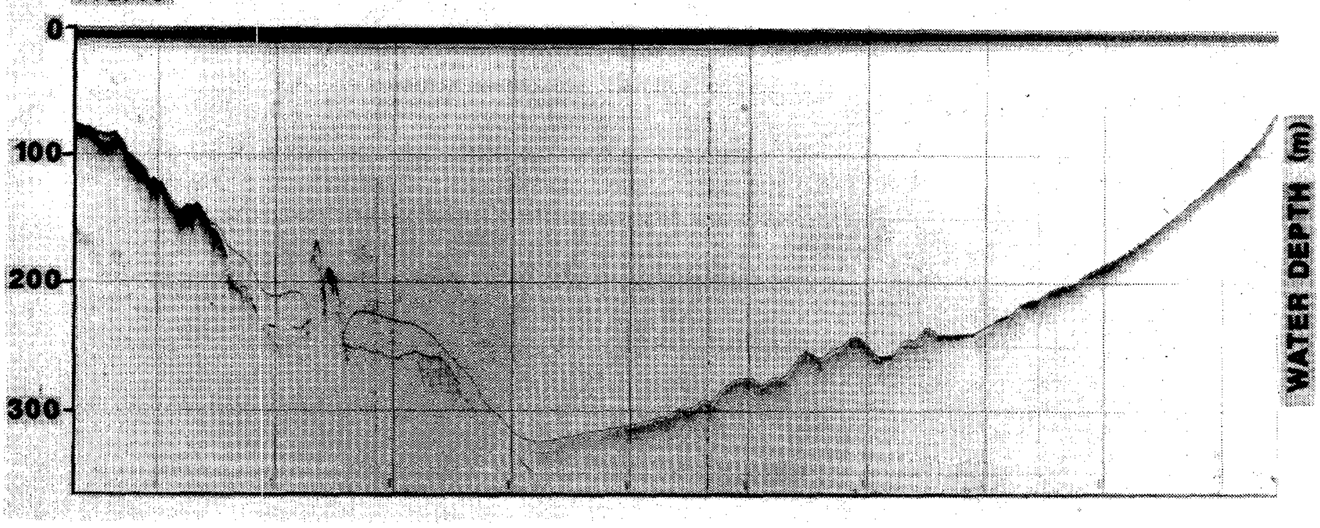
Fig. 4a: Single channel airgun, continuous seismic profile (Valdez line) across the Foreslope Hills, showing the Quaternary stratigraphy and subsurface structure. Vertical exaggeration approximately 21x. Seismo-stratigrtaphic units after Hamilton and Luternauer (1983,a, and in prep.) Bedrock comprise Tertiary and Cretaceous Sedimentary rocks of the Whatcom and Nanaimo basins Pleis(tocene) denotes dense unconsolidated sediments of Glacial age and origin. LPG (Lower Post Glacial) denotes accumulations of horizontally bedded unconsolidated sediments confined to the deepest portions of the Pleistocene erosional basins. UPG (Upper Post Glacial) denotes the youngest unconsolidated sediments, here these are dominantly deltaic facies of the foreslope and foredeep. F denotes a possible recent fault.

FIG.4b



4b: CSP line 84 across the Foreslope Hills.

FIG.4c



4c: Sounding profile (3.5 kHz) collected at the same time as CSP line 84. Vertical exaggeration approximately 24x. Note slump deposit, tensional depression and upslope region of small scale slumps and fissures.

record. Note the small scale ~ 3 m slumps further up the slope in the range from 50 to 170 m water depth and the flat lying sand and silt terraces which have been deposited since the major slump event. This acoustic picture of the infilling of the tensional depression and the outbuilding of sand and silt turbidite terraces on the advancing foreslope explains and confirms the findings of the repeated bathymetric surveys and the high sedimentation rates in the area.

Origin and Mechanism for the Formation of the Foreslope Hills

The geological community has been aware of the existence of the Foreslope Hills since the publication of Mathews and Shepard's sounding profiles (1962). From a few widely spaced bathymetric profiles alone, one would be hard pressed to demonstrate that the FSH were a product of deltaic sedimentary processes rather than some relict and partially buried pre-existing topography. Terzaghi (1962) inferred the FSH to have originated as a series of individual slumps representing mass transport for sedimentological reasons from the subtidal platform to the foredeep. However, the continuity of the ridges and troughs, evident on the bathymetric model (Fig. 3) and the continuity and character of the subsurface structure, evident in the closely spaced seismic data base of Hamilton and Luternauer (1983a), favors a single large slumping event. This in fact is the position favoured by most investigators: Mathews and Shepard (1962), Mayers (1968), Tiffin et al. (1971).

The origin of the FSH is explained by a four stage model involving loading, failure, slump and sedimentation (Fig. 5; Hamilton and Luternauer, in prep.). To understand the loading required to initiate failure, one must realize that in areas like deltas with high sedimentation rates, sediment accumulation is too rapid to permit steady state compaction, dewatering, degassification and dissipation of high pore pressures (approaching geostatic) (Prior and Coleman, 1982). On the Fraser delta where most of the sediment load arrives in one annual pulse associated with the spring freshet,

FIG. 5

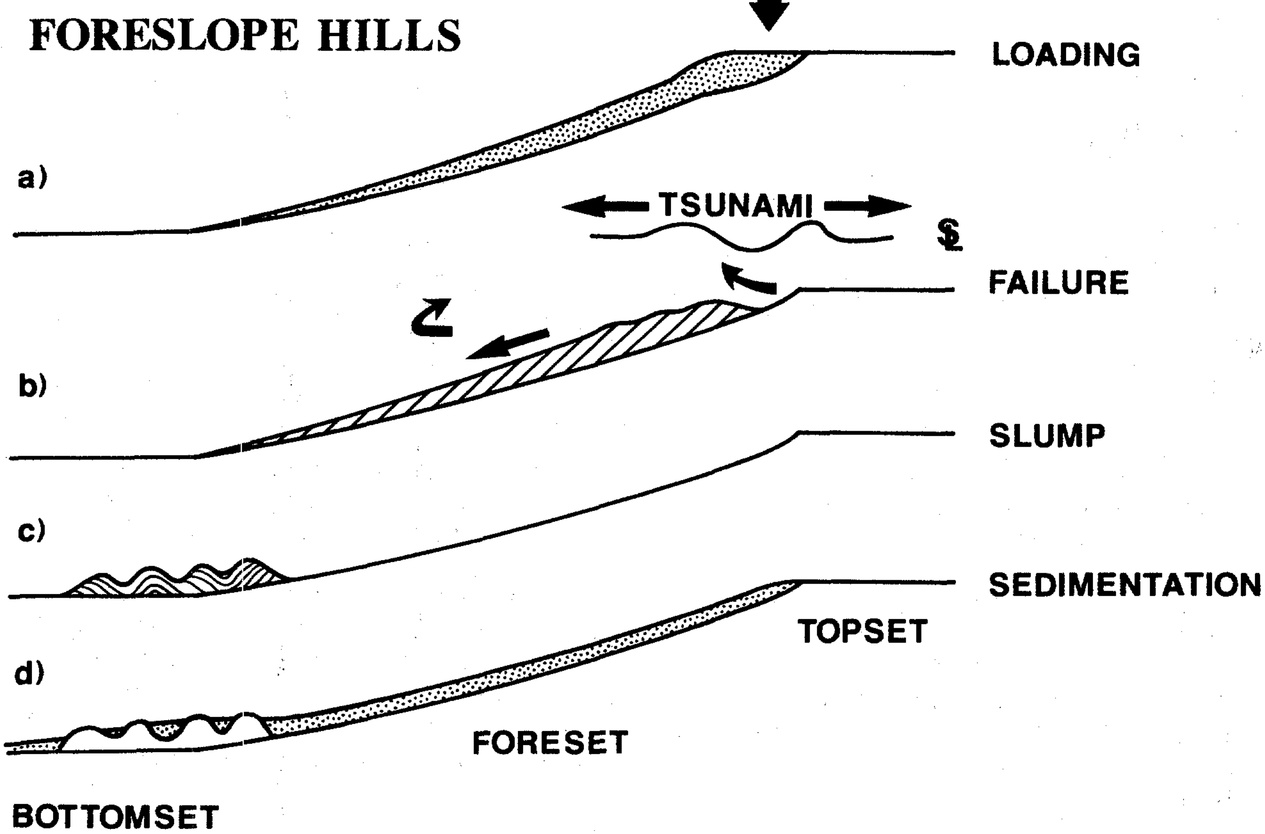


Fig. 5: A mechanism for the formation and evolution of the Foreslope Hills in 4 stages after Hamilton and Luternauer, (in prep.).

(Luternauer et al., 1983), it is easy to visualize loading induced failure. External causes all relate to increased shearing stress with possible contributions from seismic accelerational loading (shaking), sediment loading (burden of spring freshet, dredging/dumping/ landfill) water loading (high runoff and large tidal ranges, manmade barriers affecting runoff and tidal circulation) and cyclic wave loading such as occurs in shallow water during intense storms. All of these potential triggering mechanisms are present at least intermittently in the region of the Fraser delta.

The actual failure and liquefaction of the sediments results in the downslope transport of the slumped mass, and any substrates which it picks up from the overridden seafloor and incorporates on its downslope passage. The downslope transport of this semi-coherent mass probably took place over a period of a few minutes to a few hours, as suggested by the cohesive nature of the sediments and their internal structures (Hamilton and Luternauer, in prep.). According to the model, the transfer of energy and momentum from the moving sediments to the overlying water column could generate a tsunami. While examples of tsunamis related to seismic events are more common, slump (non-seismic) induced tsunamis also exist, such as the Greek tsunami of February 7, 1963 in the western Gulf of Corinth (Papadopoulos and Chalkis, 1984). Such a disturbance radiating from the vicinity of Sand Heads in the central Strait of Georgia would easily propagate both up the Strait (NW) to affect urban centres like Nanaimo, Parksville, Comox, Powell River and Campbell River, by reflection into Burrard Inlet and down the Strait (SE) to affect the American San Juan Islands and Puget Sound. The extent of effects on the tidal flats of Roberts and Sturgeon banks, on the low lying areas of Richmond and Delta and up the main arm of the Fraser would depend on whether or not there was a high tide at the time of failure.

The immediate product of the failure would have been the compacted, deformed and dewatered slump deposit at the foot of the scarred delta foreslope, either as a single mass or one principal mass with smaller isolated outrunner blocks.[1] Associated with the Foreslope Hills is an isolated southerly outlier smaller than 2 km across centred near 49°0.3' 123°23.5' which is either an outrunner block or a separate slump. In the former case the disaggregation of the slumping mass into several discreet blocks, would have caused a more complex interaction with, and momentum transfer to, the overlying water column than a simple planar push or a moving wedge. This would considerably complicate the modelling of any wave effects.

Following the formation of the slump, the upslope (source) region will have been fissured with tensional cracks and possibly oversteepened. The slide plane scale will also have exposed older sedimentary layers to erosion. The ridge crests of the deformed slump mass will have provided a new impediment to bottom current flow in the area, localizing erosion. Similar effects to concentrate erosion and reworking, and concentrate sedimentation in the upslope depression and intervening troughs have been documented for a young mass failure deposit off of Canterbury N.Z. by Carter and Carter (1985)[2]. On the Fraser delta, the subsequent resumption of normal sedimentation advances the foreslope and infills topography at and upslope from the slump deposit, preparing the foreslope for another cycle of failure.

The reflection seismic evidence for deeper deformation beneath the FSH needs to be assessed before any final interpretation of their genesis can be

[1]This latter morphology is common with slumps in deltaic environments (Postma, 1984) such as the Mississippi (Prior and Coleman, 1982) and the delta front at the head of Kitimat fjord (Prior et al., 1984). In the case of Kitimat, B.C. the delta front, slump and slope failure of 1975 was caused by marine engineering works (dyking) which led to a pore pressure build up and sediment liquefaction (by impeding the return flow of foreshore water) during the drawdown from ebbing tide. That this non-seismic slump spawned a tsunami is a matter of record (Prior et al., 1984; Murty, 1979, Campbell and Skormer, 1975).

[2]From their work it is apparent that mass failure topography on the seafloor exerts a strong modifying influence on the seafloor hydraulic regime. The high sedimentation rates determined for the FSH can be easily explained and the youthfulness of this feature further appreciated with the insights gained through a comparison with the Canterbury, ridge and gully feature.

made. Hamilton and Luternauer (in prep.) have worked out a simple sequential model to explain the structures, (Fig. 6a and 6b). The deformation of the LPG sediments is attributed to a Pleistocene reactivation of listric normal basement faults (possibly due to postglacial rebound). The style of deformation which spans the deposition of the LPG sediments is that of disharmonic folds which build in amplitude away from the listric fault (strain is taken up internally rather than as slip). This style of thin skinned deformation is fairly common in soft sediment deposits and has been well documented for the case of growth faulting of Plio-Pleistocene silty/clayey sediments in the eastern Mediterranean (Garfunkel, 1984).

FIG.6a

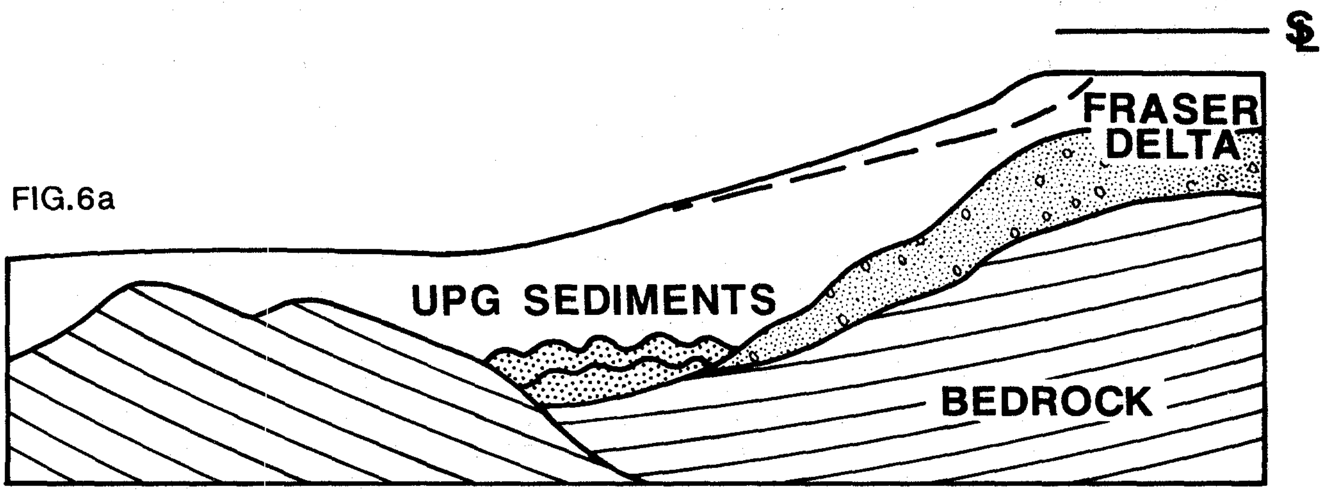


FIG.6b

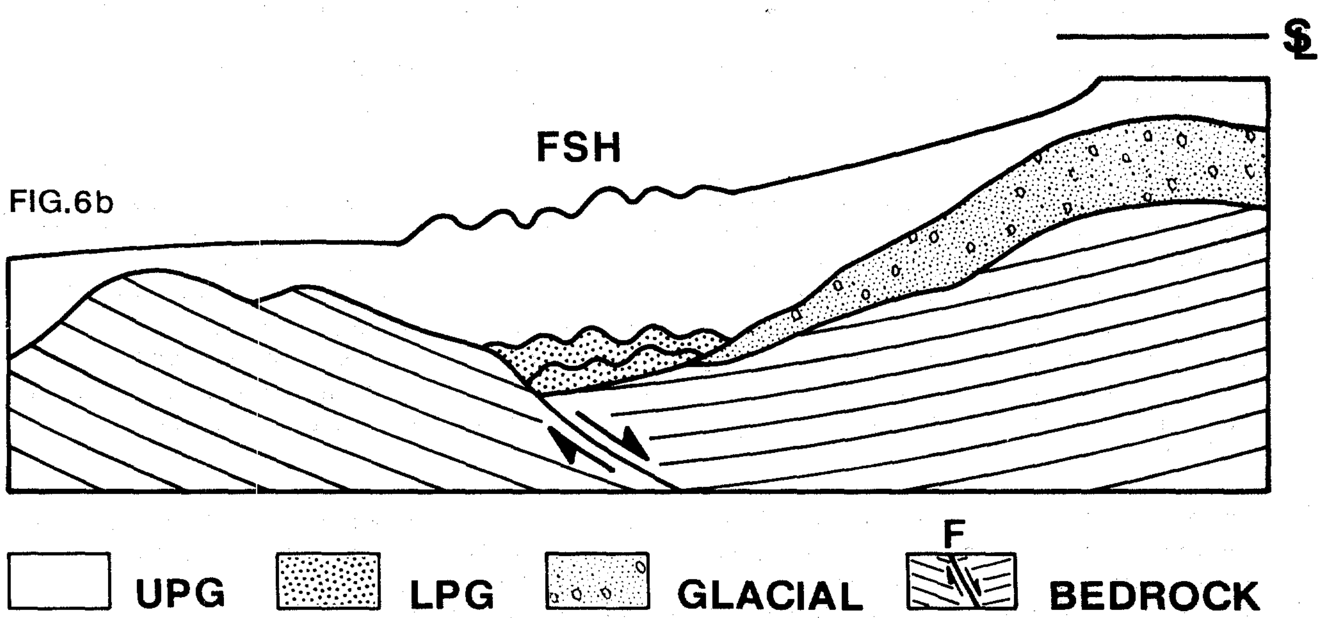


Fig. 6: Geological cartoon depicting the development of the Quaternary structures in the vicinity of the Fraser Delta after Hamilton and Luternauer, (in prep.).

In the case of the FSH, the possibility of renewed fault motion in the underlying basement as a trigger for this latest slumping event cannot be ruled out. Some seismic profiles show offsets of both UPG and LPG sedimentary layers, suggesting an underlying tectonic cause for these large scale slumps.

Two other possibilities which should also be considered are gravitational instability of the sedimentary section and standing wave type oscillation. In the former case, denser sediments are laid down on top of less dense sediments, resulting in unstable stratification and flowage, or dispersion of the underlying layers. If the LPG sediments are less dense, more hydrous or more gas laden (than the overlying UPG) this type of instability could arise; effectively explaining the underlying deformations and providing a trigger, albeit a gradual one for the catastrophic FSH failure. In the latter case it was previously noted that in this region of the Strait of Georgia the unconsolidated low strength UPG and LPG sediments are bounded by more competent formations. In the event of a major earthquake the shallow Quaternary and Holocene deposits would probably act as a wave guide to focus surface wave oscillations. Zones of maximum ground motion amplitude (and destructive periods of oscillation) can frequently be related to the topography of unconsolidated Quaternary sediments, essentially by solving for the oscillational stable harmonics of the lens of surficial sediments. Such was certainly the case for the regions of maximum damage to Mexico City in the earthquake of 19 September 1985. For the "pro delta" and foreslope sediments in Georgia Strait it is conceivable that sympathetic nodes of vibration be set up in this vicinity in times of major earthquake ground motions. This could explain the repeated deformations in this region in spite of the obvious progradation of the delta, changes in bottom slopes, sediment types and distance from likely failure sources.

Energy and Wave Height Estimates

Before embarking on speculative modeling calculations, which by their nature are fraught with untestable assumptions, it would be useful to simply compare the FSH to other tsunamigenic slumps. This method at least should provide an empirical estimate of the size of energies and wave heights to be expected. In 1975, foreshore development at the head of Kitimat Arm caused a slump which engendered a tsunami. The slump deposit associated with the 1975 event alone was on the order of 3.3×10^9 m^3. This volume includes perhaps equal parts of slumped material and reworked seafloor sediments entrained by the slump (Prior et al., 1984). The slump originated across a front approximately 1 km wide and travelled approximately 5 km down the fiord to a final water depth of 210 m. The majority of this path was over ~1/2° slopes. The associated water wave amplitude, possibly as large as 8m and water disturbance for about 1 hour (Murty, 1979), was responsible for the destruction of boats, docks and other property. On 16 October 1979 a mass failure occurred on the submarine portion of the Var River delta (Genneseaux et al., 1980). The likely cause was the loading of the subtidal platform by a manmade 300 m long bank of dredge spoils, although there is some indication that the entire delta is inactive and has been undergoing erosion through repeated mass failure in recent times. In any event, the entire artificial bank disappeared in a matter of minutes and over the next 6 hours submarine telephone cables were severed by associated turbidite flows. In this

Baiê des Anges (Cote d'Azur) disturbance the initial failure, probably about 3×10^5 m^3, caused a wave of several meters resulting in extensive damage along 100 km of coastline (Pautot, 1981). However the disturbance including wave oscillations and continued retrogressive slope failure over a 2 km width of delta front continued for several hours, either driven by the initial failure or controlled by some feedback mode of wave oscillation - cyclic wave loading - continued failure - renewed wave oscillation etc. The total amount of mass movement in the 1979 event was probably closer to 2.1×10^8 m^3. In the case of the mass failure off of the Var River all of the slumped material was carried out by turbidite or other mass flows down a submarine canyon over a distance greater than 90 km. The successive breaking times of the submarine cables allow velocities to be estimated, for the downwards slowing flows, of about 22.8 km/hr and 4.8 km/hr (several tens to a few hundreds of meters per minute). The mass failures on the steepest part of the slope (~4°) between 200 m and 2 km of water depth were undoubtedly somewhat faster than over the lower slopes (1/2°) of the abyssal plain.

The list of comparisons could easily be lengthened to include: the slump and tsunami of February 7, 1963 in the western Gulf of Corinth (Papadopolulos and Chalkis, 1984) and the tsunamis of Sanriku/Sugami Bay, Japan, 1933 and Valdez, Alaska, U.S.A., 1964 (Bolt et al., 1975). However the basic points to be made by comparison with the FSH are that slumps and mass failures lasting up to several hours in duration and less than 1% the total volume of the FSH event are capable of spawning tsunamis of several meters in height, and that the 3° to 1/2° slopes leading down to the FSH are typical of the regional slopes in these type of failure situations.

Taking total volume estimates of the FSH deposit and the path to the 200 to 300 m water depth into account, one can estimate the potential energy change associated with the 11 km front of the youngest FSH event at about 1.16×10^{16} joules. The estimated potential energy changes for the Kitimat, 1975 and Var, 1979 slumps are respectively 8.8×10^{13} joules and 3.64×10^{15} joules, which emphasizes the potential magnitude of the tsunami which could have been associated with the FSH event. Recall also the energy calculations presented in the introduction for major earthquakes and associated tsunamis, and it is readily apparent that whether one calls on an infrequent large earthquake or simply examines the potential energy change the slump has undergone, that the energy to generate a large tsunami is there. Performing the analyses of Striem and Miloh 1975, assuming only a small proportion of the total available energy is efficiently transfered to the water column, allows energy estimates to be made for a solitary wave of between 10^9 and 10^{10} joules. Repeating the type of analysis which Murty (1979) performed for the 1975 slide in Kitimat Arm one can arrive at minimized solitary wave height estimates greater than 150 m. Without being overly critical of the number of assumptions inherent in these model calculations or even accepting the concentration of this amount of energy in a solitary wave, it is still quite apparent that a major water column disturbance is implied. The most likely analogy is that of the 1979 Var River delta event where a major tsunami of several meters height was followed by a period of several hours of wave oscillation and disturbance related to the dynamics of the mass flow. To fully describe the expected wave disturbance one would need a dynamic model to assess the momentum transfer from the sediment to the water column. This would necessarily take into account not only the geometry and mechanics of the mass failure but the actual kinetics of the failure process.

While the FSH is the largest youthful slump deposit in Georgia Strait there are others which are also noteworthy and were probably capable of generated tsunamis in Georgia Strait. A prehistoric slide dated to be younger than post 644AD[1], occurred between Orcas and Sucia Islands in the American San Juans (Hamilton and Luternauer, 1983b, and in prep.). As an area of high relief, negligible postglacial sedimentation, basement structure and current high seismicity the triggering cause for this slide was most probably a local earthquake. The water depths and slide volumes were comparable to Kitimat Arm and probably could have generated waves of similar magnitude.

There has also been another slide off of Sands Heads in the main distributary channel of the Fraser River during July 1985. The slide represented a removal of a volume 37 m deep by 450 m upchannel by 280 m NS from the shallow subtidal source area in a region of active dredging. The cause was undoubtedly increased loading by sediment accumulation from the spring freshet coupled with the drawdown effect of the large tidal range during the first week of July. (Also there were no local earthquakes of any consequence on the Strait during that period). This particular failure did not register on the tide gauges at Pt. Atkinson or Steveston, although neither of those stations was well placed to record a small magnitude event originating at Sand Heads. Whether this failure was below the minimum threshold to cause a wave disturbance, or more probably it simply occurred as a slow retrogressive grain flow over many hours, the indication of modern ongoing mass failure activity near the source region of the FSH event is unequivocal. This is particularly important because a much smaller initial failure was apparently capable of triggering the massive 1979 disturbance on the Var delta.

Conclusions

From the geological, geophysical and bathymetric evidence it is apparent that the physiographic feature known as the Foreslope Hills was formed by a single submarine mass failure. The absolute age of this feature is older than 47 years but probably younger than a few hundred. While local seismicity, related to the tectonic regime, is a possible trigger for the failure, the most likely cause is the Fraser River itself, with its yearly spring freshets and continual addition of new unconsolidated materials. Regardless of how one chooses to model or estimate the accompanying tsunami effect it is evident that in a confined basin like Georgia Strait with extensive areas of low relief, high population density and foreslope development (Ferry Terminal, Coalport, Vancouver International Airport etc.) the effects of such a wave could be devastating. The geological evidence for the recurrence of slumps like the FSH needs to be seriously considered for

[1] G.S.C. radiocarbon date, shell fragment (Macoma sp.). This dates the age of the sediments involved in the slump, the slumping may have been more recent.

both the placement of major engineering projects on the delta (such as the proposed Vancouver Island Natural Gas Pipeline) and in appraising the future tsunami risk for Georgia Strait.

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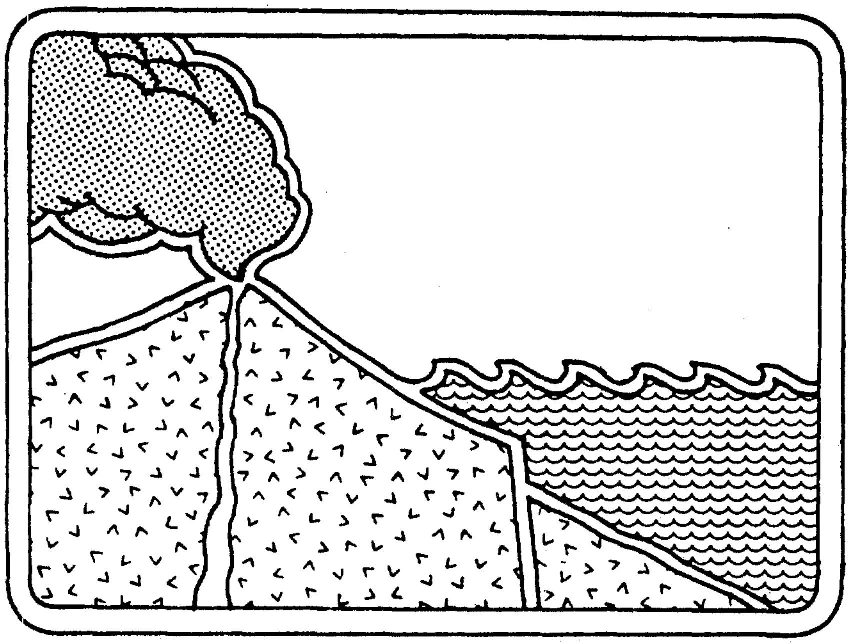

THE NOVEL SEISMIC SOURCE MECHANISM OF THE 7 MAY 1986 TSUNAMI

Wm. Mansfield Adams*

ABSTRACT

Immediately after the 7 May 1986 earthquake, all seismologists speculated that a very large tsunami had been generated: this expectation was conditioned by knowledge of the location of the epicenter at 51.3 N, 175.0 W. An early report of wave action near the source region showed only a small tsunami had been generated. According to the operating rules, a tsunami warning was issued by the Pacific Tsunami Warning Center for several areas–in particular, for Hawaii. Early observations indicated an unexpectedly short period slightly greater than ten minutes. In Hawaii, no operationally significant oscillations were observed. On the basis of these meager facts, the source can be anticipated to be of novel form: I propose a "sliver" which is defined to be an active area of a fault with a relatively short length intersecting the surface of the earth and a relatively long length along a line plunging downdip or with a strong dip component. This earthquake-tsunami event indicates the necessity of rapidly obtaining meaningful data in real time if a warning system is to improve the cooperation of the public.

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I. BACKGROUND

On 07 May 1986 at 2247Z an earthquake occurred near Adak, in the central Aleutian Archipelago. See Figure 1. A preliminary determination of Richter magnitude was 7.7 (National Earthquake Information Center (NEIC), Pacific Tsunami Warning Center (PTWC), and Alaska Tsunami Warning Center (ATWC)) on the basis of surface waves.

We are now in a position to use several months of hindsight. At the time a tsunami warning was issued, evacuation from those zones designated "inundation zones" in the Hawaii telephone directory was undertaken, and when only small waves occurred, the warning was cancelled. Now we can discuss what type of analysis might have been done and what might be done during the next event if more personnel, improved communication capabilities with a computer-based data set of the tsunami literature, and better instrumentation in the source and terminal areas become available. Some detailed recommendations concerning such instrumentation have been published (Curtis and Adams, 1985).

II. SEISMOLOGICAL ANALYSIS

Explicitly mentioned in the final PTWC log is that the period of the tsunami oscillations at Adak, only about 150 kilometers from the epicenter and having the closest tide gauge, was reported to be 15 minutes. This report is very significant for its seismological implications, as will be shown in the following development.

A novel method for quickly estimating the seismic moment has been reported recently by Talandier and Reymond (August, 1986: made available to the author as a preprint). The motivation is that the seismic moment is the most significant value for characterizing the physical parameters of the earthquake source, as argued by Kanamori (1977). Knowing the seismic moment and some other physical parameters may permit estimation of whether or not the earthquake is tsunamigenic.

The seismic moment is defined as:

$$Mo = \mu\ D\ S$$

where μ is rigidity, D is the average displacement over the fault, and S is the area of the fault that ruptures. In the warning-system environment, we can take the value for the rigidity of the source region to be that obtained from study of previous earthquakes in that region. Here we use $8 \times 10E11$ dynes for the Adak region.

The value of D, the average displacement over the rupturing surface of the faulting, can be taken from the maximum wave height in the source area. If appropriate, any observed value can be backwards extrapolated using Green's Law or some equivalent. For the 7 May 1986 event, we extrapolate the Adak observation of 175 cm back to deep water and obtain 50 cm (peak-to-trough).

We choose to consider the surface S to be a simple rectangle having width W and length L. Usually the faulting is found to be dominantly parallel to the surface of the earth and extending downward a distance that is small compared to the length. However, in the present case, the wave period of the water motion was noted to be 13 minutes at Midway (Final log: 0255). For an average water depth in the source area near Hawley Ridge of 2000 meters, the velocity of propagation is about 140 meter/sec. After Sharpe (1944) we take the value of the diameter of the source region, d, to be about two sevenths of the distance traveled in a ten-minute period, say 36 kilometers. This scales directly with the period; so for a fifteen-minute period, the estimate of the source diameter, assuming a circular source, would be 54 kilometers. See Figure 2. Thus, the problem narrows down to determining the length L. One approach to doing this is to find a good estimator of Mo, such as has been devised by Talandier and Reymond (1986).

The conventional Ms magnitude scale saturates for large earthquakes, due to the definition in terms of amplitudes of surface waves measured on conventional seismographs. Talandier and Reymond (1986) report the development of a broad-band system with response flat to 300 seconds. Using data of the vertical component of the long-period Rayleigh waves measured on such a broad-band system, they defined a new magnitude scale, Mm. Adjustments for geometrical spreading and anelastic absorption are incorporated. They take the maximum value observed over several oscillations (this choice of the maximum may be unfortunate, as an extreme usually has an undesirably large variance (Gumbel, 1967)). They constrain the slope of the relation between Mm and Mo to be the theoretical value of one. The simplified form resulting is:

$$Mm = \log Mo - 20 \text{ (dyne} - \text{cm)}$$

Their observed value of Mm for the 7 May 1986 is 8.0. Thus, we find $Mo = 10E28$ dyne-cm. From the equation defining Mo, letting $S = W \times L$, we can solve for L. This is found to be approximately 600 km. The faulting that caused the 7 May 1986 tsunami may be very unusual. The diagram of Figure 2 shows the geometry schematically, but not to scale. The distance W represents the trace of the rupturing at the surface: the value of length L is the distance downdip (rather than parallel to the earth's surface, as is customary). Because of this unusual geometry, the faulting will be specially indicated as a "sliver" fault. The ratio of L to W may be termed the "aspect ratio". The dip of the thrust plane, for purposes of operating a warning system, may be taken from previous fault-plane solutions in the source region. Billington, Engdahl and Price (1981) reported on focal mechanisms in this region during an analysis of the 4 November 1977 earthquake that occurred at 51.659 N, 175.952 W at a depth of 33 kilometers and with Ms of 6.7.

In hindsight, it is now apparent that a tsunami warning was issued when the water waves were not of a practically significant amplitude. Here the term "practically significant" means, following the game-theoretic terminology introduced by Adams (1966), that the societal cost of responding to a tsunami warning is greater than the cost to society of not responding. Indeed, this provides an objective procedure for judging the performance of any type of warning system or forecasting technique. More of the scientific literature, especially that relating amplitudes of the water waves in the source region to the amplitudes at the terminal areas, should have been available in real time to the operators of the warning system. This indicates the possible benefits of a computerized data base.

Because the geological time scale is not commensurate with man's lifetime, exceptional situations are not expected, only the generic. However, by improving the instrumentation for measuring extremely large water waves in and near the source regions and transmitting that information to the appropriate analysis center and by extending the frequency response of seismographs, as has been done by Talandier and his associates, the anomalous events may be discerned.

III. MEASUREMENTS IN THE TERMINAL AREAS

Midway Islands

Obtaining meaningful data on water level has been recognized as a difficult task. Some approaches have been described by Van Dorn (1984). The alternative is to use what exists. For Midway, intensive efforts have been made to understand the often anomalous observations. Groves and Harvey (1967) attempted to find a non-linear transformation from deep water to shallow water, where Midway was taken as the deep water "input" and Hilo as the shallow water "output." Even with inclusion of the bilinear term in the multi-dimensional Taylor's expansion, meaningful emulators could not be derived. A reason for the difficulty they encounter will become apparent from the following discussion.

A diagram of the Midway Islands, located about 2200 km west-northwest of Honolulu, is shown in Figure 3. Midway Islands consist of two small islands-Sand Island and Eastern Island, a lagoon, and the atoll surrounding the lagoon. (There is no "Midway Island" per se.) Sand Island lies to the west and has a landing strip: Eastern Island lies to the east. The harbor is created from the east end of the Sand Island and contains a protected area at its northeast corner. The tide gauge is within this protected area. The value reported in the log of Table A-1 for Midway is of the same order as the values reported for Adak in the source region. As geometrical spreading would be expected to produce oscillations significantly smaller than in the source region, this anomaly has been considered. The recorded trace is reproduced in Figure 4, courtesy of NOAA.

Note from Figure 3 that the lagoon lies north of the two principal islands, that the lagoon is essentially open to the northwest, and that much dredging has been done to create (1) seaplane lanes, (2) a taxiing basin, and (3) a ship channel between the islands from the Pacific Ocean to the protected docks. The tsunami energy enters the lagoon through the open northwest sector and sets up resonant oscillations within the lagoon, for example, between the northeast ledge of the 120-degree seaplane lane and the western side of Welles Harbor. These are resonant about twelve minutes. Such energy would then diffuse into the harbor via the ship channel. This mechanism, however, would produce a slow buildup of energy in the harbor, with successive waves being larger. In fact, Figure 4 shows that the energy of the tsunami appears abruptly in the harbor. This is explained straightforwardly as the amplification of the wavefronts occurring on the far side of an island. For Midway, wavefronts approaching from the north will wrap around the atoll, collide constructively near the mouth of the ship channel, and flood into the harbor, as observed. This energy will be complemented by that leaking in from the resonating lagoon. Perhaps numerical modeling of this situation can verify this proposed senario or suggest alternatives.

This strong response of Midway to the shorter period components of the tsunami energy is also apparent in the data studies by Groves and Harvey (1967): Figure 5 shows the records at Midway and Hilo for the 1957 and 1964 tsunamis. The Midway records show more high-frequency content. Thus their assumption that Midway represented a deep-water version of the tsunami is unfounded due to the local resonance effects.

Many mareograms for this tsunami have been made available by NOAA. The most characteristic feature is the unexpectedly short period of about 12 minutes for the dominant energy. As emphasized, this is important for crude analysis of the seismic characteristics of the source rupturing and for understanding the energy spectra observed on the monitoring instrumentation.

Water Level Records from Water Wells in Honolulu, Hawaii

The Board of Water Supply routinely monitors the water level of several wells in Honolulu on the island of Oahu in Hawaii (Chester Lao, personal communication). Two of these recordings are available at this time. The approximate locations are shown on Figure 6: one is at Beretania Monitor Well, the other at Kapalama Monitor Well. The recording for Beretania Monitor Well is reproduced as Figure 7. The oscillations of the tsunami are apparent and are essentially coincident with the ETA for Honolulu. The theory for tidally driven acquifers has been extensively developed; a significant delay would be expected. This does not seem to occur for this well. However, the region is known to contain caverns in the sediments (Adams, 1984). These have not been followed to the ocean but are of sizes (3 meters or more across and vertically) and extend (more than a half kilometer) such that a direct connection would be reasonably acceptable. The recording indicates that such a direct connection does exist.

The water-level record for the Kapalama Well is given as Figure 8: it is completely different from that of the Beretania Well. Now the oscillations are seen at the time of the earthquake–not at the time of the arrival of the tsunami. The vertical scale of this record is, for the original, one inch per tenth of a foot. The horizontal scale is, for the original, one inch equals one day: The fine lines equal four hours per unit. There are some glitches at "noon" on 7 May that could be either background noise or signal. From a calibration using the start time, shown at the extreme left, the large dark black oscillation bottoming just below 19.00 feet and passing above the top of the chart is actually occurring at about 2 o'clock p.m. These are attributed to compression and dilatation of the acquifer (analogous effects have been observed for some geysers). A description of this compressional effect has been given by Logel (1980).

Some earlier efforts at recording earthquakes in waterwells on Oahu was done by Rockne Johnson (Cox et al. 1963) but not reported in the open literature.

IV. SUMMARY

The data available in real time for the 7 May 1986 earthquake-tsunami event are reconsidered from the vantage of hindsight. A tsunami warning was issued for the State of Hawaii but no significant tsunami occurred. The threshold of the warning system should be at a level such that the tsunami that actually occurs will be visually observable at most coastal points; this will improve public cooperation with the warning system and evacuation monitoring operations.

The source rupture is found, by using results reported since the event, to have a novel geometry. Whereas the fault surface is usually long in the direction parallel to the surface of the earth and much shorter in the direction downdip, this faulting seems to have been about fifty kilometers along strike but about 600 kilometers downdip. (These figures are based on generic values for rigidity, etc., as the specific seismic data are not available at the time of writing.)

This study indicates that the Pacific Tsunami Warning System has inadequate resources to achieve its assigned tasks. Specifically, improved access to the published literature is required, such as might be provided by a computer data-base. Additional manpower is required and could pre-compute and pre-compute and pre-distribute expected arrival times for hypothetical tsunami sources about the rim of the Pacific Ocean. More characteristics of the fault might be determined in real time if standard communication procedures could be implemented and additional competent staff added to perform such analyses. Instrumentation in or near any source area should be capable of surviving, measuring, and reporting water-level amplitudes of more than 20 meters. The broad-band seismic instrumentation developed and reported by the French should allow real-time estimation of the seismic moment. "And one is well aware that the danger of a major tsunami covering the entire Pacific only exists for earthquakes the moments of which are superior to $5 \times 10E28$ dyne-cm." (Talandier and Raymond, 1986). Prediction of tsunami wave heights should be encouraged, because tsunami wave heights can be forecast.

ACKNOWLEDGMENTS

The author is beholden to George Curtis for having read several successive versions of this manuscript on several successive weekends. Hawaii Institute of Geophysics Contribution Number 1846.

COMMENT AFTER ACCEPTANCE

Since this paper was written, a special symposium on this earthquake-tsunami event has been held at the Western American Geophysical Meeting in San Francisco, California as session S11A: the abstracts appear on page 1081 ff. of EOS, Vol. 67, No. 44, Nov. 4, published by the American Geophysical Union. The fault length estimates vary from 125 to more than 200 kilometers. Estimates of the dip are about 20 degrees (which also helps explain the small tsunami, as only the vertical component of ocean bottom motion will contribute to production of the tsunami). The two major subevents appear to be located on either side of the epicenter. The twenty-four hour zone of aftershocks is about 200 kilometers. Whereas the subevent 100 kilometers to the west of the epicenter is at the western end of the aftershock zone, the subevent to the east truncates at about 60 kilometers from the epicenter, far short of the eastern end of the aftershock zone.

The seismic data on which those analyses are based should be telemetered to the PTWC and the source parameters determined in real-time in order to assist in estimating the possibility of a significant tsunami in Hawaii and elsewhere.

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APPENDIX

Table A-1, taken from a preliminary log provided by the PTWC Director, lists some of the PTWC activities prompted by this earthquake. Notice that it took less time to determine a preliminary epicenter location (51.3 N, 175.0 W) than it took for the seismic waves to travel from the epicenter to PTWC. A preliminary determination of Richter magnitude 7.8 was achieved eight minutes later. Note (2309Z) that the magnitude value from ATWC is 7.7 and is termed a "confirmation." This recognizes the variance of magnitude determinations.

Two telephone contacts with ATWC, at 2331Z and 2348Z relayed the water level information from Adak: see Figure 1 for location of Adak, northwest of the epicenter. The earlier one reports 38 cm "wave activity" and the latter reports "waves increased to 70 cm." Based on the available seismic and water-level data, a tsunami warning was issued by PTWC at 2351: this is less than one hour from the activation of the PTWC seismic alarm. Note that the water motion is not well described by the term "wave activity" which could be either peak-to-trough (or vice versa) or a deviation from some datum. Close to the source, the early oscillations will contain the maximum amplitudes, hence these water-levels could be accepted as definitive. This is not a typical situation. An epicenter may not be close to a tsunami monitoring station: also any significant tsunami will have such large water-level motion near the source area as to be off-scale for most tide or tsunami monitoring stations (Adams, 1966). This fact has prompted the Tsunami Commission of the International Union of Geodesy and Geophysics to recommend repeatedly that instrumentation suitable for very large water-level oscillations (of more than 60 feet) be developed and installed in the areas known to be tsunamigenic.

Of great import is the reconfirmation of magnitude, made at 0002Z of 08 May. The fact is not as important as the indication of the quality control apparent within the operations of the PTWC, a most commendable feature. Such verification steps have been singled out by Adams and Curtis (1986) as lacking in most warning systems for natural or man-made hazards.

Most instructive is the 0029Z discovery that the Shemya tide gauge is "out of operation." This is simply symptomatic of what must be expected in any warning system—not all the sensors are functioning properly; not all the communication channels are functioning properly; not all the analysis and interpretation operations function perfectly. The need for redundancy **AT EVERY STAGE** is apparent.

The report from USSR (0340Z) that "no tsunami had been observed" must be understood in proper context. During a conversation by this author with a Russian tsunamist, much confusion arose about whether or not a particular earthquake had produced a tsunami. The confusion was finally resolved by discovering that the Russians (at least at that time) did not consider that a tsunami had occurred unless a water-level oscillation of more than TWO METERS deviation is observed.

TABLE A-1: ITEMS ABSTRACTED FROM THE PRELIMINARY DRAFT LOG OF PACIFIC TSUNAMI WARNING CENTER (PTWC) FOR THE 7 MAY 1986 EARTHQUAKE-TSUNAMI EVENT

Only the Most Pertinent Events are Listed
(Times in Greenwich Meridan)

07 May 1986		
2247Z	-	Earthquake origin
2254Z	-	Activiation of PTWC seismic alarm
2259Z	-	Preliminary location at latitude 51.3 N, longitude 175.0 W
2307Z	-	Preliminary magnitude (Ms) 7.8
2309Z	-	Received ATWC telephone confirmation of magnitude (Ms) 7.7
2317Z	-	ETA Adak
2331Z	-	Telephoned ATWC and received simultaneous telephone call from ATWC reported 38 cm wave activity at Adak at 2317Z
2348Z	-	Telephoned ATWC to confirm wave activity at Adak; waves increased to 70 cm; advised ATWC of PTWC's upgrading Regional Watch to Warning
08 May 1986		
0002Z	-	Telephoned NEIC to reconfirm magnitude (Ms) determination
0022Z	-	Received ATWC NAWAS announcement of ATWC Bulletin 003 175 cm wave activity at Adak
0029Z	-	Telephoned ATWC to request Shemya tide data; tide gauge out of operation
0050Z	-	Telephoned ATWC for tide information from Shemya; no activity reported
0115Z	-	Received ATWC NAWAS announcement of ATWC Bulletin 004 with 150 cm wave reports at Adak and 10 cm wave reports at Unalaska
0117Z	-	Received telephone call from ATWC with report of 175 cm waves at Adak with period of 15 minutes
0201Z	-	ETA Midway
0245Z	-	Received report of 67 cm maximum waves at Midway
0357Z	-	Received report from USSR of negative tide activity at Bering Island
0505Z	-	Notified Civil Defense via phone of cancellation of Warning
ATWC	=	Alaska Tsunami Warning Center
NEIC	=	National Earthquake Information Center
PTWC	=	Pacific Tsunami Warning Center

This discussion has simply provided constructive comments on the log of the warning center; hopefully it also illustrates the trials and tribulations of operating such a warning center to those who have not attempted this. The PTWC is understaffed and underfunded, as this author has written to NOAA.

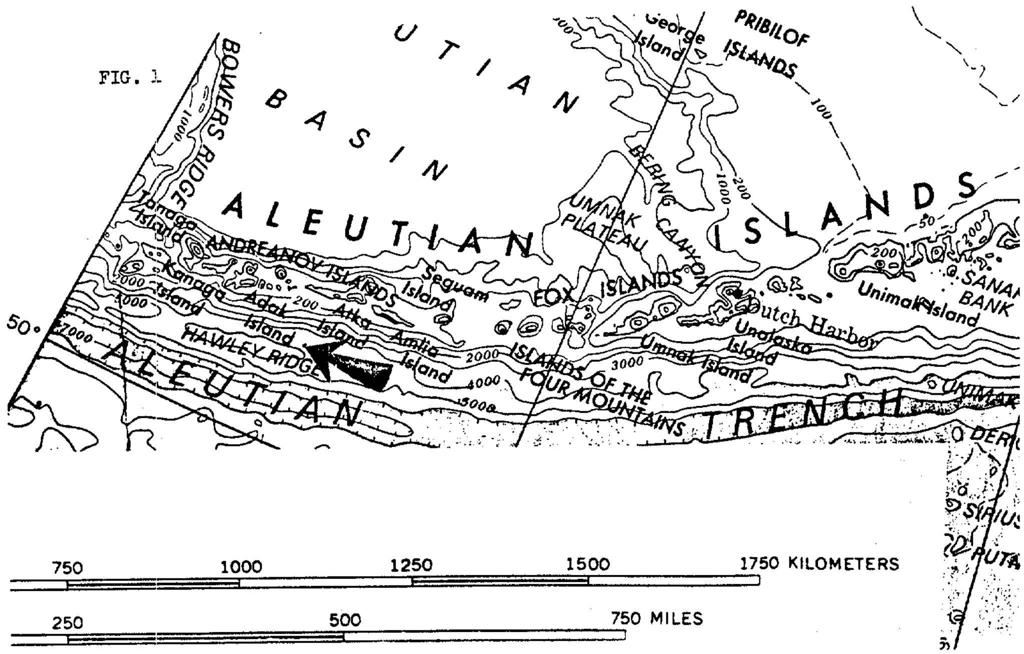

Fig. 1: Map showing epicenter of the Andreanof Islands Earthquake of 7 May 1986. Note the distance from the epicenter to Adak, the closest station observing the tsunami.

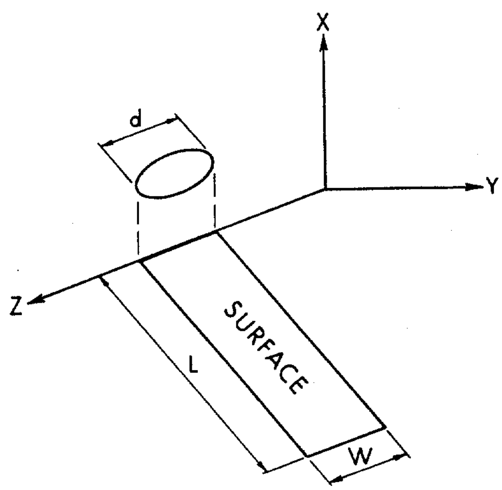

Fig. 2: Diagram showing the unusual faulting configuration. The tsunami period indicates an average diameter, d, for the source region, assumed circular. This is then taken to be one side of the fault, w. The area of faulting is S. Assuming a rigidity and estimating the seismic moment, the other length L can be found. For this earthquake, the strike length, w, appears to be much shorter than the dip length, L. The term “sliver” is suggested for this configuration of faulting.

FIG. 3

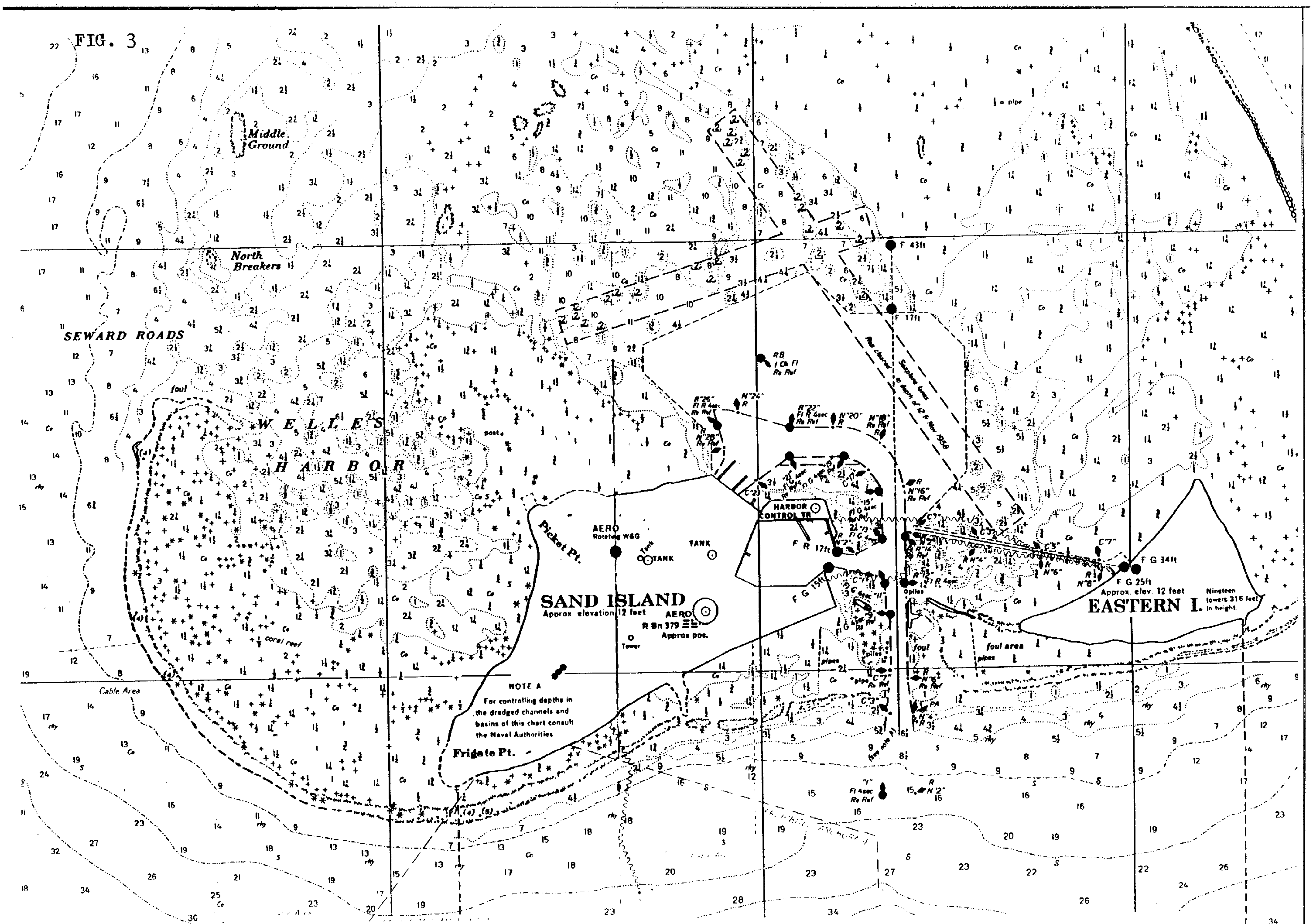

Fig. 3: Map of Midway Islands, showing the Sand Island on the west and Eastern Island on the east. Note the extensive dredging for channels, the small-boat harbor, and the sea-plane lanes.

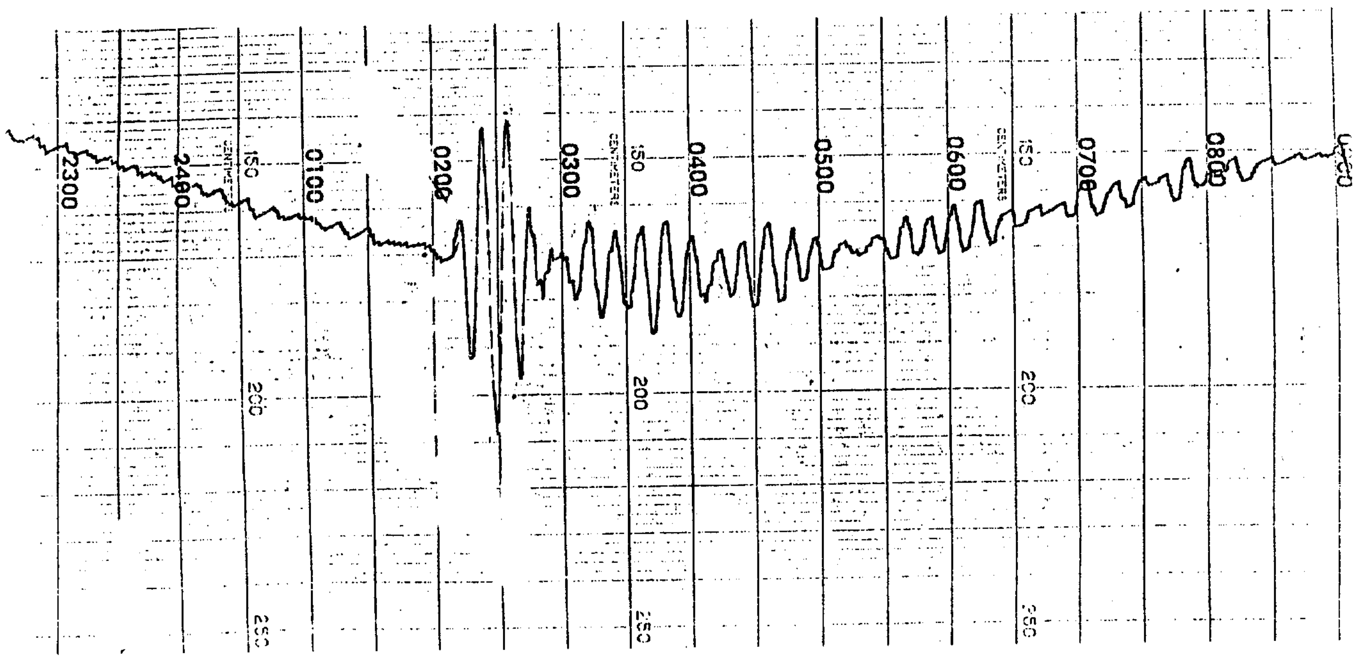

Fig. 4: A copy of the mareogram recorded at Midway for the tsunami from the 7 May 1986 Adreanof Islands earthquake. Courtesy of NOAA.

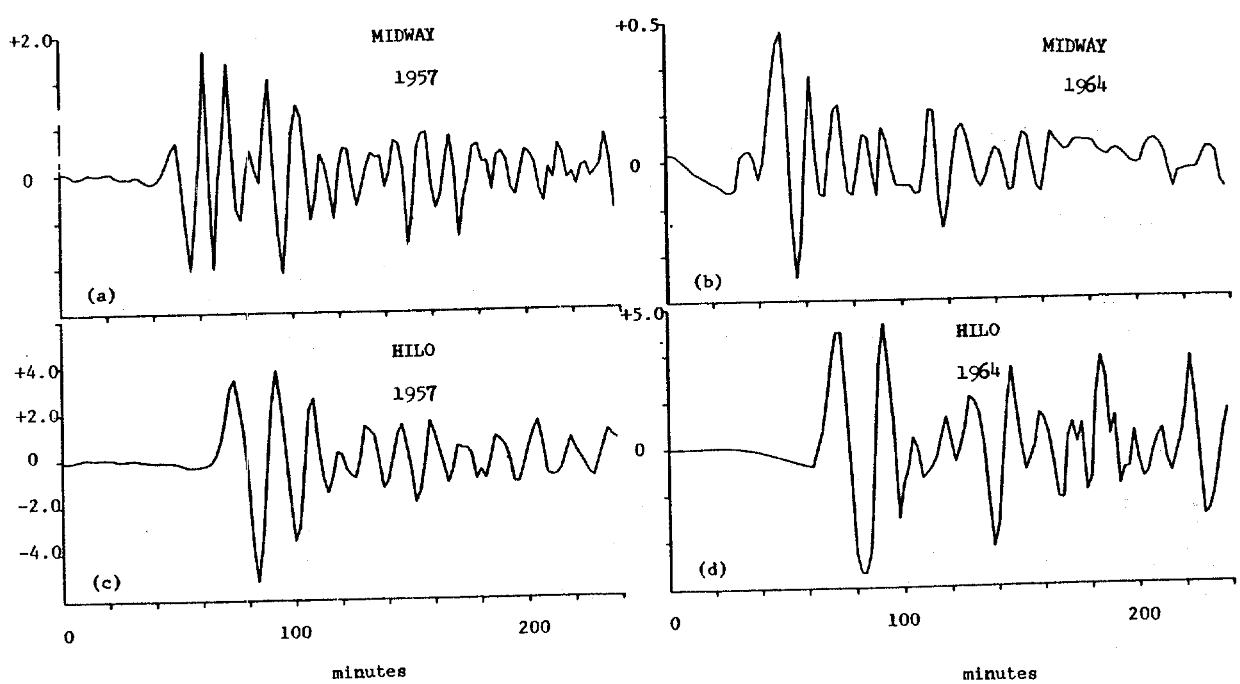

Fig. 5: Tracings of the mareograms recorded at Midway and Hilo for the 1957 and 1964 tsunamis (taken from Groves and Harvey, 1967). Note the high-frequency content for the upper, Midway, records.

Fig. 6: Diagram showing location of two water wells in Honolulu. The upper figure is the island of Oahu: the lower figure shows a segment expanded to show the positions of the Kapalama and Beretania water wells relative to the well-known landmarks of Waikiki and Diamond Head Crater. The scale is indicated by the shading on Sand Island, which extends 1500 feet inland. (The shaded area is the area to be evacuated during a tsunami warning.)

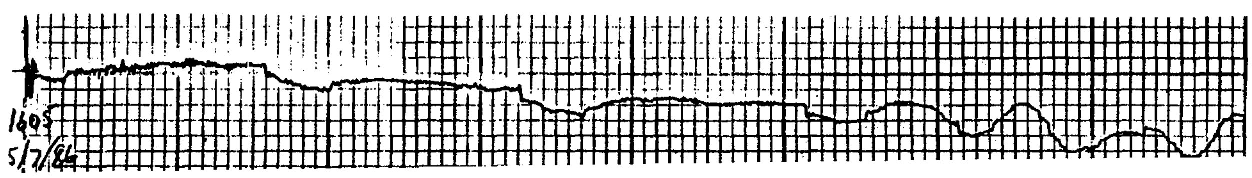

Fig. 7: Copy of the water-level recording taken at the Beretania well on a portable recorder used temporarily. The oscillations shown begin at the time of the arrival of the tsunami and have a period of about 13 minutes. (Twelve vertical lines equals fifteen minutes.) Courtesy of Chester Lao, Honolulu Board of Water Supply.

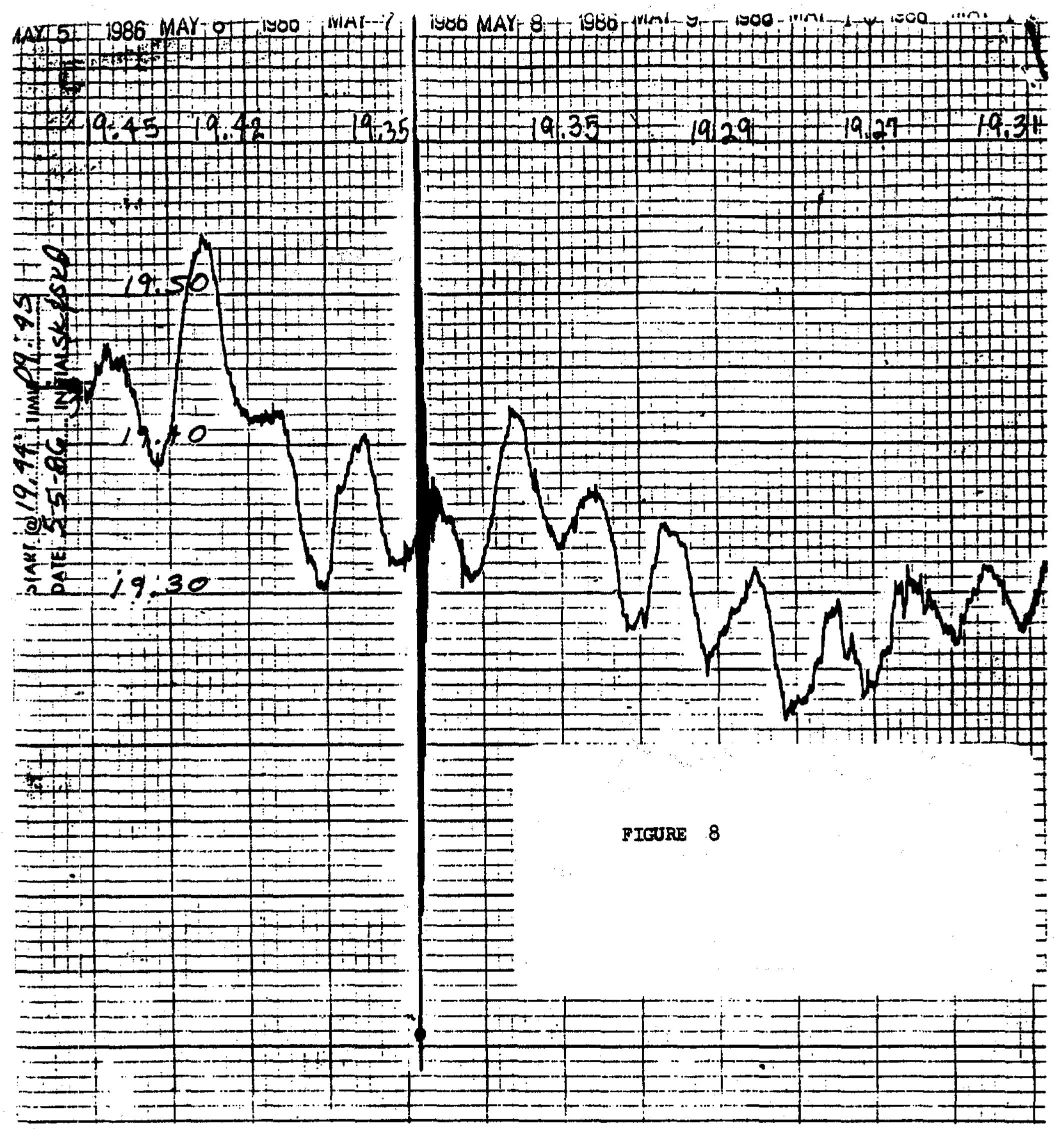

Fig. 8: Copy of the water-level recording taken at the Beretania well on the routine recorder. The oscillations shown occur at the time of the earthquake. There are no noticeable oscillations at the time that the tsunami arrived. (Twelve vertical lines equals one day.) The time that the record was begun is given on the left. This record is courtesy of Chester Lao, Honolulu Board of Water Supply

Real-Time Monitoring and Modeling for Tsunami Threat Evaluation*

by

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ABSTRACT

Data from the May 7, 1986 tsunami illustrate the importance of (i) better, prompt knowledge of source factors, (ii) multiple deep ocean sensors, and (iii) interacting this information with an adequate data base in near real-time to produce a valid assessment of a tsunami threat. A program is outlined to apply such data from the May 7 event, the Mexico event of September 19, 1985, and others, in a hindcast simulation to evaluate the utility of running a model in real-time to improve tsunami warning. An iterative procedure would be used in an actual sequence; it is believed this would help resolve anomalies such as the Midway data of May 7. Data and problems associated with that day's warning are reviewed, along with feasible improvements. Recommendations from two working groups and other references are included.

*Presented at the American Geophysical Union Fall Meeting, December 1986, abstract published in EOS, Nov. 4, 1986

Recent Events

The much-discussed May 7, 1986 mini-tsunami originated from a 7.6 magnitude earthquake near Adak, Alaska. Before the warning was issued, available data were compared with the damaging 1957 event of nearby origin. The 1957 magnitude was somewhat larger (8.3) and produced larger reported waves at the Adak tide gage, which probably was in the near field. At Midway Island, however, the 1986 waves were larger than 1957, thus apparently confirming the validity of the warning which had been issued.

The Midway gage, in the open ocean area 1000 miles from Honolulu, was in effect being used as a substitute for a deep ocean gage. The reason - in hindsight - why this was misleading is discussed in a recent paper (Adams, 1986). The May 7 event provided a much needed but very expensive (Cox, 1977) evacuation drill in several areas, plus the justification for this and other papers.

On Sept. 19, 1985, the devastating "Mexico City earthquake" (near the west coast of Mexico) coupled into the ocean enough to generate a tsunami and a watch condition was set. Due to lack of communication and gage data, it became necessary for the Pacific Tsunami Warning Center to either cancel the watch or to sound a warning without the information usually available. On the basis of knowledge of geology in the source area plus history, the watch was cancelled. Evacuation was avoided, correctly, as the maximum wave in Hilo was 22 cm. (The tsunami was significant in a portion of the Mexican coast).

What were the differences in these two events and the evaluation of their threat which lead to such different outcomes?

Primarily, it appears that the ability to rapidly access and correctly interpret available data were the key factors. The question is how can we do this every time, with the certainty that a damaging tsunami will be correctly forecast. The NSF Research Planning Group (Raichlen, 1985) and the HOE Working Group (Furumoto, 1985) urged more knowledge of the source mechanism, more sensor input, and modeling, with a goal of quantitative warning.

Modeling and Monitoring

It seems to us that at some point the model must be run in quasi real-time in order to reach the goal of quantitative monitoring, or to simply provide more accurate go/no-go warnings. We shall discuss some of these needs and their implications. The two events mentioned above provide a vehicle for study, and May 7 in particular shows the economic benefits. Factors to consider include:

- Need for geological data base

- Value of deep/open ocean real-time monitoring (especially if only single point)

- Need for rapid, better analysis of seismic factors (Adams, 1970)

- Need for a real-time model to effectively use these data

The need for better knowledge of source factors is obvious. To achieve this better input data is highly desirable - long wave/seismic moment data are probably most important. And next, a data base to which the raw inputs can be rapidly applied to produce the best geophysical estimate of source factors. This is presently done with reference manuals and experience. The emphasis on this factor from the planning groups should eventually aid real-time model work, as it is a natural tie-in.

Deep/open ocean monitoring is now accepted as both technically feasible and analytically desirable (Bernard, 1986). Note that use of the Midway data of May 7 as a substitute led to a mis-conclusion. This should not be a problem when such data become available and are fed into an adequate model. In addition, the multiple stations proposed (Bernard, 1986; Saxena and Murty, 1986) will also aid this problem. See Figure 1. Obviously, data from open ocean and coastal sensors must be available in near real-time to aid in threat evaluation, though historical data from even small tsunamis will greatly assist in analysis and future prediction. Because of location, logistic, and funding constraints, sites will be limited so we must make optimum use of any available data from them. Thus, a model is attractive. But, is it feasible?

Real-Time Model Factors

Clearly, numerical models of tsunamis propagating about the Pacific and impacting its shores have been used for many analytical purposes; there is not room (nor need) to cite the available references. Recently, Mader (1986) has incorporated his well-developed codes for water wave modeling into a book on the subject; tsunamis are a frequent example in the book. He points out that, with recent advances, it is now possible to run such programs on a personal computer, and he has done so. Normally, though, large main-frame computers have been used for such work. Can a tsunami warning center, which may not have instant access to a Cray, utilize a PC or other computer approach to threat evaluation?

Benefits and Conclusions

Thus, a hindcast procedure can be used not only to test and possibly verify the model; it should be used to improve it. A significant advance over present methods of assessment is expected, and hindcasting events of several sizes from a variety of areas will show how good a prediction can be made, how fast, and where the gaps in data or methodology are. A good model can also be used to show optimum locations for ocean sensors, and for shoreline observation points. A basic premise is that developing and operating a near real-time model will bring together the factors needed to evaluate a tsunami threat and, if fast enough, will usefully evaluate the threat.

How can it be made "fast enough"? The estimates in Table 1 are, like tsunami evaluations, based on inadequate data and must be conservative. The extrapolations are minimal, and by the time a system is implemented, electronics will probably have moved even faster. But, there are several other speed-up factors to consider.

One productive plan is to use adaptive cell size. There is no need to walk the wave across a thousand miles of relatively uniform open ocean in five mile steps. The wave can leapfrog open areas and return (with a spreading factor) to small cell spacing when it encounters shallow water or approaches within 100 km of a coast.

Some gain can be achieved by covering only a sector which involves areas of concern, or which affects the propagation (Saxena and Murty, 1986). Parallel processing by several microcomputers with separation by task or geography may prove feasible in this program, as it has in others. Progress in these areas need not be used only for speed-up; it may be used to better address near-shore effects, the shallow water assumption in the present code, or bottom friction factors. The need is to provide an adequate evaluation tool to those who provide the warning. And the goal is to provide an accurate binary (go/no-go) warning and eventually a quantitative, selective warning.

This can be achieved at a fraction of the cost of the May 7, 1986 evacuation.

Table 1 summarizes the calculations of this possibility. Note that the assumptions made are based on use of the system to aid in decision making rather than in a research analysis. The initial conversion from Cray time to PC-AT time was done with an actual wave model code (ZUNI) and problem. Other conversions are based on current data (Hinden, 1986) or other benchmark comparisons (Dongarra, 1986).

The table states that speed, not memory, is the basic limitation. This is so even though the computer has two significant memory functions, because even PC's now have both large RAM's and hard disks, to handle many megabytes of data. The two functions - storage of geophysical, bathymetric, and other reference data, and of the model and functional data - are handled readily. We suggest that considerable of these data are needed by a warning center anyway, for rapid access and use even if not run in a model, and should be in their database.

If run in a model, it would be desirable to have access to a powerful mainframe computer, but it appears that a 1990 "professional PC" could support a quasi-real time operation of the order of minutes. As an operating mode, we propose the following sequence:

- The model code has been prepared for this purpose and loaded and tested.

- The data bases have been loaded to a reasonable degree. These include: bathymetry for all significant areas, with a resolution approaching cell size where important; historic fault data and related source information; spectral characteristics of harbors and other gage sites; and historical run-up (wave) data for selected locations.

- A seismic event is inputed and the model run. Any gage input data (in a trial run) are inhibited until they actually become available.

- Some tsunami models are planned for one wave only. This should have at least three waves with gage data (e.g., Midway added at the appropriate time to affect the prediction. Thus, an iterative program is produced. Observed run-up can also be input to the model.

- The results are evaluated in hindcast, the model revised as necessary, and prepared for an actual event.

- It is exercised on such events, though small, and operationally verified.

TABLE 1. PROGRESS TOWARD A QUASI REAL-TIME TSUNAMI MODEL

Assumptions:
- 10 words (80 bytes) per cell
- 10 compute cycles per math operation
- shallow water (SWAN) code used
- speed is the variable; adequate memory is available

YEAR	MACHINE	CYCLE TIME, ns	TIME RE CRAY	TIME FOR 5 MILE RESOLUTION	TIME FOR 20 MILE RESOLUTION
Current	PC-AT	210	200	12 hr	3 hr
Current	VAX 11/750		10	30 min	8 min
Current	CRAY	2.7	1	3 min	1 min
1987 (Firm)	PC-386	63	60	2 hr	30 min
1989 (Est.)	PC-486	~30	~30	1 hr	15 min
1989 (Est.)	SUN/APOLLO (Upgraded)	~15	~15	30 min	8 min

NOTES:
1. Times vary greatly with configuration, operating system, source, graphics, etc. CRAY/PC-AT is actual wave model comparison.
2. 5 mile resolution requires 400 k cells; 20 mile requires 100 k.

SOURCES:
- Pacific models run at Los Alamos National Laboratory
- Argonne National Laboratory benchmark comparisons (Dongarra, 1986)
- Brief industry survey (Hindin, 1986)

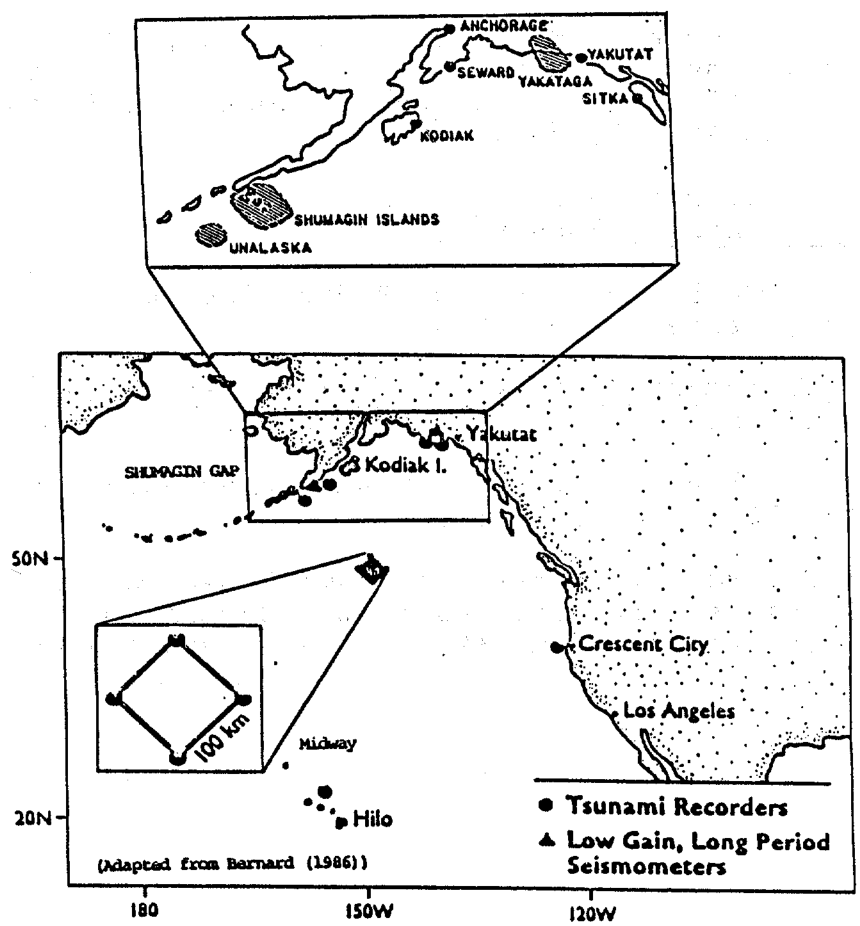

FIG 1

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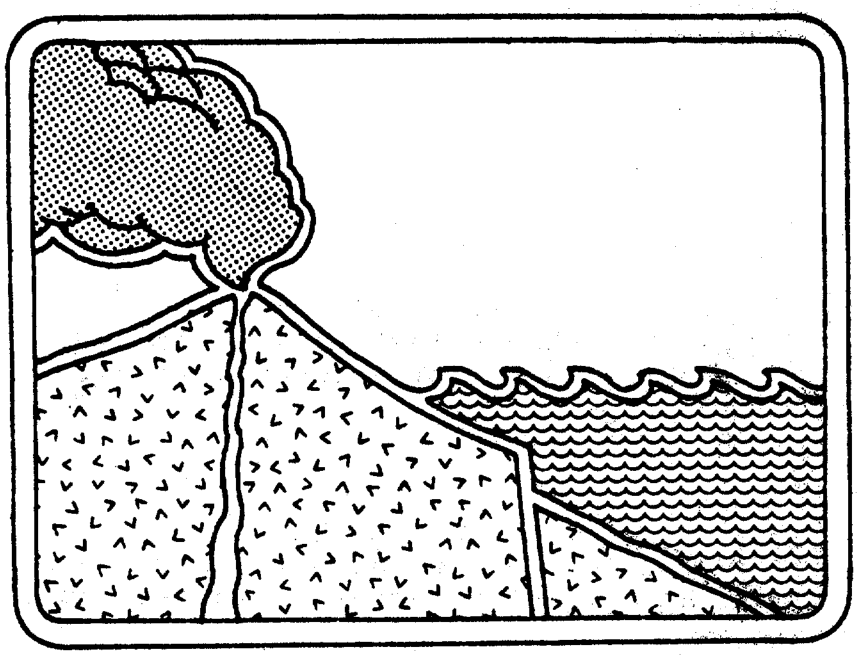

SPECTRAL ANALYSIS OF MAREOGRAMS FROM URUP TSUNAMIS OF 13 AND 20 OCTOBER 1963

by

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Introduction

In 1963 two strong earthquakes occurred near the Pacific coast of the Urup Island: those of October 13, $05^h 18^m$GMT, with magnitude M=8 8.25, and of October 20, $00^h 53^m$GMT, with magnitude M=7 7.5. The second source was situated in 100 km to east from the main source (fig.1). The first shock was felt on the Urup and Simushir Islands with intensity up to VII degrees (according to 12-degrees scale). Perceptibility of the second shock was three grades less [1, 2]. Both of these earthquakes produced tsunami waves, which were observed visually or by tide-gauges over the entire coast of the Kurils, at Sakhalin Island, in Japan, on Hawaiian Islands, etc. The maxima of tsunamis heights in the vicinity of the sources were 5 m and 15 m respectively, but in general the ratio of heights of the second tsunami to the first tsunami was equal to 0.4 [1, 2].

In the USSR the first tsunami was recorded by tide-gauges installed at Petropavlovsk-Kamchatskiy, Shumshu I., Matua I., Kurilsk, Yuzhno-Kurilsk, Katangli, Poronaisk, Korsakov, Krillon. The second tsunami was recorded by instruments at Shumshu I., Matua I., Kurilsk, Yuzhno-Kurilsk, Poronaisk, Korsakov, Krillon. Examples of typical tsunami records are shown on fig. 2.

These earthquakes and tsunamis have been investigated under the leadership of the senior author and a small booklet with full description of the events was published 20 years ago [1]. Copies of 9 mareograms from the tsunami of October 13 and 4 mareograms from the tsunami of October 20 have been reproduced, as well as tables with heights and other tsunami parameters. At the same time, efforts were started to find the spectra of the tsunamis but this work was interrupted and has been recently renewed.

Spectral analysis of mareograms is used widely in investigations of tsunamis. G. Miller in his pioneering work [3] has shown the similarity of spectra of different tsunamis at one given point and disimilarity of spectra of the same tsunami at different points. Investigating the spectra of numerous tsunamis recorded in the Hawaiian Islands, he has classified them according to their "colour" (frequency band). In the USSR, analogous investigation has been carried out by R.A. Yaroshenya [4].

Results of spectral analysis of tsunamis are often compared to spectral analysis of harbour's seiches and to results of numerical computations of harbour's resonance (see for example [5]). Spectral peaks of tsunamis usually coincide with maxima of transfer function of a harbour. In this paper the results of spectral analysis of the Urup tsunamis of 1963 are presented. Quality and duration of the mareograms permits analysis of the spectra in the frequency band of 0.1 to 10 cph, where the main tsunami energy is concentrated.

Contrary to previous authors we have tried to show that the spectra of tsumanis reflect not only the local harbour resonance but also specific properties of tsunami sources.

Spectral Analysis

Eight mareograms from October 13 and 4 mareograms from October 20, 1963 tsunamis have been processed. They were digitised with the step of 1 minute and then tidal oscillations were subtracted. There was visible tsunami effects on a majority of the tide-gauges records, the typical duration of oscillations was 8-20 h. Analysis was carried out to evaluate mean square water elevation σ^2, energy of oscillations ("energy of signal") $\varepsilon = \int_0^\infty \zeta^2(t)dt$, where ζ is the time, and tsunami spectra value of ε is correlated with the energy of tsunami waves in the point of observation as it will be explained later. The parameters are given in table 1.

Spectra were calculated with the help of the fast Fourier transform. A spectral time window was used to suppress the Gibbs effect. Smoothening of spectra was made by two methods. The spectra are shown on fig. 3. The main energy is concentrated in the range of periods from 5 to 100 minutes.

The levels of spectra vary strongly in dependence on mutual position of the tide-gauge and the source. It has been shown in [1, 2] that in the case of October 13 tsunami the decrease of tsunami height with ray distance r is described satisfactorily by the law $r^{-1/2}$, corresponding to simple cylindrical divergence. The decrease of the energy is described consequently by the law r^{-1}. It has been shown also in [1, 2] that after tsunami penetration into the Okhotsk Sea its height diminished by 2-2.5 times. The same phenomenon took place when tsunami penetrated from the Okhotsk Sea into the Japan Sea. These findings are confirmed in this investigation calculations of the values $\frac{1}{k}\varepsilon r$, where $k \simeq 0.21$ is the coefficient of penetration for tide-gauges situated in the Okhotsk Sea. Table 1 shows that the values r are almost constant at least for the event of October 13. It is interesting to point out that the coefficient of penetration is perhaps basically determined by the ratio of the total length of all deep-water Kuril straits to the length of the Kuril are which is equal to 0.25. A note-worthy feature was a high stability of upper freqeuncy boundaries of spectra for both tsunamis. They are equal to 6-7 minutes for tsunami of October 13 and to 4-5 minutes for tsunami of October 20. Low frequency boundaries of spectra vary from 60-100 minutes for the Pacific stations to 30-50 minutes for the Okhotsk Sea stations with the exception of Poronaisk. So the Okhotsk stations have narrower-band spectra. It is explained by the fact that long waves are cut off when tsunami passes through the narrow Kuril straits (narrower than 30-50 km). In other words long waves are reflected essentially from island arc while short waves penetrate easy into the Okhotsk Sea and the "colour" of spectra [3] is changed from "red" to "violet". The unusual structure of spectra at Poronaisk is explained perhaps by specific situtation of this point. It is exposed to the tsunami sources through the deep water Boussole strait.

Some of the spectral "peaks" are identical over spectra obtained at different stations. It is natural to explain this phenomenon by the influence of specific structure of tsunamis sources. Periods of 6-7, 20 and 33 minutes for tsunami of October 13 and periods of 10 and 20 minutes for that of October 20 are marked out distinctly.

However, general forms of spectra are determined basically by response functions of harbours where tsunami waves were recorded. So according to [4] seiches at Yuzhno-Kurilsk have principal period of 17 minutes and the peak on the period of 16 minutes is present in the spectrum of tsunami of October 13 (fig. 3). Seiches at Kurilsk have main period equal to 11 minutes [4] and there are peaks on periods of 9-10 minutes in spectra of both tsunamis at Kurilsk (fig. 3).

Evaluation of Parameters of Tsunami Sources

Seismotectonic processes in the source of a strong earth-quake have typical duration of the order of 10-100 sec while the time of "reaction" of strong tsunami waves can be taken equal to $L/\sqrt{gh}$, where L is the length of the source, h is the depth of the water. This time is of the order of 10-100 minutes. So in the problem of tsunami generation we can assume that the bottom is dislocated instantaneously. The relief of the initial water elevation in the tsunami source reflects the bottom deformation

However, it is very difficult to restore the inital elevation of water surface using records of tsunami obtained only by coastal tide-gauges. It is explained by tranformation of tsunami waves during their propagation due to such factors as 1) geometrical divergence, 2) dispersion, 3) absorbtion, 4) reverberation after multiple reflections of waves from bottom slopes and from shores, 5) selective amplification of some spectral components in straits, bays, etc.

It must be noted that factors 1 and 2 practically do not change the form of the energetical spectrum of waves. As for factor 5 it must be indicated that usually response functions of harbours and shelf can be calculated and their influence on the spectra can be taken into account. Thus the form of energetical spectrum of tsunami must be among the most stable characteristics of waves. This approach has been used for instance in the work [6] where total energy of tsunami has been tentatively evaluated by spectral analysis of coastal mareograms.

According to some data (see [3] for instance) tsunami spectrum in the source must be a comparatively even function of frequency. Therefore, it is reasonable to model the initial water elevation in the tsunami source by an axisymmetrical dome:

$$\zeta_o(r)=ae^{-\frac{1}{2}(r/D)^2} \quad (1)$$

where a is the maximal elevation of water, D is a conditional radius of the source. According to [7] far off the source the wave elevation (1) is described as follows (fig. 4):

$$\zeta(r,t)=a/2\int_0^\infty \sqrt{2kd/\pi r}\cos\left[k/d(r-ct)-\pi/4\right]e^{-\frac{1}{2}k^2}dk \quad (2)$$

where $c=\sqrt{gh}$ is the velocity of long waves. It is similar to the well known R. Takahashi's solution [8] for the profile of a tsunami from a piston-like cylindrical deformation of the sea bottom.

Energetical spectrum of such signal is equal to

$$s(w)=\frac{1}{2}\frac{a^2D^2}{c^3r}w\,e^{-w^2D^2/c^2} \quad (3)$$

Where D/c is the specific signal duration τ.

Of course radiation of real Kuril tsunamis excited usually on the steep contintental slope of the island arc is not isotropic. Therefore proposed model is not quite suitable for adequate description of tsunami propagation. However, it can be used for rough estimations of such tsunmai characteristics as specific duration τ.

We will try to estimate values of τ. It is understandable that if we will plot spectrum (3) using axes w^2, lg S expression (3) will transform into the straight line with inclination equal to $-\tau^2$. It has been noted already that usually spectra are distorted by local resonant properties. This distortion is more essential for periods less than 30 minutes. Low-frequency part of spectrum basically charaterizes the initial signal in the source.

On fig. 4 a the spectrum at Poronaysk for October 13 tsunami is drawn using the mentioned axes. Regression line found according to the least squares method is shown. The inclination (regression coefficient) is equal to $\rho=-13.0$ min^2, which yields $2\tau=7.2$ minutes. For the tsunami of October 20 the same operation (fig. 4 b) yields $2\tau=5.4$ minutes, which is less than in previous case. Values of τ for other points of observation are of the same order (table 2).

Now it is possible to estimate the dimensions of the tsunamis sources. We assume that the length of tsunami source L is equal to L=2.51D and that the water depth in the tsunami sources is equal to h=5000 m. Then we have L=100±15 km for the tsunami of October 13 and L=80±25 km for the tsunami of October 20. These estimations are close to dimensions of short axes of tsunami sources found by inverse refraction diagram method method (fig. 1). This result can be explained by the fact that the main tsunami energy is radiated from long side of tsunami source.

Let us attempt to estimate the energies of tsunamis. Using the method described in (6) we suppose at first that the total energy of tsunami E_t may be evaluated as

$$E_t = \frac{1}{2}\pi\rho g\varepsilon r$$

These energies are equal to 10^{22} ergs for the tsunami of October 13 and to 10^{21} erg for the tsunami of October 20. But we have overestimated energies because coastal records of tsunamis are contaminated by local seiches of long duration. To obtain real values of tsunami energies it is necessary to integrate records only during the arrival of the direct tsunami signal. Using a dome-like profile we can propose that the duration of the arriving signal is equal approximately to 6τ. Remembering that values of τ are equal to 22 and 16 minutes for the tsunamis of October 13 and 20, we conclude that the tsunamis energies are equal to 10^{21} and 10^{20} ergs respectively.

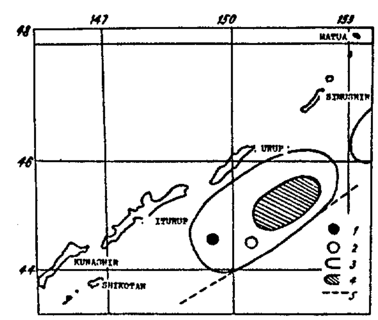

Fig. 1. Locations of sources of Urup tsunamis occured in 1963. 1 - epicenter of October 13 earthquake, 2 - epicenter of October 20 earthquake, 3 - source of October 13 tsunami, 4 - source of October 20 tsunami, 5 - axis of the Kuril deep-water trench (from [1]).

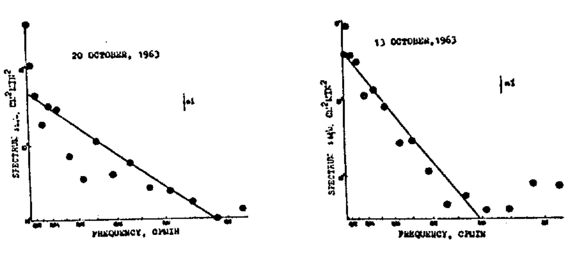

Fig. 4. Tsunami spectra at Poronaisk plotted in quadratic-logarithmic scale.

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Table 1

Some parameters of tsunami records

13 October, 1963

Station	Ray distance r, km	R. m. s. σ, cm	Signal energy E, $10^4 cm^2 min$	Adjusted signal energy Er, $10^7 cm^2 min \cdot km$
1. Kurilsk	240	14.8	28.2	3.9
2. Yuzhno-Kurilsk	560	22.0	62.0	3.1
3. Krilion	610	7.7	7.5	2.4
4. Korsakov	630	8.4	9.1	3.3
5. Shumshu	630	6.0	4.6	-
6. Poronaisk	685	8.5	9.2	3.8
7. Katangli	985	7.3	6.7	3.7
8. Petropavlovsk-Kamchatskiy	1150	1.4	0.2	-
		20 October, 1963		
1. Kurilsk	240	7.5	7.3	0.74
2. Matua	300	9.4	11.4	0.26
3. Shumshu	630	3.5	1.5	-
4. Poronaisk	685	4.3	2.4	0.73

Table 2

Some characteristics of tsunami spectra

13 October, 1963

Station	Linear regression coeff. β, min^2	Correlation γ	Signal Duration 2τ, min	Source dimension 2.51 D, km
1. Kurilsk	- 4.5	0.69	4.24	71
2. Yuzhno-Kurilsk	-10.3	0.93	6.42	109
3. Krilion	- 8.7	0.91	5.90	100
4. Korsakov	- 8.5	0.92	5.83	99
5. Shumshu	- 8.1	0.86	5.69	96
6. Poronaisk	-13.0	0.96	7.21	121
7. Katangli	-10.5	0.95	6.48	110
8. Petropavlovsk-Kamchatskiy	-10.6	0.94	6.51	110
average			6.04±0.82	102±14
	20 October, 1963			
1. Kurilsk	- 1.3	0.37	2.28	39
2. Matua	-5.3	0.78	4.60	78
3. Shumshu	-10.7	0.94	6.54	111
4. Poronaisk	- 7.3	0.87	5.40	90
average			4.70±1.56	80±26

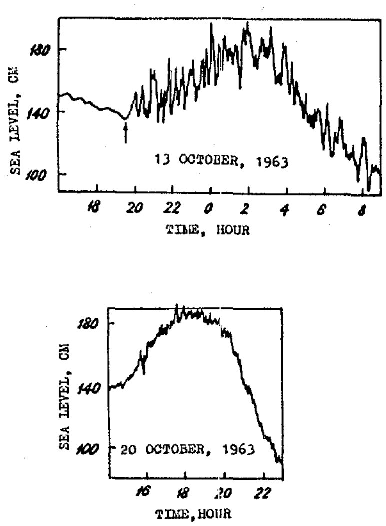

Fig. 2. Records of tsunamis at Poronaisk.

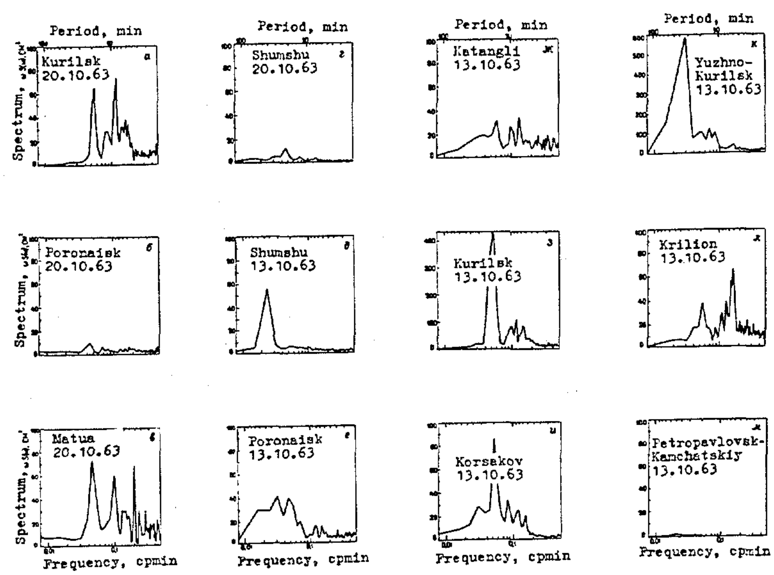

Fig. 3. Spectra of tsunamis.

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