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FOREWORD

On 23-rd of March 1988 a symposium was held in Bologna, Italy with the title: **Tsunamis generated by earthquakes and volcanic eruptions: theory and observations.** The meeting was organized in the framework of the XIII General Assembly (Bologna, 21–25 March, 1988) of the European Geophysical Society (EGS), an organization that counts over 1,000 members among the european as well as the extra–european geophysicists. EGS publishes important scientific journals with worldwide circulation, such as Annales Geophysicae, Tectonics and the Geophysical Journal, the publication recently formed by merging together the Geophysical Journal of the Royal Astronomical Society, the Journal of Geophysics and the solid earth part of Annales Geophysicae. EGS promotes annual General Assemblies consisting of several symposia, workshops, special lectures, joint and interdisciplinary sessions... covering the whole spectrum of geophysics from planetary and solid earth sciences to hydrosphere and atmosphere sciences and external geophysics. Next year, the EGS General Assembly will take place in Barcelona, Spain from 13-th to 17-th of March.

The tsunami meeting was convened by prof. W.M.Adams (USA), by prof. S.L.Soloviev (USSR) and by prof. S.Tinti (Italy) and received sponsorizations from the Istituto Nazionale di Geofisica (ING), Rome and from the Intergovernmental Oceanographic Commission of Unesco (IOC), Paris. Special thanks must therefore be addressed to prof. E.Boschi, President of the ING and to Dr. M.Ruivo and Dr. I.Oliounine (IOC), who all encouraged the symposium organization by offering advise and help since the very initial stage.

The significance of the Bologna tsunami meeting goes far beyond the fact that it was a very successful event from a scientific point of view, due to the participation of several distinguished experts and scientists. It was first of all a remarkable occasion to emphasize to the european geophysicists that tsunamis are not only a "monopoly" of the Pacific Ocean, as many erroneously think, but that, as historical records clearly testify, large–scale tsunamis did occur in the past also in the Mediterranean Sea and in the Atlantic Ocean, where of course nobody can exclude that they can be generated again even in the future. Indeed, the most catastrophic tsunami ever seen in historic times pertains to the Aegean Sea and was caused by the explosive eruption and the consequent caldera collapse of volcano Santorini in the Bronze Age. It was an event analogous, but much more violent than the well known tsunami that destroyed the villages on the coasts of Sunda Straits after the Krakatua explosion in 1883. But there is no need to go back to the ancient times to have notion of disastrous events. Even in the present century large tsunamis did affect the Mediterranean coasts. On 28th of December 1908 a tremendous tsunami invested the Sicily and Calabria coasts of the Messina Straits, ensuing a normal–fault $M = 7.1$ shallow earthquake; waves higher than 5 m were reported in several places of the Straits where they caused many victims and severe destruction. On 9th of July 1956 a tsunami, very likely caused by landslides triggered by a large $M = 7.5$ earthquake, occurred in the Greek Archipelago in the Aegean Sea, where in the isle of Amorgos wave amplitudes up to 20–30 m were observed. The Bologna symposium had therefore the merit to catalize the attention of the whole XIII EGS General Assembly on the fact that tsunamis are natural phenomena of general concern. The Assembly recognized that little efforts were made in the past to study tsunamis not originated in the Pacific Ocean and, in order to promote research and international cooperation, approved the constitution of an international Working Group entitled *"Tsunamis in the European Seas and in the Atlantic Ocean"*. The activity of the Tsunami WG, coordinated by prof. S.Tinti, will concern a quite broad spectrum of items: the assembling of high–quality catalogs of tsunamis, the tsunami zoning, the calculation of the travel–time charts, the numerical simulation of the major tsunamis as well as of the future predicted events, the theoretical aspects of the tsunami generation and propagation, the coastal hazards assessment and mitigation, the tsunami acquisition and processing, the tsunami warning system. The Tsunami WG is open to all researchers and scientists who are willing to participate in its activities, irrespective of the EGS membership, and who are invited to contact the WG coordinator.

The scientific importance of a meeting is known to be related even to the possibility of disseminating the ideas and the results illustrated in the course of the congress to the maximum extent. One of the best way to achieve the goal is the publication of the proceedings. Well aware of this, the convenors accepted promptly the offer of having the tsunami symposium proceedings published by a scientific journal. Among the others, the Science of Tsunami Hazards was preferred, because i) it is a prestigious journal, specifically devoted to the tsunami research; ii) it is distributed among all the tsunami experts of the world; iii) it gives the chance of a quick and capillary circulation. Of course, the convenors are sincerely grateful to prof. T.S.Murty, the technical editor of the journal and to all the editorial staff. Editor of the proceedings are prof. S.Tinti and prof. T.S.Murty. They wish to inform the readers that the papers included in the proceedings are essentially unrefereed and that neither the journal nor the proceedings editors take any responsability of the content of the papers, responsability that fully pertains to the authors. The choice to eliminate the lengthy procedure of the referee revision was uniquely motivated by the opportunity to speed up the proceedings publication.

Bologna, Italy S.TINTI
May, 1988

ON AUDIBLE TSUNAMI ON THE COAST *

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ABSTRACT

A simplified model of acoustic tsunami precursor is developed in order to have a more reasonable physical understanding of an audible boom on the coast heard just after the occurrence of a big earthquake and just before the first tsunami arrival. Such acoustic precursors have been found in the historical descriptions on the coast of the northwestern Pacific. A quite similar remark was found also in a recent report of the 1977 Sumbawa tsunami prepared and issued by the Indonesian Governmental Publication Office. The model introduced in this work deals with the problem of an acoustic ray generated by a sound source associated to an undersea earthquake and subsequently refracted across the air–sea interface. A theoretical estimate of the audible boom is studied to give a possible case of an audible tsunami precursor. This model is yet hypothetical, though it is clear that the past hazardous tsunamis in the northwestern Pacific accompanied the large undersea earthquakes with audible offshore or atmospheric booms that were real tsunami precursors. So that, the boom can be exploited for a short–range prediction if this model is appropriately improved.

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* A part of this work was presented at the general assembly of the European Geophysical Society held in Bologna in March 1988.

INTRODUCTION

A simplified model of acoustic tsunami precursor is developed in order to help a more reasonable understanding of an audible boom on the coast just after a large undersea earthquake and just before the tsunami arrival. The boom can be found in the historical descriptions in Japan, especially in the coastal zone of the northwestern Pacific. The most remarkable point is that only the hazardous tsunami is accompanied by the boom in the western Pacific. The related researches and reports are for example the ones published by the International Tsunami Information Center (ITIC) in 1977, by Pusat Metereologi dan Geofisica (1977) and by Nakamura since 1979. The Indonesian event is expressed as "explosive sounds heard in a district after the earthquake and before the tsunami arrival". The boom is depicted in the past description as a cannon's boom or as a thunder roll in the northwestern Pacific.

A simplified model of a sound field in the sea is assumed first and its application is made in order to give a physical basis to the reported occurrence of the boom. A hypothetical modelling is extended to cover a possible atmospheric boom as the tsunami precursor. In the last part of this work a possible case study of the boom is introduced for the area of the Ionian Sea in the neighbourhood of the Messina Straits, Italy.

ACOUSTIC RAYS UNDER THE SEA SURFACE

There have been many contributions on tsunami research so far, but their main interest was only restricted to learning the dynamical relation between the main generating shock and the ensuing tsunami. In some descriptions, optical anomalies in the sky have been also remarked and pointed out in relation to the hazardous earthquakes; however, they have been taken so far only as a minor problem. Therefore, seismologists tend to develop their theories to solve problems concerning the prediction of the forthcoming hazardous earthquake, whereas many works in hydrodynamics have been devoted only to clarify the properties of the tsunamis on the coast. Recent innovations in the field of electronics and the consequent applications to high-speed computers have been combined with several successful techniques of numerical modelling of tsunamis. Notwithstanding the general progress in the tsunami research, the author however could not find any work concerning an acoustic tsunami precursor except his primitive modelling (see Nakamura, 1986b).

One problem is to consider the possible subsurface sources of a large-tsunami precursor from a theoretical point of view. One kind of such possible sources is the fault line of an earthquake on the sea floor. The other possibility is the formation of a concentrated caustic in the surface layer of a stratified ocean. By this time, no consideration of the oceanic stratification has been taken into account for the problem of the acoustic tsunami precursor except by Nakamura (1986b).

Now, we consider a simplified layered ocean. The stratification in the ocean is determined by the vertical distribution of the water temperature and salinity. In addition, the sound speed may be determined as a function of the water temperature, salinity and pressure. When the sound speed C is given as a function of the water depth z, then the surface condition can be written as $C = C_0$ at $z = 0$.

The Snell's law is applicable for the acoustic refraction at the interfaces of a stable multiple-layer ocean. If we assume that an acoustic ray in the i-th layer with an incidence angle w_i (with sound speed C_i) crosses the interface into the neighbouring j-th layer with refraction angle w_j (with sound speed C_j), then:

$$sin(w_j)/C_j = sin(w_i)/C_i. \qquad (1)$$

When each one of the stratified-layer thickness is negligibly small, it can be taken as a vertically continuous stratification with thin layers dz; and an acoustic ray at an arbitrary depth z can be simply related to the same ray at the reference depth z_i as follows:

$$sin(w)/C(z) = sin(w_i)/C(z_i) = a. \tag{2}$$

For a path of an acoustic ray ds, the depth dz and the travel time dt are related as:

$$ds = dz/cos(w), \tag{3}$$

$$dt = ds/C(z). \tag{4}$$

And for the horizontal travel distance dR:

$$dR = tan(w)dz. \tag{5}$$

Using these relations, we can get the horizontal distance from an acoustic source at the point (z_i, R_i) to the receiver station (z_f, R_f) as follows:

$$R_f - R_i = \int_{z_i}^{z_f} aC(z)\left(1 - a^2C(z)^2\right)^{-1/2} dz, \tag{6}$$

while the travel time from the source to the reciever station is:

$$t_f - t_i = \int_{z_i}^{z_f} C^{-1}(z)\left(1 - a^2C(z)^2\right)^{-1/2} dz. \tag{7}$$

One special case is the one where the sound speed varies linearly:

$$C(z) = C(z_i) + b(z - z_i). \tag{8}$$

With this, the integration of (6) and (7) can be easily obtained analytically (see for example Clay and Medwin, 1977). The results are written in the simple form:

$$R_f - R_i = \frac{1}{ab}\left(cos(w_i) - cos(w_f)\right), \tag{9}$$

and

$$t_f - t_i = \frac{1}{b}\log\left[\frac{V_f(1 + cos(w_i))}{V_i(1 + cos(w_f))}\right], \tag{10}$$

where for $k = i$ and $k = f$:

$$V_k = z_k - z_i + C(z_i)/b, \tag{11}$$

$$w_k = 1 - a^2b^2V_k^2. \tag{12}$$

Now, the acoustic ray forms a part of circle. The radius of the circle is :

$$r = 1/(ab), \tag{13}$$

where no dissipation is considered for the sound propagation along the acoustic ray.

In the real ocean, the vertical distribution of the sound speed can be replaced by a two–layer model with the sound channel at the bottom of the main thermocline. In order to get the sound speed for given values of the water temperature, salinity and depth, a formula obtained by Clay and Medwin (1977) can be utilized, following the example of Nakamura (1986a) for a case off the south of the Japan Islands. An application of the above theory has been made by Nakamura (1986a,1986b, 1987, 1988a). As Nakamura (1987) pointed out, only a faint subsurface caustic can be expected in the higher latitude area even in the northwestern Pacific. This must be caused mainly by (1) a significant seasonal or annual thermocline up to 100m thick in summer and (2) an almost uniform water of low temperature in winter. However, reconcentration of acoustic rays emitted from the source can be expected when a specific condition is satisfied for the horizontal water temperature distribution. This problem regards the horizontal acoustic ray for the tsunami precursors recorded in the descriptions concerning the events in the eastern Japan.

ACOUSTIC TRANSMISSION

Sound generated at the epicenter on the sea–floor propagates in the sea to form a subsurface concentrated caustic which is another sound source as seen above; almost all of the sound energy, however, must be transmitted vertically as the author expects after some theoretical considerations.

Now, assume that two kinds of fluids, i.e. fluid 1 (sea water) and fluid 2 (air) are separated by a horizontal interface $P - P'$. The interface is assumed to be flat and no disturbance is considered initially. For our convenience, the densities and sound speeds for the two fluids are denoted for the sea water as ρ_1 and C_1 and for the air as ρ_2 and C_2 respectively.

When an acoustic ray r_i in the sea gets to the sea surface and is partly transmitted into the air as r_t and partly reflected as r_r, we can start to consider the energetics of the tsunami precursor. The energy E_i of z_i is a sum of the energies E_t of the ray r_t and E_r of the ray r_r. This means to consider a system where the energy is conserved, that is $E_i = E_r + E_t$. As for energy transfer:

$$\int C_1 E_i df = \int C_1 E_r df + \int C_2 E_t df, \tag{14}$$

where f is the frequency of sound for the interested acoustic ray and the integral extends over the range of the frequency band width of the sound. In this step, it is more simple if no reflection can be assumed in (14), since the problem becomes:

$$C_1 E_i = C_2 E_t + constant. \tag{15}$$

As the acoustic wave energy is proportional to the density of the interested medium and to the square of the amplitude of the acoustic wave in the medium, then, $C_1 E_i$ is proportional to $\rho_1 A_i^2$. Hence,

$$(C_2 E_t)/(C_1 E_i) = (\rho_2 A_t^2)/(\rho_1 A_i^2). \tag{16}$$

With the assumption of E_r above, the absolute value of E_t has the same value of E_i. So that, when we consider the absolute values, the above expression can be written in a more simple form:

$$C_2/C_1 = (\rho_2/\rho_1)(A_t^2/A_i^2). \tag{17}$$

Now, remind the properties of the sea water and the air. That is to say,

$$\rho_1 = 1.03 \quad \text{and} \quad C_1 = 1500 m/s \quad \text{for sea water} \tag{18}$$

and

$$\rho_2 = 0.00129 \quad \text{and} \quad C_2 = 330m/s \quad \text{for air of } 0°C \tag{19}$$

approximately, so that:

$$(A_t/A_i) = (\rho_1/\rho_2)^{1/2}(C_1/C_2)^{1/2} = 0.986 * 2.132 = 2.1. \tag{20}$$

As far as we refer to acoustic energy in relation to wave amplitude A in a medium of density ρ, the relation is $E = (1/2)\rho A^2$. Then,

$$(C_2/C_1)(E_t/E_i) = (\rho_2/\rho_1)(A_t^2/A_i^2) = 0.00125 * 4.42 = 0.0055, \tag{21}$$

or

$$(E_t/E_i) = 0.0055(C_1/C_2) = 0.0118. \tag{22}$$

This result shows that the rate of acoustic energy transmitted from the sea water into the air is an amount of approximately 0.0118 part of the initial energy across the sea surface. Since the sound intensity is proportional to the square of the sound pressure (see Clay and Medwin, 1977), the above result can be taken to show that the sound energy detected in the air after crossing the sea surface is almost one hundredth of the sound energy in the sea. If the acoustic energy in the sea is strong enough, say $10^{10} erg$, the transferred energy into the air must be of about $10^8 erg$, which must be felt by naked ears on the coast as a boom or a thunder roll.

ACOUSTIC ENERGY DENSITY ON THE SEA SURFACE

The author previously referred to the tsunami catalogs (Iida, 1984;Iida et al., 1967;Soloviev et al., 1974;Watanabe, 1986). In this work, the author utilized the observed result of the vertical distribution of water temperature and salinity at a station in the Ionian Sea (Böhm et al., 1987). The location of the station is shown in Fig.1. The vertical distribution of salinity and water temperature are shown in Fig.2. The right part of Fig.2 shows the vertical distribution of sound speed. Sound is assumed to be generated on the 1.2 km deep sea–floor. Then, a propagation pattern of the acoustic ray is computed as seen at the bottom of Fig.3.

Fig.1

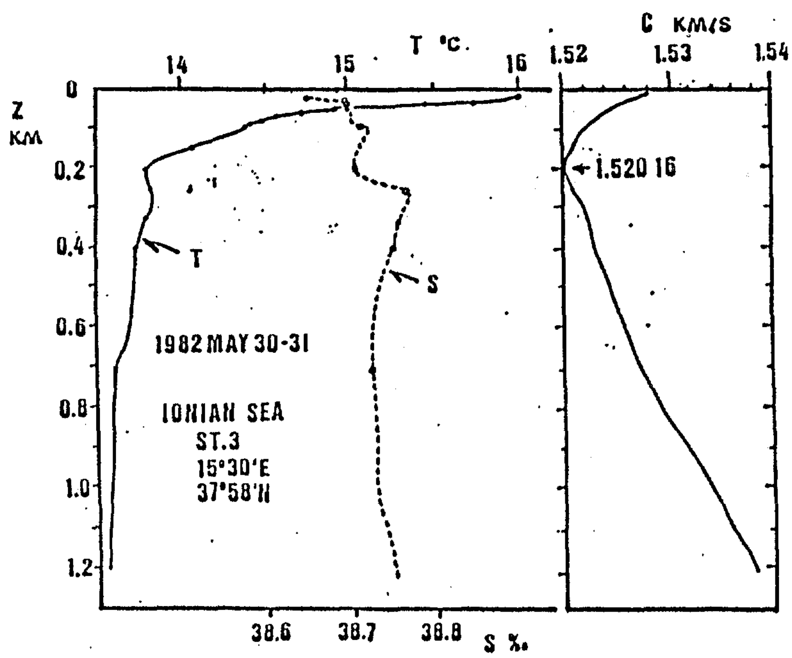

Fig.2

Relative acoustic energy density on the sea surface is estimated and shown at the top of Fig.3. In this case, no subsurface concentrated caustic is expected. Nonetheless, a boom can be detected by a person on the coast at the heigth of 10m in a distance less than about 80 km directly, even if the earth's surface curvature is considered. This work was stimulated by contributions by Tinti and his group in 1987. With Sapia et al. (1987), the more detailed study must be necessary there. As for the atmospheric tsunami precursor, it is necessary to consider the vertical distribution of the air temperature up to the thermosphere at least.

SEISMIC WAVE-T PHASE

As for the so-called "T phase", it is found in observations by Ewing and his group. Although, they know only that their T phase propagates at a speed nearly the same as the sound does in the water. The author considered here the oceanic stratification, though Ewing and his group never took into account any effect of the oceanic stratification on the tsunami precursor or on the formation of any concentrated caustic, which must surely be the cause of a boom as a tsunami precursor soon after the occurrence of a large earthquake or before the arrival of a hazardous tsunami.

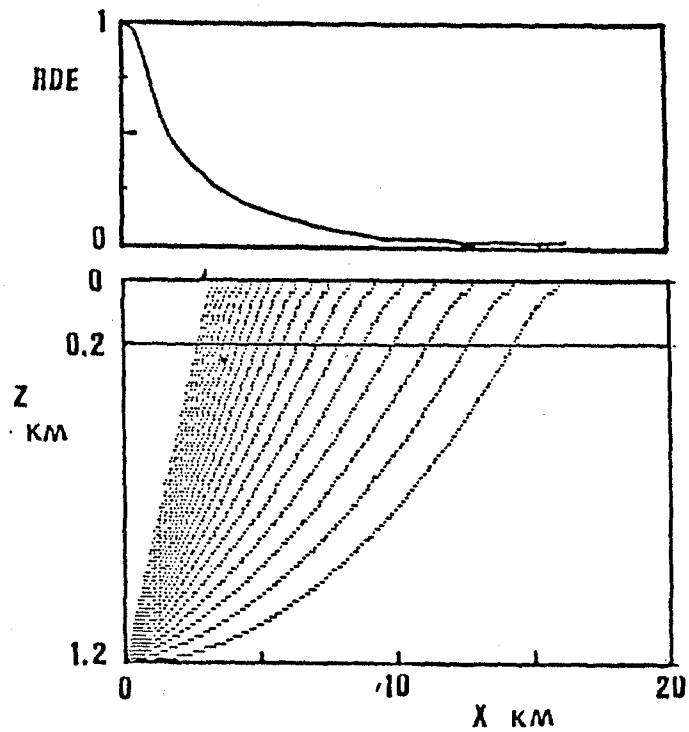

Fig.3

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A REVIEW OF THE HISTORICAL 1627 TSUNAMI IN THE SOUTHERN ADRIATIC

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ABSTRACT

One of the largest tsunamis that invested the Italian coasts in the Southern Adriatic sea occurred on the 30th of July in 1627 in the promontory of Gargano. It was generated by a large earthquake with epicentral intensity $I_0 = XI$ (MCS scale). The analysis of the evolution of the tsunami implied the precise evaluation of the earthquake effects. The epicenter was a few km on land, according to the analysis of the coeve sources, that are particularly abundant of information on the shock only for the inhabited areas. The shallower coastal region, however, was wild and unhinabited in the 17th century, since the settlements were situated away from the shore, in places reputed safer and more protected from the raids of pirates. This entails that the original sources on the tsunami are not extremely detailed. Notwithstanding, we have been able to locate and evaluate the most important effects of the waves, that involved a considerably large region. The area most affected by the event was the shoreline between Fortore and San Nicandro, close to the Lake of Lesina in Northern Gargano, that after the first sea withdrawal was deeply inundated by the waves. As to the tsunami risk, we point out that a tsunami with the same characteristics as the one occurred in 1627 could be highly destructive today, mainly because the settlement network has grown larger and larger and many touristic structures are now in existence close to or on the shore, that was previously uninhabited.

INTRODUCTION

The catalog of the Italian tsunamis due to Caputo and Faita (1984) reports four major historical events that were generated by earthquakes in the Adriatic Sea and that were rated larger than intensity V in the tsunami Ambraseys scale (Caputo and Faita, 1982): as to three of them, ascribed to the Northern Adriatic, they have been carefully reviewed in a recent paper by the authors (Guidoboni and Tinti, 1988), who showed that they were overestimated. The 1106 case in the Lagoon of Venice was indeed proven to be a false event, whose origin was due to a curious chain of misinterpretations in the historiographic tradition. The present work deals with the fourth one of the largest Adriatic tsunamis, that invested the coasts of Puglia in 1627. In the Caputo and Faita's catalog the tsunami is associated to an earthquake with estimated intensity IX and epicenter on land at $41^0\,50'$ Lat E and $15^0\,20'$ Long N, a few kilometers distant from the shoreline in the Gargano promontory (Puglia).

The seismic history of Gargano shows that the area is the most unstable region of the South-Western Adriatic. A recent study to determine the most active seismogenic zones of the Italian territory (see Mulargia et al., 1987) shows that the whole peninsula of Gargano together with the adjacent marine belt may be considered seismogenetic, whereas the southern part of Puglia is substantially quiet and aseismic. The local tectonic setting of the Gargano is characterized by a system of surface and deep normal faults (see Ciaranfi et al., 1983), predominantly running from NW to SE, paralleling the Apennine chain. Southern Gargano is crossed by a right–lateral strike–slip fault, the so–called Carbonara fault, with antiapennine alignement, that has been well defined on the basis of the instrumental seismicity (Favali, 1978). Earthquake mechanisms of strike–slip type as well as of dip–slip type have been found from first–arrival analysis consistently with the above picture of the region (see Gasparini et al., 1985). The focal depth of the earthquakes ranges from surface to the deep crust, although some determination exceeding 70 km depth seems scarcely credible.

THE EARTHQUAKE AND THE TSUNAMI

The analysis of the available historical sources has inevitably implied the joint revision of the tsunami and of the generating earthquake. The most recent catalog of the Italian earthquake (the PFG catalog edited by Postpischl, 1985) gives the intensity X–XI (MCS scale) to the shock, that occurred one hour before noon (GMT) on the 30th of July 1627. Our revision entailed a careful analysis of a large amount of several types of documents as may be seen from the Table I: administrative as well as ecclesiastical documents, local chronicles and annals, treatises, letters, accounts of damage surveys... There has been a strong effort to find the original sources (either published or unpublished), since only the original sources were proven to be reliable and free from the subsequent historiographic distortions. Most of the encountered difficulties resided in that the documents were spread out in many central and peripheral offices. To name only a few, we should mention the State Archives of Naples and Foggia, the documents filed in Vatican as well as in local episcopal archives (see for example Molfetta's) and even in parish churches, such as San Severo's, or in monastic institutions.

TABLE I – Type and number of sources utilized to study the 1627 Gargano earthquake and the following tsunami.

TYPE OF SOURCES	Absolute Frequency	Relative Frequency
Administrative Documentation	10	6
Chronicles/Annals	78	43
Ecclesiastical Documentation	27	15
Diaries/Private Deeds	3	2
Letters	5	3
Accounts/Relations	8	4
Epigraphs/Inscriptions	6	3
Treatises/Manuscripts	28	15
Earthquake Catalogs	11	6
Periodicals	4	3
Total	180	100

The most important work on the 1627 earthquake is the contemporary monography written by Antonio Lucchino (1628), who was eye–witness and circumstantial reporter of the damages caused in the town of San Severo and the neighbouring area and was read and appreciated by most of the successive investigators (see Baratta, 1894 and 1901). In our research, we were however able to find out references to the tsunami that neither directly nor indirectly had been taken into account in the previous literature. Among them it is worth recalling the coeve contributions by Del Vasto and by Ballerani (both written in 1627 a short time after the events) as well as the seismic catalog assembled in the same century by Da Secinara (1652).

The Figure 1 summarizes the results of our work. The isoseismal lines close to the epicenter are substantiated by a reliable amount of observations, whereas on the other extreme the line of intensity VI is somewhat uncertain. According to our judgement, the main shock, that was followed after 15 minutes by a violent aftershock, must be attributed the intensity XI (in MCS scale) with macroseismic epicenter to NE of San Severo. The destruction of life and property was widespread, involving towns and villages as well as solitary houses and small hamlets. The reported victims were more than 5,000. As to the tsunami, there is a number of witnesses' accounts regarding the effects on the northern coasts of Gargano close to the lake of Lesina. The lake, that is approximately 20 km long and 4 km large, is nearly a lagoon, connected to the sea by means of two very narrow channels.

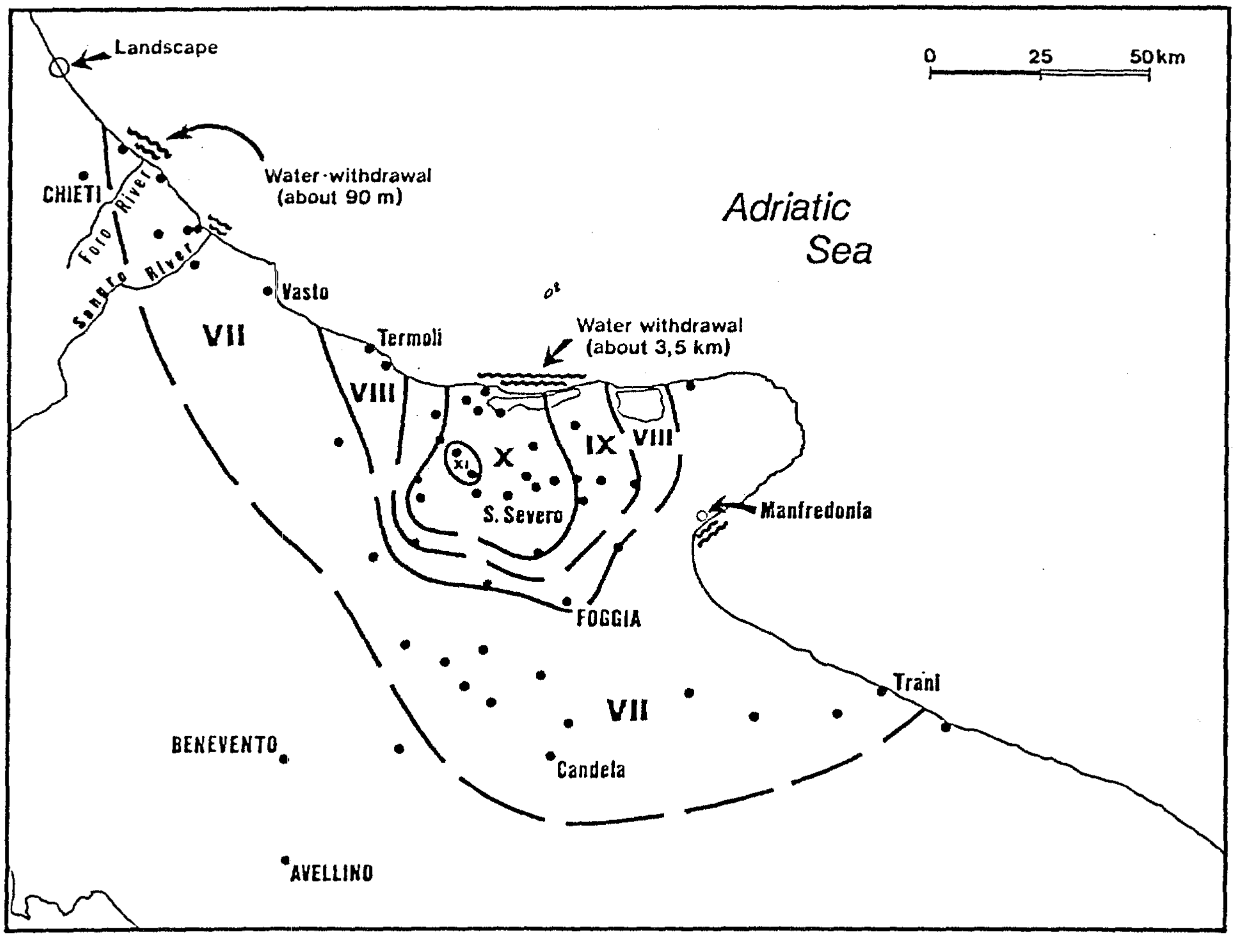

Figure 1. Isoseismal map of the 1627 earthquake according to the present research. Black circles denote the points where macroseismic observations are available. The external intensity VI isoline is rather uncertain and it is drawn only for completeness. The waves along the coasts denote the shorelines where tsunami effects are reported.

From the sources we know that the sea water first withdrew nearly two miles from the shoreline between the places called Fortore and San Nicandro, that the lake was devoid of water for a while and that the water invaded afterward the land for some length, inundating the country and the village of Lesina (see e.g. Del Vasto, 1627 and De Poardi, 1627). Lesina was a little and poor fishing settlement in those times and it is practically impossible to discriminate the destruction due to the earthquake from the effects of the tsunami. As to the estimate of the height of the waves, even considering the mild slope of the bottom in the coastal region facing the lake of Lesina, the waves should have been remarkable and in the order of a few meters. Tsunami effects involved a very large area. The coastal town of Termoli is reported to *"collapse"* into the sea, with a hyperbolic expression stressing the tragedy and the size of the event (Ballerani, 1627). A water withdrawal of about 90 m was observed at the mouth of Saro (identifiable with the river now called Foro), some 100 km to the North of the source region (Antinori, 1782). Da Secinara (1652) reports that further away to the North a mountain collapsed into the sea close to Montignano and was completely submerged by the water. Of difficult explanation are the high waves observed in the Gulf of Manfredonia on the southern side of Gargano, i.e. in a zone that should have been virtually sheltered by the promontary itself. The town of Manfredonia, placed on the coast, suffered very negligible damages from the shock (see Figure 1), but the sea is reported to have invested the town walls up to the half of their height (see Cerqua, 1627), which implies a run up in the order of 2–3 m.

In order to attempt the interpretation of the event, recourse must be made to geology and seismology. The tectonic setting of Gargano is not simple. Normal faults predominate to the North, reflecting the tensional regime characterizing the stress field in the area (see Cristofolini et al., 1985). Our interpretation is that the 1627 earthquake was a shock associated to a dislocation on a normal fault with strike nearly parallel to the coast, some 20 km inshore. The local magnitude corresponding to intensity XI is M_L =6.3, according to the intensity–magnitude relationships assessed for the region (see Tinti et al., 1987). The fault length should have been in the order of 20 km, the northern block experiencing a downlift. As to the slip on the fault surface, the data are not sufficient for any evaluation. The above picture is consistent with the isoseimal pattern, with the available tsunami data, with the whole regional setting and with the recent seismological data.

The final consideration is on the tsunami risk. In the 17th century, the shallow coasts near the lake of Lesina were almost uninhabited. Most people lived more inshore, which gives the explanation of why the earthquake impressed them much more than the tsunami. In the present days, however, many things have changed dramatically and the shallow coasts in the northern Gargano as well as the little bays in the rocky eastern part of the promontory have been transformed by several permanent settlements and by seasonal touristic villages and campings. And therefore a tsunami with the same size as the one occurred in 1627 could potentially cause a high number of victims and great destruction. This is one of the reasons that encourages us to continue our investigations on the historical earthquakes occurred in Gargano.

ACKNOWLEDGMENTS

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REVISION OF THE TSUNAMIS OCCURRED IN 1783 IN CALABRIA AND SICILY (ITALY)

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ABSTRACT

In 1783 Calabria in Southern Italy was shaken by the most violent and persistent seismic crisis that occurred in the last two thousend years. In the short interval of time from February up to the end of March, five earthquakes of intensity $I_0 \geq$IX (MCS) occurred south of Catanzaro, determining damages and destruction in a vast area embracing the whole Southern Calabria. The earthquakes gave origin to tsunamis. The most relevant ones are the tsunami ensuing the 5 February I_0 =XI earthquake and the tsunami occurred in the night between the 6th and the 7th of February. The latter was very likely generated by the sudden collapse of the flank of the mountain called Capo Pací down into the sea, after a series of small and intermediate–size local earthquakes. The waves inundated the shore of Scilla, a little Calabrian village a few miles to the north of the upper end of the Messina Straits, where they killed more than 1,500 people. The present work is a contribution to the reconstruction of the events, performed by means of a rigorous analysis of the original sources. We devoted our attention also to the study of the minor tsunamis generated during the seismic period, in order to give a more complete view of the facts and to provide a more sound basis for assessing the tsunamigenic potential of the region.

INTRODUCTION

The Calabrian Arc is one of the Italian regions most subject to investigations, since it is the center of a very intense geodynamic activity essentially driven by the continental collision between the African and the Euro–Asiatic plates (see Jackson and McKenzie, 1988; Lavecchia, 1988). From the analysis of the Italian catalog of earthquakes, it may be seen that the region is characterized by extremely active seismogenic structures (see Tinti and Mulargia, 1985 and Mulargia et al., 1987). The high level of seismicity combined with the narrowness and the length of the peninsula–like land of Calabria, explains the large tsunamigenic potential of the area (see the catalog of the Italian tsunamis by Caputo and Faita, 1984). The seismic period that started in 1783 is the most severe crisis in the seismic history of Calabria. It was preluded by several local foreshocks since 1780 (see Baratta, 1901) and according to recent studies, by forerunners occurred on the greek margin of the Adriatic microplate in 1781 (see Mantovani et al., 1986). The whole Southern Calabria was affected by the sequence. The earthquakes hit an area of very large extent with devastating effects. Several towns and villages were fully or almost completely destroyed and the territory experienced permanent geomorphological changes: ground fissures, local rise and fall in the land level, landslides, obstructions of river with consequent formation of new small lakes... (see Cotecchia et al., 1969). The documentation regarding the earthquakes is incredibly abundant, which might be the reason why so far no attempt was made for a thourough and complete analysis of the sources. We were able to consult about one thousend documents of various types: administrative and ecclesiastical documentation, notarial deeds, letters, accounts, chronicles, annals and periodicals... Most precious, however, for our research were the several monographs, that were written by scholars soon after the events. Some of them, hundreds of pages long, contain very detailed accounts of observations, inquiries, conversations, measurements made throughout the area where the earthquakes were felt.

THE EARTHQUAKES

The most relevant earthquakes occurred in the first two months of the seismic sequence, but it took an interval of several years for the regional seismicity to fall down to the ordinary background level of activity: aftershocks, mostly in bursts and swarms, were felt almost everywhere in Southern Calabria (Baratta, 1901). In Table I the major earthquakes are listed together with the focal coordinates and the MCS intensity, according to the PFG Italian seismic catalog (Postpischl, 1985). The solid circles in the first column mark all the earthquakes that were object of particularly accurate investigation in the present work. The last two columns refer as well to our analysis and indicate respectively the earthquake intensity and whether or not the shock was followed by a tsunami. As may be seen, our epicentral intensity evaluations differ slightly from the previous ones. Quite significant is only the discrepancy in the estimate

TABLE I – Major earthquakes of the 1783 seismic sequence. Solid circles denote the events analyzed in our work. Origine time and focal coordinates are taken form the most recent catalog of the Italian earthquakes, known as Progetto Finalizzato Geodinamica (PFG) catalog (see Postpischl, 1985). PFG epicentral intensities are compared with our estimates. In the last column tsunamigenic earthquakes are denoted with an asterisk.

		PFG SEISMIC CATALOG					THIS WORK	
	DATE	ORIGIN TIME(GMT)	LAT. (^{0}N)	LONG. (^{0}E)	DEPTH (km)	INT (MCS)	INT (MCS)	TSU
•	1783 Feb 5	12 15	38 20	16 00	13	XI	XI	*
•	1783 Feb 6	00 06	38 15	15 42	–	X	VIII–IX	*
•	1783 Feb 7	13 20	38 36	16 15	6	XI	X–XI	*
	1783 Feb 7	15 00	38 11	15 33	–	VIII		
•	1783 Mar 1	01 30	38 46	16 18	–	IX	IX	*
•	1783 Mar 28	00 16	38 50	16 30	18	X	XI	

regarding the second shock in the list for which we were unable to find evidences supporting the PFG–catalog I_0 =X estimate. We remark further that the 28 March earthquake is changed to an I_0 =XI earthquake after our analysis, becoming one of the largest events of the 1783 seismic series. Nevertheless, its importance for our discussion is very small, since there is no evidence of tsunami generation for this shock. Besides the assessing of the epicentral intensities, our investigation was devoted particularly to determine the whole macroseismic field associated to the various earthquakes. To illustrate our work, in Figures 1a and 1b we show the isoseismal lines of two of the shocks that were carefully examined. The 5 February I_0 =XI earthquake, that is displayed in Figure 1a, abruptly opened the seismic period and it is of special interest for our present purposes. The isoseismal curve I =XI relative to the most severe effects embraces an area considerably large, that is mostly elongated in the Apennine chain direction and that, however, inglobes to the south the coastal zone between Palmi and Bagnara. Seismic wave attenuation was much stronger eastward and the Ionian coast suffered comparatively much less damages than the other regions. The isolines terminate on the Tyrrhenian coasts, with the exception of the I =VII curve, for which observations available in the Aeolian islands and in the northeastern corn of Sicily allowed a reasonable continuation. The 7 February I_0 =X–XI earthquake shown in Figure 1b is characterized by a more isotropic pattern of the isoseismal lines, with the source region confined in the high terrains of the Apennine chain, some 40 km to the north of the epicenter of the earthquake previously discussed. The two shocks, though

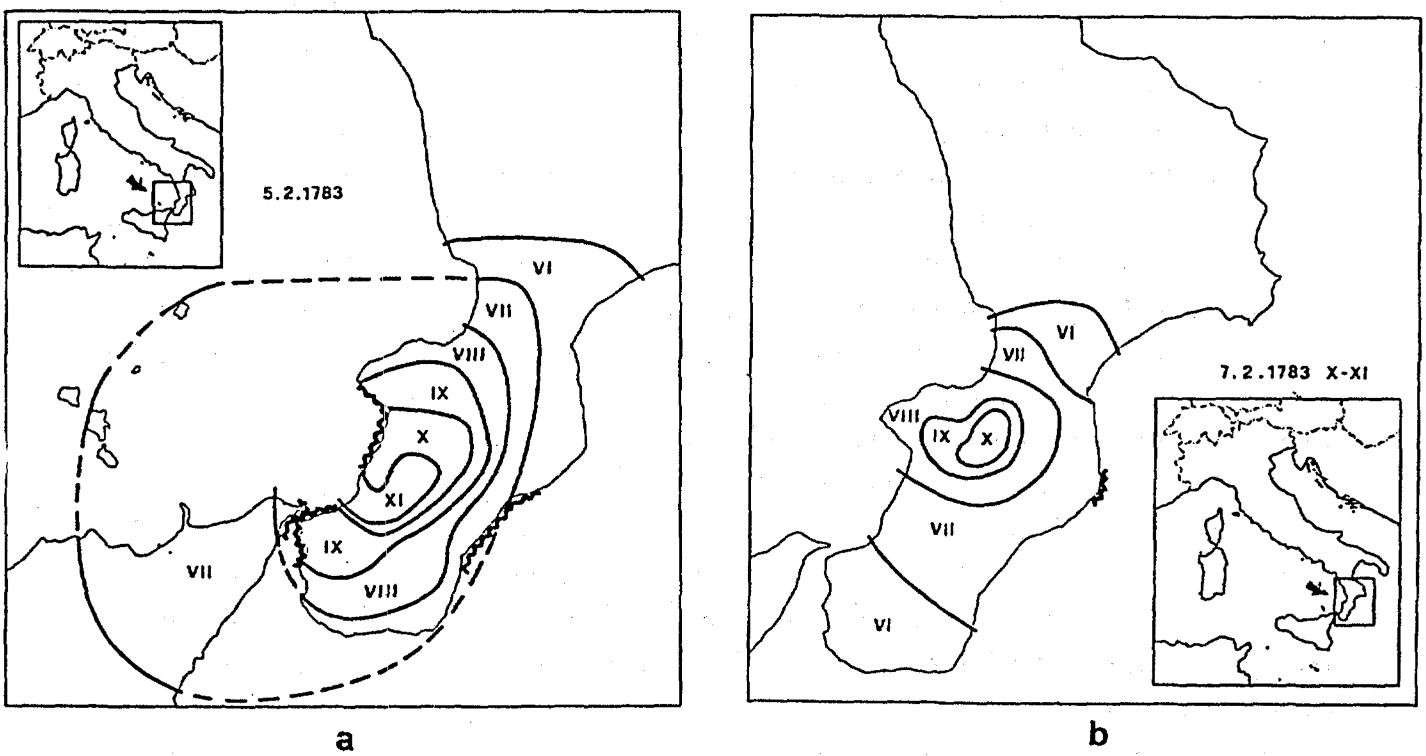

FIGURE 1. Isoseismal maps of the 5 February 1783 earthquake (a) and of the 7 February 1783 earthquake (b), as determined in this work. The coasts for which historical sources report tsunami effects are denoted by the wavy thick line.

close in time, occurred sufficiently away from each other so as to allow to discriminate quite clearly their macroseismic effects at least in the neighborhood of the epicentral areas. Some minor difficulties only arise for the determination of the peripheral I =VI isoline of the 7 February earthquake, because its southern branch crosses a territory experiencing continuous vibrations since the beginning of the crisis (see Figs. 1). In both figures the wavy lines immediately offshore denote the coastline where some information on the tsunami was reported, as we will see in the next section.

THE TSUNAMIS

In the Italian tsunami catalog by Caputo and Faita (1984), there are 8 events pertaining to the seismic period started in 1783, of which 5 occurred in the first half of the year and 3 in January 1784. Only three of the major earthquakes listed in Table I were associated to tsunamis, all other cases being related to very-small-size seismic events. Our revision changes somewhat the global picture: 4 entries should be definitely deleted from the catalog, since they refer either to sea water agitation of meteorological origin or to anomalous behaviour of the water in rivers very far from the sea; 2 new insertions are proposed, concerning the tsunamis following the 7 February earthquake shown in Figure 1b and the 1 March earthquake, included in Table I (see Galimi, 1783 and Vivenzio, 1788); other 2 episodes in the tsunami catalog seem quite dubiuos, but since our analysis could not find any conclusive answer, we prefer to consider the tsunamis as possible cases, though not certain.

The most important tsunamis generated in 1783 seismic period are by far the one following the 5 February earthquake and that occurred during the night between the 6th and the 7th of February. The macroseismic observations and the recent geological studies (Tortorici et al., 1986), consistently with the structural setting of the area, suggest that the 5 February shock was a normal fault shallow earthquake. The fault plane may be imagined to have width and length in the order of 20–30 km, to be parallel with the Apennine chain and to dip with a small angle towards the Tyrrhenian Sea. The earthquake caused the western block of the fault to slip downward, the displacement certainly involving also the sea bottom close to the coastline. The information available about the tsunami is really valuable, regarding many aspects of the phenomenon. The tsunami was observed in the Gulf of Gioia, in the Messina Straits and even at Roccella Ionica on the eastern Calabria coasts, where it was probably caused by submarine landslides triggered by the earthquake. The tsunami was strong, but not disastrous, since it did not determine heavy damages or victims. The water was generally reported to first recede from the coastline and then to inundate the shore: recessions and inundations repeated at least three times at intervals of about 10 minutes, the wave amplitude being in the order of 1–3 m (see e.g. Sarconi, 1784; De Lorenzo, 1895).

The 6 February tsunami was tremendously disastrous because of the very high number of people killed: most victims were inhabitants of Scilla, a little calabrian village close to the northern entrance of the Messina Straits. The 6 February I_0 =VIII–IX earthquake listed in Table I was very likely not responsible of the tsunami. In the epicentral area, the first waves were observed about half an hour later than the ground shaking (see Vivenzio, 1788), that is a delay too large to support the hypothesis of the seismic origin. The tsunami was probably determined by a huge rockfall, triggered by the earthquake: a portion of the Mt.Pací that is a promontory bordering the western side of the largest beach of Scilla, called Marina Grande, collapsed suddenly into the sea: a large volume of matter was involved in the rockfall and the local geomorphology of the coastline was dramatically changed (see Minasi, 1783 and 1785; Sarconi, 1784). The tragedy was augmented by the circumstance that most of the population of Scilla, deadly frightened by the terrible series of the earthquakes, had escaped from their houses to the open beach: at the night time many took refuge in Marina Grande, others in the close beaches of Marina di Chianolea and Oliveto, both sheltered by the little promontory of the Scilla castle. The waves originated by the rockfall swept with extreme violence

Marina Grande, a strip of sand about 800 m long and 100 m wide, where very few survived. The total number of victims is difficult to ascertain, but it surely exceeded 1,500 (see Mercalli, 1906). The description of what happened is very detailed, as the accounts of numerous direct witnesses are available in the historical sources. Sarconi (1784) was even able to draw a precious topographic map of Scilla to help the reader to follow his narration. Inundation heights in the range of 6–9 m were observed in Marina Grande, that is the place most affected by the tsunami. Reports insist on the quick sequence of three waves investing the shore, the third of which was the strongest and the most destructive. The region affected by the tsunami seems to be rather limited around the source. At Torre del Faro, located in Sicily on the extreme head of the promontory facing Scilla, waves 6 m high caused damages and victims. However, there are only episodic accounts of the waves in the other coasts of the Messina Straits and in the Tyrrhenian Sea, which confirms the very local nature of the tsunami.

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THE TSUNAMI GENERATED FROM THE ERUPTION OF THE VOLCANO OF SANTORIN IN THE BRONZE AGE

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ABSTRACT

A tsunami source mechanism resulting from the collapse of the cone of the Santorin Volcano in forming a submarine caldera following the last violent eruptive phase, and augmented by additional large scale crustal movements, oscillatory vibrations, and atmospheric shock waves, can account for the generation of the catastrophic sea waves observed in the Aegean Archipelago and the Eastern Mediterranean in the Bronze Age. The tsunami and destruction resulting from the explosion and collapse of the volcano of Santorin are documented in this study. It is believed that the final caldera collapse resulted from a large earthquake along the NE-SW trending normal fault along which the Santorin Volcanic field has developed. On the basis of recent geological evidence, a source mechanism is proposed which can account for the size and destructiveness of this Bronze Age Tsunami with interesting archaeological and historical implications.

INTRODUCTION

Numerous large destructive earthquakes and tsunamis have occurred from antiquity to present in the Eastern Mediterranean Seas and particularly in the Agean Archipelago. There is historical evidence that such large destructive earthquakes and tsunamis have ravaged the Aegean islands and the Greek mainland resulting in extensive destruction of the Minoan and early Greek settlements (Pararas-Carayannis, 1973). Of a total 613 known historic earthquakes, at least 41 major events generated documented tsunamis that struck the coasts of Greece. Sixteen of these resulted in really damaging or disastrous tsunamis. Between 1801 and 1958, 482 earthquakes with intensity equal or greater than VI, and 170 with intensity greater than VIII have occurred. Twenty of these earthquakes resulted in tsunamis, and six of these tsunamis were particularly damaging or disastrous in the Aegean and the Eastern Mediterranean Sea (Galanopoulos, 1960).

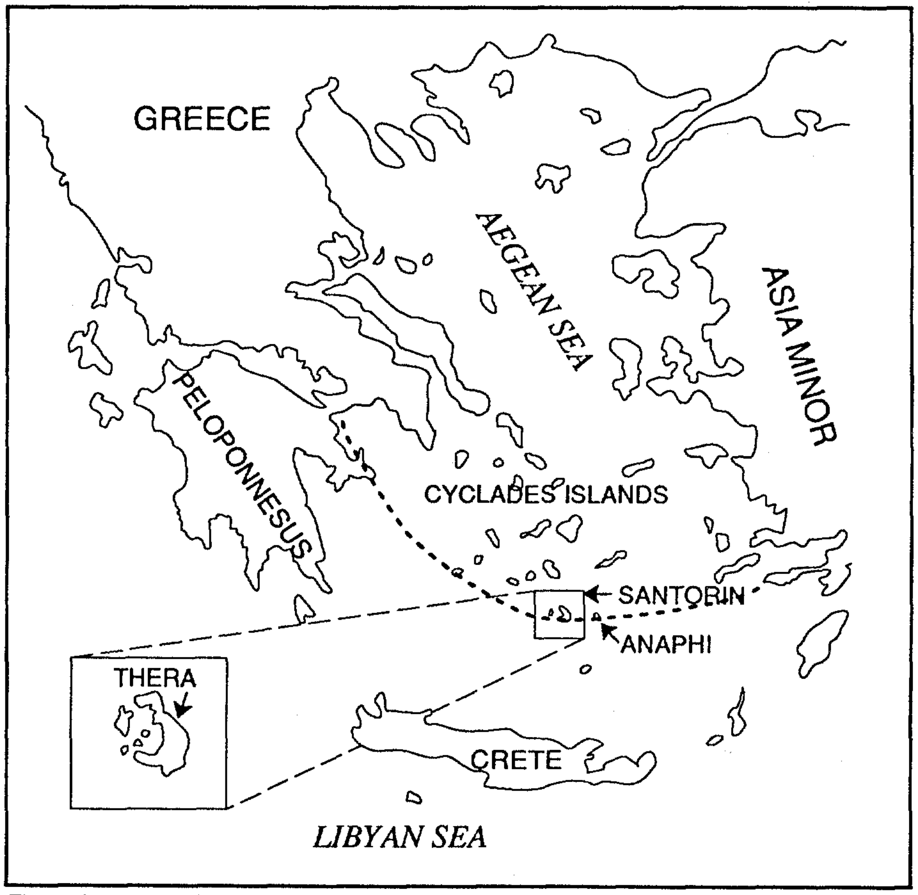

Fig. 1 Location Map showing Thira (Santorin) and the Cyclades Volcanic chain (dotted line).

Thus, the occurrence of large tsunamis is quite usual for the Eastern Mediterranean. Most of the destructive tsunamis have originated from a region of the Hellenic arc near the island of Santorin. At 365 AD, a destructive tsunami struck the Island of Crete and was reported as far as Alexandria, where ships were carried inland and left in the streets of the city. On 26 September 1650, a destructive earthquake was accompanied by a submarine explosion from the Columbo Volcano, whose crater lies in the sea on the northeast of the island of Santorin. There was a devastating tsunami observed on the island of Ios, and waves of up to 16 meters were reported. In 1672, the islands of Cyclades, and particularly Santorin, were again shaken by an earthquake. The island of Kos was reported to have been swallowed up presumably by the resulting tsunami. The best documented and most recent tsunamigenic earthquake is the one that occurred on 9 July 1956 near the southwest coast of the island of Amorgos, killing 53 people, injuring 100 and destroying hundreds of houses (Galanopoulos 1957). The waves were particularly high on the south coast of Amorgos and on the north coast of the island of Astypalaea. At these two places the reported heights of the tsunami were 25 and 20 meters respectively (Galanopoulos 1960).

Of all the historical tsunamis in this area, the best known but least documented has been the one presumably associated with the explosion-collapse of the volcano of Santorin in 1490 B.C. There is a substantial and undisputed evidence that the final explosion of the Volcano of Santorin in the Bronze Age coincided with a catastrophic tsunami in the Aegean Archipelago and the Eastern Mediterranean. There is a great deal of speculation about the effects of this tsunami on the ancient world but very little is known about the tsunami source mechanism, the time history of events leading to tsunami generation, and finally, of the initial tsunami height, and the tsunami height distribution in the Eastern Mediterranean. All studies completed thus far, attribute this tsunami to the explosion-collapse of the volcano of Santorin in forming a large submarine caldera. Although large tsunamis can be generated by such an explosion-collapse mechanism, this particular event alone could not have generated the large destructive tsunami observed in the Aegean and Eastern Mediterranean Sea during that period. The reasons are provided in the subsequent sections. In addition to understanding volcanic tsunami generation, careful examination and analysis of this event has important historical and archaeological implications. In this paper, such an analysis and examination of the Bronze Age tsunami is undertaken, particularly on the basis of recent geological and geophysical findings.

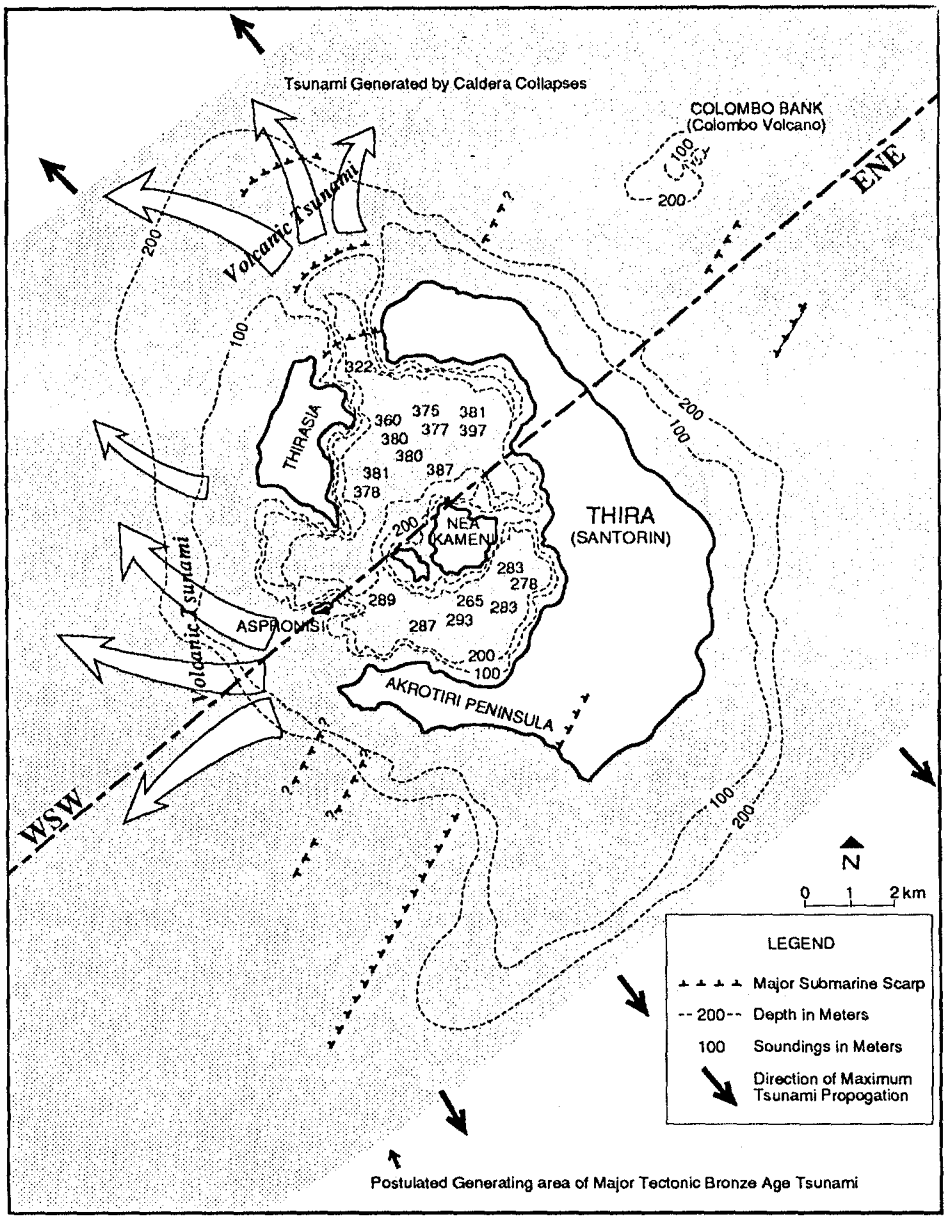

Fig. 2 Map of the Thira (Santorin) volcanic field, indicating source areas of volcanic tsunamis and postulated generating area of major tectonic tsunami.

Source Mechanism of Volcanically-Generated Tsunamis

To understand the Bronze Age tsunami in the Aegean, the source mechanism of volcanically-generated tsunamis must be examined. Marine volcanic eruptions of the Krakatoan variety, when associated with explosion-collapse processes in forming submarine calderas, are very efficient tsunami generators. However, volcanic eruptions affect relatively small portions of the sea floor and waves generated are very catastrophic locally, their energies dissipating rapidly with distance. Tsunami generation mechanism from volcanic sources is a complex phenomenon involving, not only the explosion and collapse of the volcanic cone, coupling of atmospheric shock waves, but possibly, in some instances, larger tectonic movements of the adjacent sea floor.

In order to understand the volcanic tsunami generation mechanism, we must examine caldera formation processes and all other related concurrent geotectonic activity.

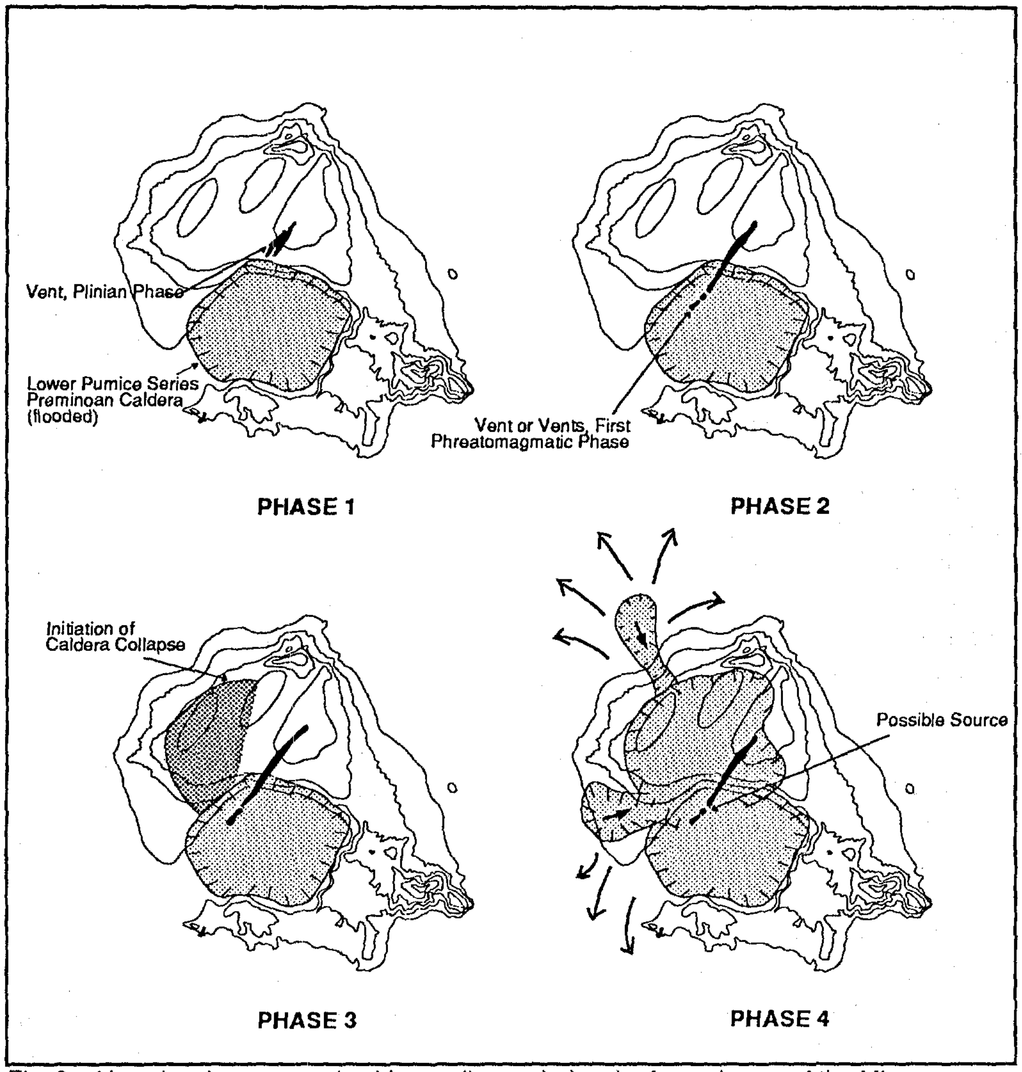

Fig. 3 Vent development and caldera collapse during the four phases of the Minoan eruption of the Volcano of Santorin. (after Heiken and McCoy, 1984)

It is generally accepted that volcanic calderas of the Krakatoan type are formed by the engulfment of the unsupported upper volcanic cone into the drained magnetic chambers below. However, this theory lacks detail and is somewhat contradictory to evidence. It has been observed that the volume of the ejected pumice and other pyroclastic debris is often considerably less than the volume of the caldera depression. Therefore, the volume discrepancy suggests a possible mechanism for the explosive removal of the upper volcanic cone rather than its total engulfment, or perhaps a combination of the two processes. This in turn is related to tsunamigenic efficiency.

Also, it is not known with certainty whether the volcanic collapse phase is sudden and total, or periodic and partial. The time-history of the caldera collapse and the geometry of the collapse are very important to know for understanding tsunami source mechanism. Similarly, the triggering mechanism of the last violent paroxysmal eruptive phase of the volcano is doubtful. It is not known with certainty if this phase is purely hydromagmatic in origin, or the result of extreme gaseous pressures building below high viscosity magmatic residues. Furthermore, certain other large depressions that resemble Krakatoan calderas, though associated with regional volcanic activity, result from tectonic subsidence along fractures controlled by regional fault patterns. Such fault patterns may be localized along ring fractures and may indeed form circular caldera-like depressions, but may also be associated with larger tectonic displacements along major fracture zones. This may have been the case for the volcano of Santorin. Large tectonic displacements caused by a large earthquake in the area could have caused the final collapse of the volcano of Santorin, the formation of the large caldera, and the destructive Bronze Age tsunami or tsunamis.

Source Mechanism of the Bronze Age Tsunami

It is not known how high were the waves of the Bronze Age tsunami, but researchers have found evidence of tsunami inundation on the island of Anaphi, 25 Km east of the island of Santorin, ranging from 40 to 50 meters above sea level on the west side of the island. Other evidence was presumably found as high as 160 m and 260 meters on the northeastern side of the same island. At greater distance away from Santorin evidence of the tsunami height was found at a height of 5 meters above sea level north or Jaffa-Tel Aviv. This tsunami height, if corrected for eustatic change in sea level in the Mediterranean for the last 3 1/2 thousand years, would have been at least 7 meters.

Could the tsunami generated by the explosion of Santorin, or any tsunami for that matter, have been as high as 160 and 260 meters on the northeastern side of the island of Anaphi? These values appear to be extremely unreasonable for either tectonically or volcanically generated tsunamis on any open coastline. An evaluation of the tsunami source mechanism is necessary. Such an evaluation of the tsunami source mechanism can be undertaken, particularly in view of recent studies of the Santorin Caldera formation (Heiken and McCoy, 1984), and other geophysical measurements and observations.

There is evidence of a NE-SW trending fault within the Aegean plate which is consistent with a similarly oriented graben along which the Santorin volcanic field has developed (Figure 2). Eruptions of Santorin have occurred from fissures located within this graben. Furthermore, according to Heiken and McCoy (1984), there is evidence of a much older flooded pre-Minoan caldera present on the southern half of the volcanic field of Santorin before the 1490 B.C. eruption. This caldera was approximately 5-6 Km in diameter. The original depth of this caldera is not known, but its present average depth is 280 meters. The presence of this pre-Minoan caldera would reduce the volume of the Minoan caldera to 19 Km^3 which is reasonably close to the volume of magma erupted during the Minoan eruption as estimated by Watkins et.al (1978). Furthermore, according to Heiken and McCoy, the vent development and caldera collapse occurred in four phases, (Figure 3). Thus, a long-time history of caldera collapse is inferred.

The sequence of events was as follows: A vent was developed in the first phase. When the subsequent eruptions extended the vent into the flooded caldera, a series of large phreatomagmatic eruptions occurred which widened the vent and increased the volume of the emitted lithic fragments. According to Heiken and McCoy, it was not until the third and more massive phreatomagmatic phase, that collapse of the caldera began with subsidence in the western part of the island. Finally, eruptions in the fourth phase, although with a phreatomagmatic component, were mostly from subaerial vents in the east. Finally, after this phase, the final collapse extended from west to east and resulted in the Minoan caldera of approximately 8 x 9 Km with a present depth averaging 380 below sea level.

Based on such a time-history of events, it is highly improbable that such caldera formation could have generated tsunami waves conceivably of the size that have been suggested in the literature. The caldera collapse process would have been rather slow for large tsunami generation. Furthermore, the geometry of the collapse would not have allowed the generation of extremely high waves. Tsunami waves could have been generated only from the two openings that were formed in the northwest and in the west of the present Santorin island. Undoubtedly, several smaller tsunamis over a period of time could have been generated from the periodic caldera collapse. The fact that the fourth eruptive phase was primarily from subaerial vents, (as indicated from the emission of ignimbrites as shown by Heiken and McCoy) with only a small phreatomagmatic component, indicates that the bulk of the Santorin volcano was still up. Thus, a different source mechanism for the final caldera formation and tsunami generation must be found. What triggered the final caldera collapse and the largest of the Bronze Age tsunamis, could not have been a phreatomagmatic eruption, but some other geophysical event. Such an event could have been a large earthquake with an epicenter close to Santorin island along the same NE-SW trending fault that has generated the 1650, the 1672 and the 1956 tsunamis. Such an earthquake may have been responsible for triggering the final Santorin volcanic caldera collapse and can account for the large tsunami that resulted.

A tsunami source mechanism is proposed here resulting from the collapse of the volcanic cone in forming a submarine caldera, following the last violent eruptive phase, but augmented by additional larger scale crustal movements, oscillatory vibrations and atmospheric shock waves. Only such a mechanism can account for the generation of the extensive catastrophic sea waves documented for the Aegean Sea following the eruption and collapse of the volcano of Santorin in 1490 B.C.

There is substantial evidence that on the island of Crete large earthquakes destroyed the Minoan palaces of the island, including Knossos, throughout the life of this Kingdom. The first major destruction of the Palace of Knossos by earthquakes occurred around 1720 B.C. After the palace was rebuilt and restored to its original splendor, it was again destroyed by the earthquakes of the fourteenth century B.C. (Pararas-Carayannis, 1974). So, evidence of large destructive earthquakes exists and Sir Arthur Evans, in his excavations at Knossos, verified that. Specifically, he found many Minoan houses ruined by huge blocks of rock, displaced as much as 6 meters from their original positions (Pararas-Carayannis, 1973). Only an earthquake of great magnitude could accomplish this. Such an earthquake could have been associated only with the seismic zone of the convex side of the Hellenic arc (Hellenic Trench), which is an established tsunamigenic region (Papazachos et.al, 1985, 1986).

Height of the Bronze Age Tsunami

There is no doubt that a number of tsunamis were generated from the gradual collapse of the Santorin volcano over a period of time. Also, there is no doubt that a much larger tsunami was generated that acted as the catalyst in the declination of the Minoan civilization (Pararas-Carayannis, 1972). There is conclusive evidence that Minoan cities on the north and east coast of the island of Crete were struck by huge tsunami waves (Marinatos, 1939). These included Amnisos, Malia, Niroun Chani, Psira, Ghournia and Zakros. Nothing is definitely known about the tsunami on other Aegean Islands. However, a rough estimate of the Santorin tsunami at Anaphi island, the closest to the origin, can be extrapolated from the 7 m. tsunami (corrected for eustatic change), as documented at Jaffa-Tel Aviv, 900 Km away. An estimate of the tsunami height at Anaphi island can be obtained on the basis of tsunami height attenuation at Jaffa-Tel Aviv assuming entirely geometrical dispersion and not effects of refraction, diffraction, or resonance. Based on such a calculation, the height of the Bronze Age tsunami at Anaphi island was estimated to be 41.9 meters, which is somewhat consistent with the 40-50 meter elevation at which pumice deposits were found by Marinos and Melidonis. Such pumice deposits were found at 350 meters inland on the west coast of the island. Also, the west coast of Anaphi island would have been closer to the tsunami source area and would have experienced the highest tsunami waves.

The estimate of 42 meters at Anaphi appears to be reasonable. The highest possible tsunami wave at the source could not have exceeded 50 meters. The pumice deposits found on the northeastern side of the island of Anaphi, at 160 m and at 250 m could not have been carried by tsunami waves of any eruption or of any other earthquake. These values are simply too high. Furthermore, even a 50 meter tsunami cannot be supported by a tsunami source mechanism which involves only the explosion-collapse of the Santorin Volcano even on the shortest time scale. The time-history of events and the geometry and orientation of such a tsunamigenic source could not explain a large tsunami of 40-50 meters at Anaphi Island, near Santorin. Only a mechanism which involved caldera collapse in combination with a much larger scale tectonic crustal movements along the prevalent NE-SW trending fault caused by a large earthquake can account for the generation of the largest of the Bronze Age tsunamis.

Conclusions

Based on the above evaluation the following conclusions can be reached:

1. The time history of caldera formation indicated by Heiken and McCoy is too slow for large tsunami generation. Thus, a different tsunami source mechanism must have been responsible for the extreme tsunami observed in the Aegean and Eastern Mediterranean Sea in the Bronze Age.
2. Following several paroxysmal eruptive phases and partial caldera collapses, final Santorin caldera collapse was triggered possibly by tectonic subsidence resulting from a large earthquake along the NE-SW trending normal fault which has also formed the graben along which the Santorin volcanic field has developed.
3. Several tsunamis must have occurred with different source mechanisms. The gradual collapse of the western portion of the caldera must have generated several smaller tsunamis originating at the northwest and west opening of the Santorin caldera. The several paroxysmal phases and atmospheric shock waves probably generated other small tsunamis.
4. The collapse of the remainder of the volcano into the empty magmatic chambers possibly triggered by a large earthquake, generated a larger tsunami at the northwest and west openings which may have been destructive in adjacent islands, in Crete and elsewhere.
5. The much larger tsunami, the Bronze Age tsunami, was generated by a combination of normal faulting resulting from a suspected large tectonic earthquake and possible underwater landslides of unstable volcanic tuffs on the outer perimeter of the Santorin volcano. This event may have occurred concurrently as the tsunami generated by the collapse of the remaining volcano of Santorin, or at a different time.

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MATHEMATICAL SIMULATION OF TSUNAMI EXCITATION BY DISLOCATIONS OF OCEAN BOTTOM

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A B S T R A C T

Theoretical investigations of tsunami generation by underwater earthquakes are reviewed. The effects associated with form of transient bottom displacements, dispersion and nonlinearity of waves, density stratification of ocean are discussed.

INTRODUCTION

Three stages of tsunami development are usually distinguished: (a) formation of a localized initial disturbance and its evolution near the source; (b) propagation of waves in the open ocean; (c) propagation of waves in shallow water and on the shore. At each stage tsunami is described by its own mathematical model with initial conditions being found from previous stage of the dynamic process. Therefore tsunami forecasting at shore depends on precision of mathematical simulation of waves field at first stage. As a rule disturbance of ocean near the source during the earthquake is unknown.

Using a theory of "elastic" tsunami generation you can find the wave field at the stage (a) if parameters of earthquake source are known (Alexeev and Gusyakov, 1973). Within the framework of this model various situations are possible and special numerical experiments of studying initial disturbance formation near the source are necessary.

Mathematical modelling of waves generated by the known vertical displacements of ocean bottom is the simplest way of studying the initial stage of tsunami development. Setting different bottom deformations and sea-water properties in models one can carry out a comparison analysis of tsunami generation mechanisms and investigate the effects which accompany waves.

MECHANISMS OF TSUNAMI GENERATION DURING UNDERWATER EARTHQUAKES

The main part of tsunamis is excited by underwater earthquakes. About 90% of tsunamis on the west coast of the Pacific Ocean are bound up with underwater earthquakes (Soloviev and Go, 1974). Essential mechanisms of seismotectonic displacement transition to ocean are: (a) rapid upward or downward displacements of the ocean floor; (b) strong elastic displacements or oscillations of the ocean floor; (c) the abrupt horizontal shifts of steep underwater slopes or strong horizontal seismic impulses transmitted through a vertical or inclined wall. The main mechanism of tsunami generation is the first one.

The available data on the two most throughly investigated tsunamis, the Alaskan tsunami of 27 March 1964 (Hwang and Divoky, 1970) and Niigata tsunami of 16 June 1964 (Soloviev and Militeev, 1967), support the idea that "piston-like" movements of ocean floor are the main mechanism of tsunami excition. In both cases there was upheaval of extended parts of ocean bottom, by 3-10m and 2-5m respectively.

Many weak tsunamis result from strong elastic displacements or bottom oscillations. Such mechanism of excitation must have taken place on the 15 of June 1911 during an earthquake on Ryukyu Islands (Soloviev and Go, 1974).

Effective generation of tsunami by propagating envelope of seismic waves is possible if parameters of gravity water waves and elastic waves are close to each other (Belokon et al., 1986). However, as Belokon et al. (1986) have shown, in frame of linear model developed by Podyapolskij (1978) such resonance excitation of tsunami is not possible for periods $10^2 - 10^3$ s.

Iida (1963), Watanabe (1964), Hatori (1966), Soloviev (1968), Iwasaki (1977), Pelinovskii (1982) studied the correlation between parameter of tsunami generation region and the magnitude of tsunamigenic earthquake.

MATHEMATICAL MODELS OF TSUNAMI GENERATION

The linear problem of tsunami generation by small bottom deformations in homogeneous ocean of constant depth H has been studied most thoroughly. The two-dimensional linearized shallow-water non-dispersive fundamental equations are

$$\frac{\partial u}{\partial t} = -g\frac{\partial \zeta}{\partial x} + X, \quad \frac{\partial v}{\partial t} = -g\frac{\partial \zeta}{\partial y} + Y, \quad \frac{\partial \zeta}{\partial t} + H\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) = \frac{\partial h}{\partial t} \qquad (1)$$

with initial conditions

$$u = v = \zeta = 0 \qquad (t=0), \qquad (2)$$

where x, y are horizontal co-ordinates, t is time, u, v are the x and y components of fluid velocity, ζ represents the displacement of the water surface from the mean level, h is a small transient bottom displacement ($h=0$ for $t \le 0$, $|h| << H$), g the accelaration of gravity.

A horizontal body force (X , Y) in the area of tsunami excitation for modelling of horizontal impulse transfer to the water mass has been introduced by Voit et al. (1981). The above mentioned transfer is possible in the region of continental shelf.

If $X = Y = 0$, $h = h_0(x,y)q(t)$ ($q=0$ for $t \le 0$) the problem (1), (2) can be solved by Laplace and Fourier transformations. Finally, we obtain

$$\zeta = \frac{1}{4\pi^2} \iint\limits_{-\infty}^{+\infty} H_0(m,n) \left\{ \int\limits_0^t \cos\left[cz(t-\tau)\right] q'(\tau)d\tau \right\} e^{i(mx+ny)} dm\,dn, \qquad (3)$$

where H_0 is Fourier transformation of h_0, $c=\sqrt{gH}$, $z=\sqrt{m^2+n^2}$.

The total energy $W(t)$ of wave field is given by

$$W(t) = \frac{\rho}{4\pi^2} \iint\limits_{-\infty}^{+\infty} |H_0(m,n)|^2 \left[\int\limits_0^t q'(\tau)\Psi(z,\tau)d\tau \right] dm\,dn, \qquad (4)$$

where ρ is the density of fluid,

$$\Psi(z,\tau) = Hq''(\tau) + g\int\limits_0^\tau q'(\xi)\cos\left[cz(\tau-\xi)\right]d\xi,$$

Solutions in the form of integrals may be found in general linear statement of the problem and when body force (X , Y) is taken into account. The behaviour of waves for large distances and time $t \to \infty$ are studied by Gazaryan (1955), Kajiura (1963), Van den Driessche and Braddock (1972), Voit et al. (1981) and others.

Numerical calculations of integrals were carried out by Kajiura (1970), Selezov et al. (1982). Integrals (3) and (4) are calculated by Dotsenko and Sergeevsky (1985) for transient bottom deformation of a spacial form.

Numerical models allow to calculate two-dimensional tsunami waves when nonlinear terms of governing equations are taken into account (Hwang and Divoky, 1970; Marchuk et al., 1983). Numerical investigation of 'piston-like' tsunami generation with small nonlinearity and weak dispersion were carried out by Dorfman (1977).

To investigate physical effects accompanying tsunami generation and propagation the generalization of above-mentioned models is required. For example, the process of tsunami generation by small bottom deformations in baroclinic ocean of constant depth is decribed in hydrostatic approximation by the equations

$$\frac{\partial u}{\partial t} - lv = -\frac{1}{\rho_0}\frac{\partial p}{\partial x}, \qquad \frac{\partial v}{\partial t} + lu = -\frac{1}{\rho_0}\frac{\partial p}{\partial y} \qquad (5)$$

$$\frac{\partial p}{\partial z} + \rho_0 N^2 \zeta = 0, \qquad \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial^2 \zeta}{\partial z \partial t} = 0 \qquad (6)$$

with boundary and initial conditions

$$p = \rho_0 g \zeta \quad (z=0), \quad \zeta = h \quad (z=-H) \qquad (7)$$

$$u = v = \zeta = 0 \qquad (8)$$

where z is vertical co-ordinate, p is dynamic disturbances of pressure, ζ is the elevation of fluid particles from mean positions, $\rho_0(z)$ and $N(z)$ are the vertical distributions of undisturbed density and buoyance frequency of the fluid, l the Coriolis parameter.

The process of internal waves generation is close to excitation of waves by step-function ground motion with an initially undisturbed fluid ($q'(\tau) = \delta(\tau)$, δ is delta-function). The solution of the problem (5)-(8) for such case is given by

$$\zeta = \sum_{j=0}^{\infty} Q_j(x,y,t)\,Y_j(z) \qquad (9)$$

(Dotsenko, 1982, 1986), where Q_j is the same as integral (3) after replacing C by the phase speed of internal gravity waves for the mode number $j \in \overline{0,\infty}$. Modes $Y_j(z)$ may be numerically determined if buoyance frequency distribution is known. Surface waves correspond to barotropic mode (j =0).

TRANSIENT BOTTOM MOTION

Let us consider the generation of tsunami by a transient bottom motion of the form

$$h = h_0(R)q(t), \quad h_0 = a_e L^3 (R^2+L^2)^{-3/2} \quad (R=\sqrt{x^2+y^2}) \tag{10}$$

for two types of functions $q(t)$ shown in Fig.1.
They describe non-elastic (a) and elastic (b) vertical displacements of the ocean bottom. For bottom motion (10) integrals (3) and (4) may be calculated by analytical methods.

It will be assumed below that the radius of bottom deformation zone is equal to the distance $R=R_e$ for which $h_0(R_e)=0{,}1a_e$.

Processes of waves formation for two types of bottom motions are shown in Fig.1. Profiles of generated waves depend on the time history of ground motion due to an earthquake. For example, the amplitudes of leading waves for bottom motion (a) may be essentially larger than for (b). The amplitude of the wave in the case (b) depends considerably on the duration T of bottom deformations.

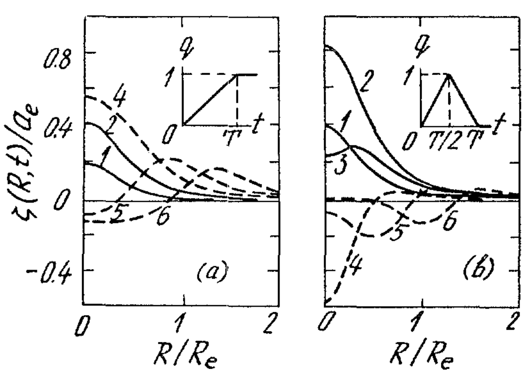

Fig.1. Processes of waves formation for 'non-elastic'(a) and 'elastic'(b) tsunami excitations. $H=4 \cdot 10^3$m, $R_e=5 \cdot 10^4$m, T=120s
1 - =24s, 2 - 60s, 3 - 84s, 4 - 120s, 5 - 240s, 6 - 360s

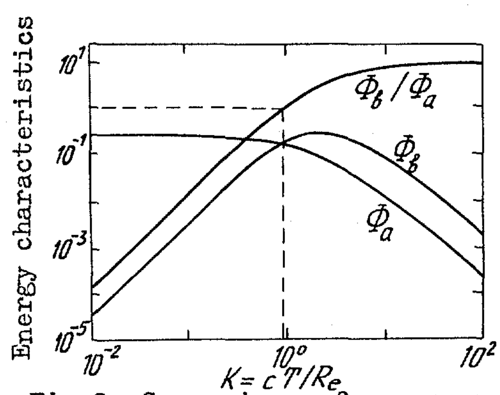

Fig.2. Comparison of energy characteristics of wave fields for 'non-elastic' and 'elastic' tsunami excitation

Total energies of wave fields $W(t)$ for 'non-elastic'(a) and 'elastic'(b) tsunami excitation are equal to

$$W_a = \pi \rho g a_e^2 L^2 \Phi_a(K), \quad W_б = \pi \rho g a_e^2 L^2 \Phi_б(K)$$

respectively, where $K=cT/R_e$. Variations of Φ_a, $\Phi_б$ and $\Phi_б/\Phi_a$ with K are shown in Fig.2. 'Non-elastic' excitation of tsunami is more effective than 'elastic' excitation if $K<0{,}96$. As it is known the most destructive tsunamis were preceded by visible changes of the ocean bottom after the shock (Soloviev and Go, 1974).

THE DIRECTIVITY OF WAVE RADIATION

This effect is typical for many tsunamis. The Chilean tsunami of 22 May 1960 is a bright example (Soloviev and Go, 1975). For the causes of the directivity, the following factors are considered to be most impotant: the effect of the shape of a source, the transfer of the impulse in the horizontal direction to the water mass, and large-scale variation of ocean depth.

For rectangular and elliptic shape tsunami generation zones in the uniform ocean the highest waves are radiated in the direction perpendicular to the major axis of the bottom deformation region (Kajiura, 1970; Marchuk et al., 1983; Dotsenko et al., 1986). Asymptotic laws of wave attenuation are similar for different directions. They are $R^{-1/2}$ for non-dispersive waves (Kajiura, 1970; Dotsenko et al., 1986) and R^{-1} for dispersive waves (Van den Driessche and Braddock, 1972).

Waves radiation directivity increases when the ocean has received a horizontal impulse during the earthquake. Using Voit's et al.(1981) model Sebekin (1986) showed that leading dispersive waves attenuate in the direction of horizontal impulse with distance as $R^{-\alpha}$ ($1/3 \leq \alpha \leq 1/2$); the decay is faster (R^{-1}) in other directions.

Bottom displacements $h = h(x+Ut,\ y)$, where U is the velocity of the movement of bottom disturbances, when $U>C$ generate directed waves of two types (Marchuk et al., 1983). The first type waves are similar to longitudinal ship waves. The angle between the crests of waves and the direction of bottom deformations movement equals to $\theta = \arcsin(c/U)$ (Novikova and Ostrovsky, 1978). The second type waves locate near the leading edge of moving bottom deformation area. The resonance tsunami excitation takes place at $U \approx c$.

EFFECTS OF DISPERSION AND NONLINEARITY

Wave dispersion increases with the time and when the radius of deformed area of bottom decreases. An oscillating 'tail' is formed following the leading wave, the latter is not always the highest one (Gazaryan, 1955).

The degree of dispersive transformations of a long axisymmetric wave depends on the parameters $\beta = R_e/H$ and $\tau = ct/H$. Kajiura (1970) showed that at $\beta \geq p\tau^{1/3}$ dispersive distortions are weak under the suitable value of constant p .

If the horizontal dimensions of a tsunami source are small enough the dispersion of waves may be essential both at $t \to \infty$ and at initial stage (Voit, 1987).

Pelinovskii (1982) showed that Urcell's parameter for spatial tsunami waves decreases at all stages of wave process. So spatial wave which is linear at its initial stage remains linear further . For description of waves in the open sea (the water depth is not less that 50 m) nonlinearity and dispersion are not essential (Soloviev, 1978; Voit, 1987).

Dorfman (1977) using nonlinear-dispersive numerical model of 'piston-like' tsunami excitation showed the possibility of wave transformation into a system of solitary waves when the bottom moves up. This system 'remembers' the amplitude of bottom displacement and R_e but 'forgets' the shape of bottom elevation. When the piston moves down, no solitary waves are formed. In case of one-dimensional waves the formed solitary waves depend weakly on the law of bottom deformation when the duration and the amplitude of ocean bottom displacement are the same (Hammack and Segur, 1974).

EFFECTS OF TRANSIENT BOTTOM DEFORMATION DURATION

The dependences of amplitude and energy of tsunami wave on deformation duration T were studied by Kajiura (1970), Selezov et al. (1982), Marchuk et al. (1983), Dotsenko and Sergeevsky (1985). The length of the leading tsunami wave increases according to quasi-linear law, the energy of the wave field decreases when T increases. The amplitude of excited gravity waves is maximum in case of impulsive bottom shifts.

EFFECTS OF BAROCLINITY

The excitation and propagation of tsunami are accompanied by processes of different nature. In baroclinic ocean of constant depth internal waves as well as surface waves are generated. For model (5)-(8) excited wave field is described by expression (9). Because of small velocity of propagation (1-2 m/s) internal waves cannot be used for tsunami warning.

Nevertheless effects of baroclinity and Earth's rotation present some interest for tsunami problem.

The influence of density stratification on surface tsunami waves is negligibly small (Cherkesov, 1970; Dotsenko, 1982, 1986).

The amplitudes of internal tsunami waves for lowest mode in Kuril-Kamchatka zone do not exceed the value of 0.72 $max\,|h|$. One cannot exclude a possibility of considering tsunamigenic zones as geographic region of internal waves generation (Dotsenko, 1982, 1986). The evolution of long internal wave is shown in Fig.3(a).

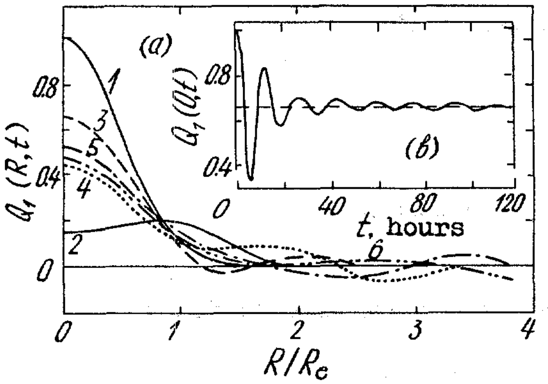

Fig.3. Process of impulsively generated internal wave radiation for lowest mode in Kuril-Kamchatka zone. $H=4\cdot10^3$m. (a) Profiles of wave. $R_e=5\cdot10^4$m, $t=0,5,10,18,26,36$ hours for curves 1-6 respectively. (b) Evolution of elevations in the origin of tsunami. $R_e=7.5\cdot10^4$m. Amplitudes of waves measured in relative values.

Due to stratification and Earth's rotation at 'piston-like' non-elastic excitation there forms a long-living geostrophic vortex in the region of tsunami generation (Dotsenko, 1982, 1986). The formation of vortex is accompanied by inertial oscillations of hydrophysical fields (Fig.3(b)). One cannot also exclude a possibility of using of an excited vortex to indicate tsunami sources.

During laboratory experiments of 'piston-like' excitation of plane internal waves in two-layer fluid a good agreement with the linear theory of potential waves has been obtained (Hammack, 1980).

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HISTORICAL AND RECENT TSUNAMIS IN THE EUROPEAN AREA

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ABSTRACT

A review of the main historical and recent tsunamis in the European area, specially those observed on the coasts of mainland Portugal and Azores, is presented, as well as an attempt for the explanation of the origin of some of them. In the Mediterranean Sea the main tsunamigenic zones are situated in the Aegean Sea and in its the central part on the coasts of Italy. Particular attention is given to the tsunami of 9 July 1956 in Eastern Mediterranean and to the tsunami of 28 Dec. 1908, which followed the Messina earthquake. Focal mechanism solutions of the earthquakes that generated those tsunamis are discussed. In the Atlantic Ocean the main tsunamigenic zone is situated off the Portuguese coast, southwestwards of St. Vincent cape, where were generated the tsunamis of 382 AD and 1755. A reference is made to some minor tsunamigenic zones, one of them in the continental margin at Galicia Bank, and others, far from it, in the abyssal plains.

In the Azores region tsunamis are generated by strike-slip earthquakes; they are local and very scarce.

INTRODUCTION

Tsunamis are not much frequent in the European area, although some coastal regions of the Mediterranean Sea and of the Atlantic are sometimes struck by those phenomena. The most prominent regions struck by tsunamis in the European area are the coasts of insular Greece, in Eastern Mediterranean and the coast of southern Italy, in central Mediterranean. In eastern Mediterranean, the region in the European area where more tsunamis have been observed, only about 20% of the tsunamis were really damaging ones, and this circumstance is attributed to the fact that most tsunamis in this area are generated by landslides.

In the Atlantic, Gorringe Bank area, situated off St. Vincent cape, is the most prominent tsunamigenic zone.

Nevertheless tsunami waves undergo normally a big attenuation both in the Mediterranean and in the Atlantic which is attributed to causes like water depth or slope of the sea bottom at the epicentral region and, above all, to their origin which is frequently attributed to landslides. Other causes like submarine block movements on the ocean bottom, volcanic eruptions and tsunami excitation by dislocations of the ocean bottom must, however, be considered.

TSUNAMIS IN THE MEDITERRANEAN SEA

The tsunami which followed the explosion of Thera volcano (Santorin island) in about 1500 BC is one of the most important tsunamis which occurred in the Mediterranean Sea. There are no reports indicating whether the collapse of the crater took place gradually or at once. If the collapse took place at once a huge tsunami should have been set off, even geater than that originated by the explosion of the Krakatoa volcano in 1883, because the crater originated by the explosion of the Thera volcano, as can presently be observed, is about 4 times greater than that of the Krakatoa. This tsunami appears to be responsible for the destruction of the Crete-Minoan civilization.

Another important tsunami in the Mediterranean Sea occurred on 21st July 365. Reports say that it followed an earthquake that destroyed several towns in Crete. It was very intense at Cirene, in Libya, and Alexandria, in Egypt, where the waves carried ships over the buildings. The coasts of Sicily and Calabria were hit by this tsunami. On the other hand there are reports that exactly on the same day, an intense tsunami hit the coast of southern Spain. At Malaga, a large withdrawal of the sea was observed; then the sea returned throwing the ships against the buildings and drowing many people. The Roman historian, Amiano Marcelino, contemporary of the event, connects this tsunami with an earthquake in southern Spain (Galbis Rodriguez, 1932). If these two tsunamis are not distinct, as it appears, it would be the only known tsunami that propagated across the Mediterranean from side to side without important attenuation of the waves. Anyway this tsunami had certainly tectonic origin.

Another important tsunami originated in eastern Mediterranean occurred on 9 July 1956. On this day a 7.8 magnitude earthquake with epicentre in the Aegean Sea originated a very strong tsunami. The waves reached 30 metres high in the epicentral region, but they attenuated rapidly and were recorded with small amplitudes on the Egyptan coast. The focal mechanism solution of the earthquake that generated this tsunami is consistent with strike-slip, whatever the nodal plane chosen as fault plane (McKenzie, 1972). Galanopoulos (1960) refers that the tsunami was set off by submarine landslides originated by the earthquake.

Important tsunamis have been generated on the coast of southern Italy. On 28 December 1908 a 7.5 magnitude earthquake destroyed Messina. This earthquake set off a tsunami with waves that reached 8 metres high in some bays along the Messina Strait (Schick, 1977). The submarine cable to the Lipari islands was broken and

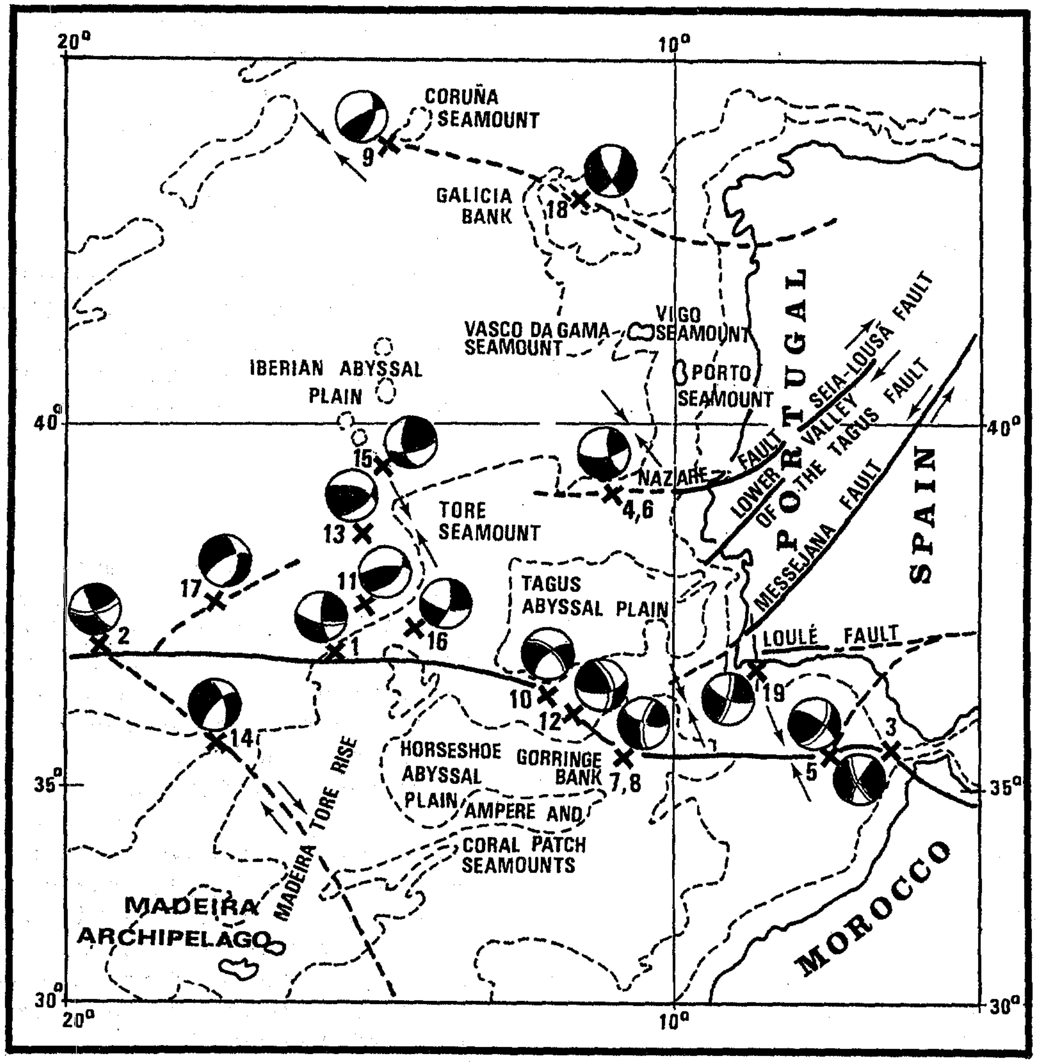

Figure 1. Focal mechanism solutions of earthquakes with epicentres in the Portuguese continental margin and its adjacente area.

the cable Malta-Zante failed. The waves reached the Egyptian coast with small amplitudes. The focal mechanism solution obtained by Schick (1977) is consistent with normal faulting showing a large amount of dip component, which is compatible with the assumption that block movements reflected on the topography of the sea bottom. On the other hand the rupture of submarine cables suggests a strong turbidity current set off by the landslides. This tsunami was probably set off by both causes - block movements due to normal faulting and landslides.

Another important tsunamis occurred in Italy. Among them can be referred the tsunami of 30 July 1627 which followed the Garganian earthquake, the epicentre of which is situated in land (Barata, 1901) and the tsunami set off by the earthquake of 6 February 1783, in southern Italy, which caused the collapse of Mount Paci into the sea, originating very high waves.

In western Mediterranean have been generated some small tsunamis related to earthquakes with epicentres in land, near the coast of Algeria. Focal mechanism solutions of the most important recent earthquakes which occurred in this area have shown that they are normally compatible with overthrusting. Earthquakes with epicentre in this area have caused turbidity currents connected with the generation of tsunamis. The Orleansville earthquake (9 Sept. 1954), caused a turbidity current which broke 5 submarine cables at a distance of 40 to 70 miles from the epicentre. The Oran earthquake (9 Oct. 1790) which killed 2000 people at Oran and destructions in southern Spain, originated a tsunami which reached the coast of Spain. At Cartagena, people working on the wharfs were obliged to flee. The Algiers earthquake (1773) originated waves 6 ft. high at Algiers which propagated in the Mediterranean. Due to the V-shaped coastline of the Alboran Sea and to the decreasing depth of the sea bottom when approaching the strait of Gibraltar the amplitudes of the waves increased and reached 30 ft high at Tangier.

TSUNAMIS IN THE ATLANTIC OCEAN

The area where is situated Gorringe Bank, west-southwestwards of St. Vincent cape (Portugal), is the most important tsunamigenic zone of Europe in terms of magnitude. The minimum depth of the ocean in this area is 25 metres and here was generated the well known tsunami of 1 Nov. 1755; on the southern coast of Portugal the waves reached more than 30 metres high, in some places, and penetrated more than 2 km in land in other places. The waves crossed the Atlantic and were observed in some places of the coasts of the American continent with considerable amplitudes. The epicentre of the earthquake of 28 Feb. 1969 is situated in the same area. It generated also a tsunami. This earthquake and the corresponding tsunami are considered to be reduced models of the earthquake and tsunami of 1 Nov. 1755. Its focal mechanism solution is consistent with thrust (Moreira, 1982; 1985) and the strike of the nodal plane taken as fault plane is in agreement with the fault trace of Messejana fault in land (Fig.1). Messejana fault appears to be responsible for the high magnitude earthquake and tsunami which occurred in 382 AD (Brito, 1597; Moreira, 1984). This earthquake and the corresponding tsunami originated the disappearance of two small islets which were localized near St. Vincent cape. These two islets are described in Strabo's Geography written about four centuries before this event. These tsunamis - 382 and 1755 - were probably generated by thrust faults, like the tsunami of 28 Feb. 1969, which leads to the assumption that block movements, with large throws, occurred on the sea bottom, although landslides took also place certainly. Another high intensity tsunami occurred on the Portuguese coast around 60 BC (Brito, 1597; Moreira, 1984) which probably was generated in the same area; there are also vague reports of a very intense tsunami which affected specially the northern coast of Portugal around 949 AD (Galbis Rodrigues, 1932).

Galicia Bank, where the minimun depth of the ocean is 492 metres is a minor tsunamigenic zone. On 20 June 1936 a 5.6 magnitude earthquake with epicentre in that area generated a small tsunami which was observed on the coast of Galicia.

Some minor tsunamis have been generated by earthquakes, with epicentres far from the continental margin, in the abyssal plains. The earthquakes of 25 Nov. 1941 and 26 May 1975, (Fig.1, Nos.2 and 14), both of magnitude 8.0, generated small tsunamis which were recorded by mareographic stations in mainland Portugal, Azores, Spain and Morocco. Both earthquakes show strike-slip focal mechanism solutions, whatever the nodal plane chosen as fault plane, and it is known that the earthquake of 25 Nov. 1941 generated a turbidity current which broke the submarine cables Brest-Casablanca and Brest-Dakar. These tsunamis were probably both generated by landslides on the sea bottom.

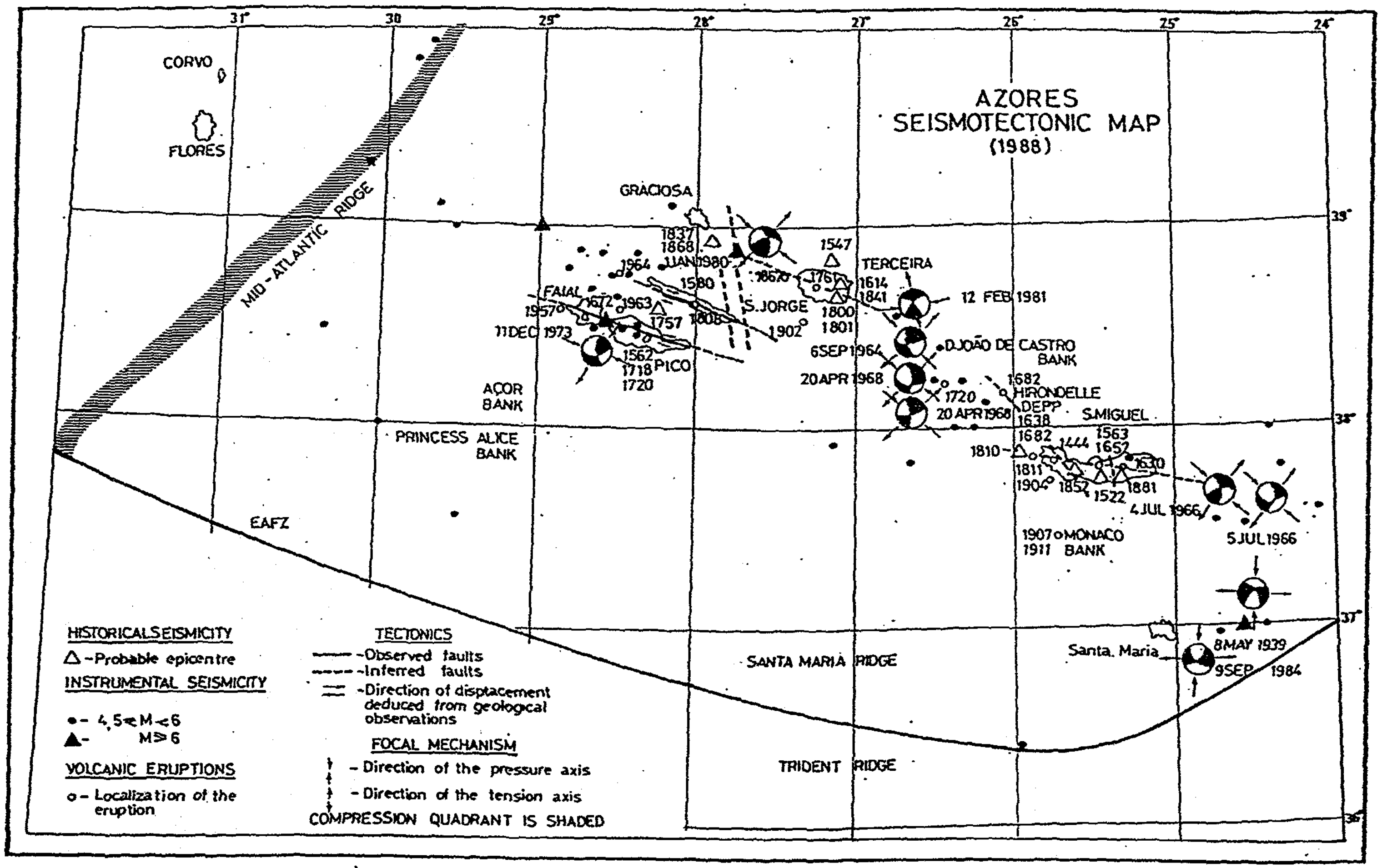

Figure 2. Seismotectonic map of the Azores region where are indicated the epicentres of the most important historical and recent earthquakes which occurred in that region.

In Fig. 2 is shown the seismotectonic map of the Azores in which are indicated the epicentres of the main historical and recent earthquakes situated in that region. All focal mechanism solutions are consistent with strike-slip, whatever the nodal plane chosen as fault plane, and with a tensional seismic stress field (Udias et al., 1976; Moreira, 1985). The most important earthquake that occurred in that region was certainly the earthquake of 1522 which caused a large number of deaths and heavy destructions in St. Miguel island. Although the epicentre of this earthquake appears to be situated in land, but not far from the Ocean, historical reports do not refer any tsunami or even agitation of the sea generated by the earthquake. The most important tsunami generated in the Azores region, although of small intensity, was set off by the earthquake of 9 July 1757, with epicentre situated in the channel that separates the islands of St. Jorge and Pico (St. Jorge channel). The earthquake was felt in ships that navigated in the Azores region and a low intensity tsunami struck Terceira, Graciosa, Pico and St. Jorge islands. The earthquake of 31 August 1926 with epicentre situated in the channel that separates the islands of Pico and Faial set off a small tsunami too. Waves about 50 cm high were observed but attenuated and disappeared rapidly. The earthquake of 8 May 1939, with epicentre situated eastwards of St. Maria island (Fig. 2) set off a small tsunami which was recorded only by mareographic stations in the Azores. The earthquake of 1 January 1980 set off also a small tsunami which was only recorded by mareographic stations in the Azores. Tsunamis generated in the Azores region are therefore very scarce and are originated by strike-slip earthquakes; they do not propagate much and the waves never reached considerable amplitudes. On the other hand it was not found any relationship between high waves in the Azores, referred by some authors, and earthquakes, even with epicentres out the Azores region. It is assumed that these waves are probably of meteorological origin.

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BASING IMPROVEMENT OF TSUNAMI WARNING SYSTEMS ON PRECIOUS EXPERIENCES

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Abstract

The evaluation of warning systems for rare events, such as tsunamis or meteorites, is difficult because of the rarity. For this reason, each experience is that much more valuable. Here we evaluate the Pacific Tsunami Warning Center and the Hawaii Tsunami Warning Center for the performances on the two large events of 1985 and 1986: the Mexican earthquake of 19 September 1985 and the Aleutian earthquake of 7 May 1986. Evaluation is guided by the operational research approach of Adams (1966); the final logs provided much of the information on times and sequencing. In hindsight, using a deadline of three hours before ETA, it now appears that a warning should have been issued for the Mexican earthquake and no warning should have been issued for the Aleutian earthquake: both are the opposite of what actually happened! These evaluations suggest that (1) the tsunami history was not adequately available in realtime; (2) the tsunami literature is not available to the warning system in a form useful in realtime; (3) instrumentation operating elsewhere in the Pacific was not being used; and (4) the decision-making process is distorted by the unrealistic demands of users. Some constructive suggestions for modifying the procedures of the systems are provided.

Quote: "Man who lives in a world of hazards is compelled to seek for security." (Dewey, 1929, page 3)

Background:

A recent vogue in government is to create a warning system for some natural hazard. This is considered by some to indicate that significant progress is being made in mitigating the losses due to that hazard. Unfortunately, there is either no objective evaluation of the performance of the system or the system is evaluated by members of the agency that is operating the warning system—so an objective evaluation can hardly be expected. See, for example, the statement by two employees of the U.S.G.S. concerning the performance of two volcano observatories operated, incidently, by the U.S.G.S., "The two observatories have excellent records of successful predictions of eruptions and well organized warning and response procedures with the officials in the area." (Gori and Shearer, 1987). While an internal "audit" might be of value internally, certainly, as in the financial world, an audit by a third party is to be preferred. In this present article, we attempt to serve as the third party in providing a preliminary audit of the Pacific Tsunami Warning Center (PTWC) and its ancillary activity as the Hawaii Tsunami Warning Center (HTWC). For accountability of public agencies, see Green et al. (1978)

Quote: "The more widespread become the habits of intelligent thought, the fewer enemies they meet from those vested interests and social institutions whose power depends upon immunity from inspection by intelligence, the more matter of course they become, the less need there seems to be for giving knowledge an exclusive and monopolistic position." (Dewey, 1929, page 297.)

Here we are concerned with a well-understood natural phenomenon, gravity waves, manifested as tsunamis. Unfortunately, some journalists are not aware of the fundamental simplicity of gravity waves (e.g. Taggert, 1987, who dramatically headlines his article, "Tsunami one of the sea's most baffling phenomena". Certainly misleading, considering that two bibliographies of the tsunami literature have been issued (Adams, 1967 and Pararas-Carayannis, Dong, and Farmer, 1982) and that there is a scientific journal solely devoted to the scientific aspects of the hazards of tsunamis.)

Early Efforts at Tsunami Warning

The history of attempts to operate a tsunami warning system in Hawaii has been well-documented since 1926, when Jagger, using seismic data on the Big Island, issued notices to the public. Here we focus on the yet more difficult problem of evaluating the performance of such a warning system. In 1966, Adams proposed using standard techniques of decision-making for such evaluation and showed how such an approach could improve the system by making the sensitivity of the performance to the system parameters more apparent to the operators of the system. This was carried further by Adams (1971, and 1972) when computer languages specially for modeling were used to model the interactions of the warning system with the public officials and the public. The feedback effect of a good performance–cooperation of the public, and the feedback effect of a false warning–reduced cooperation of the public, became demonstrable on the model.

Cox (1968) elaborated on the decision-making model for evaluating the system performance. Much difficulty was apparent for two reasons: first, many terms being bandied about were not quantitatively defined; and second, the Pacific Tsunami Warning Center (PTWC) lacked instructions that are cognizant of the feedback effect of poor decision-making. A glossary of terms for tsunami discussions has now been published (Adams and Nakashizuka, 1985), but no change has been made in the instructions to the PTWC.

These studies were supplemented by an outstanding field investigation on attitudes of Hawaii residents living near the shoreline and the relationships of these attitudes to the experiences and expectations of those residents (Havinghurst, 1967).

Recommendations by Tsunami Research Division (TRD)

Other ways of possibly improving the PTWC occurred to investigators in the Tsunami Research Division (TRD) located at the University of Hawaii in the Institute of Geophysics, and were written down, formalized, and provided to the PTWC Director (TRD recommendations 1 through 10, 1967).

Why PTWC is in the National Weather Service

The reader may wonder why the PTWC is in the National Weather Service instead of some other government agency, such as the U.S. Geological Survey which is responsible for related matters of earthquake prediction, strong-motion measurements, and seismic measurements. The motivation originally was to put all the warnings made to the public concerning natural phenomena in one agency thus concentrating the expertise of managing the information released to the public. (The warnings systems operated by the U.S.G.S. report to officials of the various states rather than directly to the public: a very significant difference.)

Surprisingly, although the National Weather Service has developed some excellent procedures for optimizing the weather forecasts disseminated to the public, none of these techniques have been applied to upgrading the PTWC. I attribute this to the anomaly of an oceanographic-seismic operation being within an organization knowledgeable about the atmosphere. Furthermore, National Oceanic and Atmospheric Administration NOAA seems to have selected the directors of the PTWC with great perspicacity. Not infrequently the selection, during the past twenty years, has been an officer of the NOAA Corps.

Within the past couple of decades, the training technique of "business-gaming" has become big business and is now ubiquitous in the world of computers. Despite its drawbacks, the benefits justify the resources required. For example, NASA's approach to simulating situations during training is well-known and requires extensive resources. No similar facilities exist in the United States for training personnel to operate a tsunami warning system.

Quote: "The common fact that we prize in proportion to rarity has a good deal to do with the exclusive esteem in which knowledge has been held." (Dewey, 1929, page 298).

PRECIOUS EXPERIENCES

Regrettably, knowledge must often be obtained the hard way, by experience.

Two Recent Experiences of PTWC

There have been two events during two recent calendar years that have prompted the public and scientists to focus attention on the PTWC: these are the Mexican earthquake of 19 September 1985 and the Aleutian earthquake of 7 May 1986. The logs of the activities at the PTWC for these two events have been made available to this author through the courtesy of the Director of the PTWC. These are public documents and are available upon reasonable demand to any taxpayer. See also Farreras and Sanchez (1987).

Selected activities from the log for the 7 May 1986 event have been used to suggest some improvements that may be possible for PTWC (Adams, 1987).

In 1980, I wrote to the Washington D.C. office of NOAA and made numerous recommendations for improving the performance of PTWC. Three of these were: double the staff; double the budget; allow some autonomy. The present staff is seven; the present budget is about $300,000 U.S. dollars per year; still there are very few discretionary funds.

HINDCASTING

PTWC Personnel and Funding

NOAA, like most government agencies, knows that legislators can see instruments and hence the agency buys many relatively inexpensive instruments. NOAA can scatter these about the Pacific Ocean rim, attempting to improve political relations. Often relatively large sums of money have been spent for many water-level gages, only to have them installed at sites so protected that there is little or no chance of useful information being obtained in real-time. Furthermore, these instruments usually have a very limited dynamic range, being designed for measuring the tide.

Here we are going to imagine that NOAA had actually followed my advice of 1980 and doubled the PTWC staff to fourteen and doubled its budget. Furthermore, we are going to consider two different deadlines for issuing a warning to the State of Hawaii: the two-hours before arrival, used prior to April 1978 and the three-hours before arrival, which is presently used.

I assume that of the added staff, one would have specialized in developing a computer data-base of the tsunami history and another would have been specializing in the analogous work of developing a computer data-base of the existing computer literature. Both of these are very feasible because the amount of information involved is small compared to the capabilities of the computers being marketed today. (The computing facilities that have existed for about the past five years at PTWC have two excellent features, viz. backup by full duplication and air-conditioned environment. Unfortunately, these aspects are vitiated by the fact that the software only allows for 64K words of RAM to be used on any particular problem. That is, the computer is obsolete.) Computers had been used previously in warning systems for epicenter determination (Adams, 1958) and proposed for Tsunami Research database (Walling, Freeman, and Adams, 1970).

The present staff is outstanding; they are to be commended for their dedication.

Assuming Doubled Staff and Doubled Funding Resources:

Handling of the 19 September 1985 Mexican Earthquake:

The specialized historian would have rapidly discovered that high waves due to a magnitude 8.1 earthquake occurred in 1932, sweeping away train tracks (Iida, Cox, and Pararas-Carayannis 1967).

Using leadtime of three hours prior to arrival in Hawaii: The PTWC should have issued a warning. That no warning was issued means that the State of Hawaii was subjected to the risk that a wave was generated.

Using the leadtime of two hours prior to arrival in Hawaii: The PTWC could have waited until additional information became available from other sources: seismic or water-level. No warning would have been issued.

Handling of the 7 May 1986 Aleutian earthquake:

The hypothetical new employee devoted to storing the tsunami literature in the computer database would be able to access predictions, such as Adams (1966), which show that the source area must have waves of several tens of feet before a damaging wave can occur in Hawaii–at least for a magnitude of 7.2, which implies a fault length on the order of the width of the strained zone. Comparison of this prediction with the information available to the PTWC from the Alaska Tsunami Warning Center (in Palmer, Alaska) would have indicated that NO warning should be issued. (See also Curtis and Måder, 1987.)

If, in addition, someone at PTWC has made contact with the Tsunami Warning Center in French Polynesia, he would have obtained a real-time estimate of the seismic moment. This was so low, about 10E28 dyne-cm, that no tsunami could have been generated that would be significant at a distance of thousands of nautical miles. (Talandier and Reymond, 1986).

The cost to the State of Hawaii for the false alarm of 7 May 1986 is estimated at about 30 million dollars. Ultimately, the public's loss of confidence in the warning system will prove even more expensive. This loss of confidence extended to satire: in Figure 1 is shown, by permission of JoAnn Ruppert–the artist, the design for a T-shirt that became an immediately sell-out.

Clearly, it would be to the advantage of the State of Hawaii to have its own tsunami warning center rather than depend upon the underfunded and understaffed federal center.

Figure 1: Silkscreen design by JoAnn Ruppert for T-shirt commemorating the tsunami alert of 7 May 1986: reproduced by permission.

THE HAWAII TSUNAMI WARNING CENTER

NOAA created the Alaska Tsunami Warning Center at Palmer, Alaska. This author informed the Office of Civil Defense in the State of Hawaii that such a system had been funded by NOAA and suggested that Hawaii should have such a system. Upon request, NOAA funded the installation in Hawaii by this writer and his colleagues of a telemetry system: two offshore bottom-mounted water-level gages were installed near Kailua, Kona and Honoapo on the Big Island and the signals transmitted, together with seismic signals from several locations, via relay, to Ewa Beach. Seismic signals from Mauna Kea and Haleakela were also transmitted, through the courtesy of the UH Institute of Astronomy, which provided housing in return for signal drops. This radio-based system was delivered to NOAA complete with manuals, calibration equipment, and permission authorizations. The data may have been useful because NOAA has continued much of this system ever since. The offshore water-level gages were not maintained by NOAA: however the AEC did maintain the Kona gage for two years.

The Hawaii Tsunami Warning Center is said to be at the Pacific Tsunami Warning Center. Unfortunately this arrangement has two bad effects. First, it places more work and responsibility on a staff already overloaded and second, the differences between the PTWC and the HTWC are not apparent. This is very serious, because the PTWC serves the Pacific Basin participants whereas the HTWC serves the State of Hawaii: the objectives are, therefore, very different. Overlay of the two systems may inadvertently lead to faulty performance.

POLICIES AND PRACTICES OF STATE OF HAWAII OFFICE OF CIVIL DEFENSE

The Office of Civil Defense for the State of Hawaii has the policy of keeping all procedures the same for all islands. Presumably this is to obtain maximum reliability via simplicity. However, in 1978, the Mayor for the Big Island insisted that the Big Island must have a warning at least three hours before the expected time of arrival ETA of the tsunami (Lam, 1987). The State of Hawaii Office of Civil Defense has, in keeping with its policy of standardization, required the Hawaii Tsunami Warning Center to provide warning at least three hours before the expected time of arrival ETA of the tsunami (Lam 1987).

Prior to the current procedures instituted from April 1978, two hours was used for the minimum time before expected time of arrival of the tsunami for which a warning was issued to the public. In practice, news services informed the public that a large earthquake had occurred and that the possibility of a tsunami was under evaluation. The significance of this is that those requiring long lead-time, such as owners of small boats, could take action deemed appropriate.

The shift from the two-hour deadline to the three-hour deadline has caused decisions to be made based on inadequate data, as in the case of the Mexican earthquake of 1985, or with inadequate time for interpreting or verifying given data, as in the 1986 Aleutian earthquake. Adams and Curtis (1986) have described the importance of obtaining verification of incoming data, as so much of it is erroneous and, of course, which data are erroneous is not apparent in real-time. See also recommendation No. 6 (Tsunami Research Group 1967).

Based on current technology and the budgets levels presently assigned to PTWC by NOAA, the three-hour deadline is not realistic.

COMPARISON OF PTWC WITH TSUNAMI WARNING SYSTEMS OF OTHER NATIONS

Japan: In January, 1987, this writer visited the Japan Meteorological Agency (JMA) in Tokyo, Japan. I was fully briefed, provided with descriptive material, and shown the new equipment being installed. The investment in an upgraded, highly automated computer-based system was prompted by the earthquake-tsunami event in the Sea of Japan 1983.

The objective of the new system, which became operational in March of 1987, is to reduce the time required in the sensing, processing, interpreting, and disseminating stages of the warning. The present system, which actually consists of several regional subsystems, is now able to operate within a time limit of under five minutes: this is a reduction of about fifty percent from the previous time attained of nine minutes. Indeed, now it is not apparent how the system can be improved, in an a posteriori sense: further research would be directed to prediction, so a priori.

Each of the regional centers are manned continuously with two or more persons. The system uses at least seven times more people than the PTWC.

Russia: The Tsunami Warning System of the USSR has used hardened water-level gages for more than ten years: these are sensors with remote readout that allow measurement of water amplitudes of tens of meters.

The number of persons involved in operating this system is not known.

French Polynesia: The Tsunami Warning System of French Polynesia uses more than fifteen people--even though the objective of the system is far more restricted than that of PTWC. This system has installed numerous long-period seismometers and developed a technique for rapidly estimating the seismic moment. (Talandier and Reymond, 1986).

As a result of these comparisons, we find that the PTWC of NOAA is characterized by: (1) least manpower; (2) lower quality instrumentation in the terminal areas than the USSR system; (3) lower quality seismic instrumentation than the French Polynesian system; (4) lower quality processing, interpreting, display and dissemination than the JMA system. In summary, we find that the tsunami warning-system of the USA has enormous opportunity for improvement.

ACKNOWLEDGMENTS

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TSUNAMI: INDUCED OSCILLATIONS IN CORINTHOS BAY MEASURMENTS AND 1-D vs 2-D MATHEMATICAL MODELS

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ABSTRACT

The recorded, on a tidal recorder in the bay of Corinthos, tsunami generated by a recent earthquake with epicenter in the area of Alkyonides islands is analysed and compared to the results given by 4 mathematical models in 1 and 2 dimensional space,for the generation of tsunamis and for the estimation of eigenperiodes and eigenmodes in geophysical coastal basins.

INTRODUCTION. SCOPE OF THE STUDY.

The present study was triggered from a recent earthquake event strongly felt in Athens. It had a submarine epicenter and induced a minor tsunami in the bay of Corinthos. The sunami induced oscillations were recorded by the tidal recorder of Isthmia.

PHYSICAL CONDITIONS

The bay of Corinthos extends between Peloponesse and the mainland of Greece. It has an elongated form and direction E-W. From N, S and E is bounded by coastal boundaries and on the W has a narrow opening (straits of Rio-Antirio). It has characteristic length 120 km, a maximum width 25 km and depth reaching 900 m. Close observation reveals a peculiar plan form and bathymetry. An eastern main basin of mean depth 500 m, length 75 km mean width 20 m can be distinguished. It is connected to a smaller western basin of length 45 km, width 10 km and mean depth slightly less than 300 m.

On the northeastern edge of the bay lies the group of Alkyonides islands. There layed the epicenter of the strong earthquake (6.9 Ricter scale) of February 24, 1981 which was strongly felt in Athens and generated a minor tsunami in Corinthos gulf.

The bay topography, the earthquake epicenter, deduced from the earthquake elements (aftershock tsunamogenic area) and the recorded tsunami at Isthmia, are described in Fig.1 and Fig.2.

The seismic origin water wave superimposed on the normal semidiurnal tide. The max amplitude (assuming that the tidal recorder fully responds to that frequency) was of the order of 30 cm, the induced seiching lasted, before full decay, almost 4 days and from the simple counting of wave peaks it is deduced that the seiching period was approximately 37 min.

a. Models of long waves propagation.

2-D case:

The linearisation (dropping the nonlinear convective acceleration terms and linearising bed friction) and the elimination of the horizontal velocity components u, v in the 2-D long waves model, leads to useful and computational feasible equation in $\zeta(x,y,t)$: (see notations of fig. 3).

$$\frac{\partial^2\zeta}{\partial t^2} = \frac{\partial}{\partial x}\left(gh\frac{\partial\zeta}{\partial x}\right) + \frac{\partial}{\partial y}\left(gh\frac{\partial\zeta}{\partial y}\right) - \frac{k}{h}\frac{\partial\zeta}{\partial t} + \frac{\partial^2\zeta_b}{\partial t^2} \qquad (1)$$

The field equation is coupled with the following boundary conditions:
Coastal boundaries $\partial\zeta/\partial\eta=0$.
Open sea boundaries $\partial\zeta/\partial t+(gh)\,\partial\zeta/\partial\eta=0$
Tsunamogenic region $\zeta_b(x,y,t)$= a given variation from a zero initial value a final bed deformation.

1-D case:

In the case that the one horizontal dimension (say ox) of the geophysical basin investigated is of higher order of magnitude that the second one, the domain can be concidered as a 1-D enlongated channel of variable cross sections (variable depth and width) and the continuity and equilibriun equations can be formulated in the x, t space according to the notations of fig. 4

$$\frac{\partial\zeta}{\partial t} + \frac{1}{B}\frac{\partial Q}{\partial x} = \frac{\partial\zeta_b}{\partial t} \qquad (2)$$

$$\frac{\partial Q}{\partial t} + \frac{\partial Q^2/A}{\partial x} = -\,gA\frac{\partial\zeta}{\partial y} - gAsf \qquad (3)$$

Where Q (x,t) the discharge (Q=AV) A the cross section area (which for the orthogonal parallelepipedic approximation is A=Bh) and sf the frictional loss of energy sf $=v^2/c^2R$ (R=A/P, P= wet perimiter and C the Chezy bed friction coefficient.

The field equations are completed by boundary conditions:

On coastal boundaries V=Q=0

On open sea boundaries Q/B=V $h=\zeta\sqrt{(gh)}$

b. Models for the investigation of natural periods and modes of basin oscillation.

1. One dimentional basins.

For elongated quasi 1-D basins Defant's method is used. This try and error method is started with the use of an arbitrary period (usually based on Merian's Formula). The equations are resolved numerically fron one end of the basin to the other. The realisation of the known boundary condition at the end, is a control for the success of the selected period. The attempt is repeated with modified T value up to the realisation of the expected standing wave. (for more details see Defant, 1961, p.165)

2. Two dimensional case.

A more realistic and accurate method is to consider the two dimensionality of the basin and apply the long waves 2-D model assuming that the free surface temporal variation is a simple trigonometric function : $\zeta=\zeta_0(x,y)\sin(\omega t)$, $\omega=2\pi/T$. Substitution of that form for ζ in (1) and simple mathematic manipulations lead to Helmholtz equation.

$$\frac{\partial}{\partial x}\left(gh\frac{\partial\zeta_o}{\partial x}\right)+\frac{\partial}{\partial y}\left(gh\frac{\partial\zeta_o}{\partial y}\right)+\zeta_o\omega^2=0 \qquad (4)$$

On the coastal boundaries the full reflaction condition is $\partial\zeta_0/\partial\eta=0$ and on the open sea boundaries the formation of nodes permits the approximation $\zeta_0=0$. From mathematic point of view it is a boundary value (eigenvalue) problem. The eigenvalue of the system are the eigenfrequencies and the corresponding eigenvectors are the amplitude values on the specified nodes on the 2-D grid.

MODELS' APPLICATION IN CORINTHOS BAY

The difference in the characteristic horizontal dimensions of the bay of Corinthos suggest that 1-D models may be adequate. In order to verify the observed period of tsunami induced free oscillation of the bay, Defant method was applied. The flow domain was discretised in 120 sections of length 1000 m. The results are shown in table 1.

The physically paradoxic indication is that the bay although under free oscillations, incited from one end, has a preference to the second mode of oscillation instead of the first one.

To investigate further that seemingly paradox the 2-D Helmholtz equation was applied for the reestimation of the natural periods and modes of oscillations of the bay. The flow domain was discretised by a square grid with dx=dy=7500 m with 39 interior nodes. The eigenperiods are shown in table 1 in descending order.

It is deduced that the longer eigenperiod given by the 2-D model is close to the observed one. This can be atributed to the peculiar form and the bathymetry of the bay. An eastern main basin of mean depth 600 m of length 75000 m and width 20000 m can be distinguished from a narrower western extern extension of length 45000 m width 10000 m and depth < 300 m. The eastern basin establishes the basic period of 37 min, a fact that can not be detected by the 1-D Defant procedure.

For the final modelization of the tsunami generation and propagation in Corinthos bay both 1-D and 2-D long waves models were applied.

For the application of the 1-D model the space discretisation step was dx=5000 m. A uniform unit upthrust evolving sinusoidaly over 60 sec from zero to the unit final value generates a long unidimensional wave that propagates between the two lateral reflective boundaries. The mareograms corresponding to locations corresponding to the coastal towns of Corinthos, Galaxidi and Nafpactes (Lepante) and the corresponding periodograms resulting from the Fourier analysis of the mareograms are depicted in fig. 5a and fig. 5b. From the peaks of the periodograms the deduced eigenperiods are shown in table 1.

The weaknes of the 1-D model to describe both the natural periods and the real time evolution of the seismic wave is reconfirmed.

The linear long waves model in two dimensional space was finally applied on the entire complex of Corinthos and Patra bays up to the Ionian open sea boundary.

The flow domain discretised by 5000x5000 m square grid. The time step used was dt=30 sec. Along the aftershock tsunamogenic elliptic area a uniform unit upthrust evolving sinusoidally from zero to the final value was assumed.

The program ran for 270 min real time and the computed mareograms at 6 coastal locations are depicted in fig.6. The Fourier analysis gave periodograms at the coastal towns of Corinthos, Xilocastro, Egio, Patra, Nafpactos, Galaxidi and Itea depicted in fig.7. The main eigenperiods revealed from the periodograms are shown in table 1.

CONCLUSIONS

For the specific case of Corinthos bay, the tsunami record revealed the difference in the descriptive ability of the 1-D and 2-D long waves models and the superiority of the 2-D models in describing the hydrodynamic idiomorphy of the basin.

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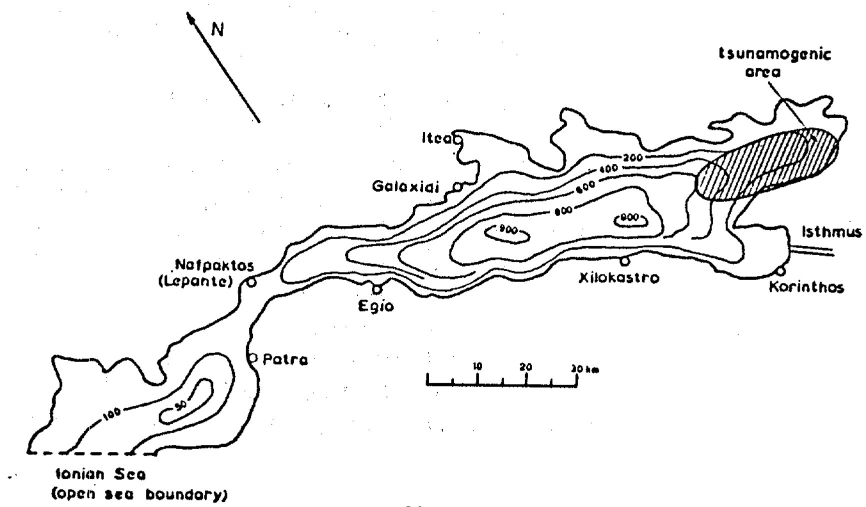

figure 1

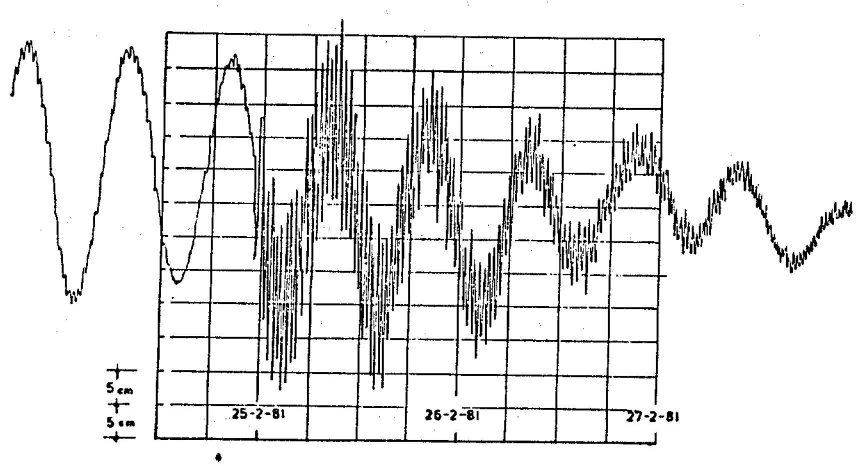

figure 2

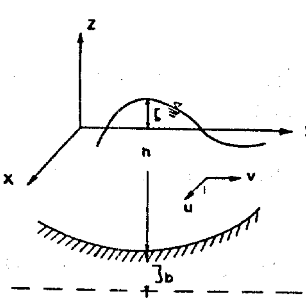

figure 3

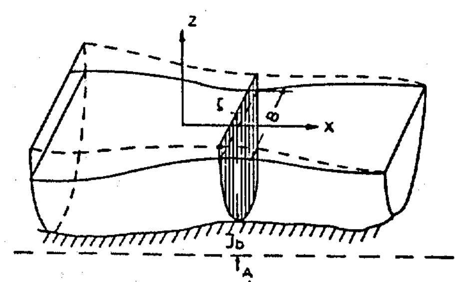

figure 4

MODEL	1-MODE	2-MODE	3-MODE	4-MODE
DEFANT	57	37	27	21
HELMHOLTZ	37.3	24.5	17.4	13.2
1-D	50	29	22	17
2-D	38	24	18	13.5

Table 1

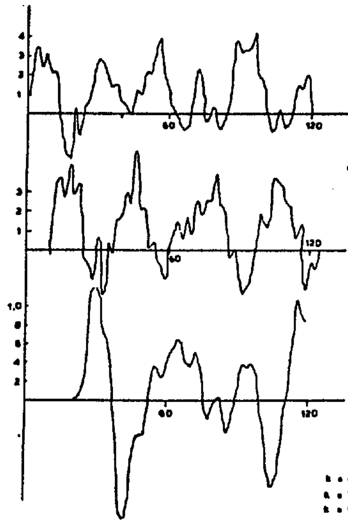

figure 5a

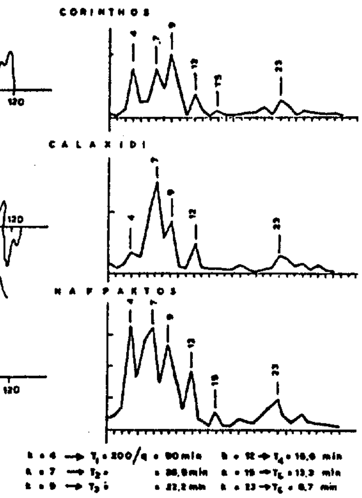

figure 5b

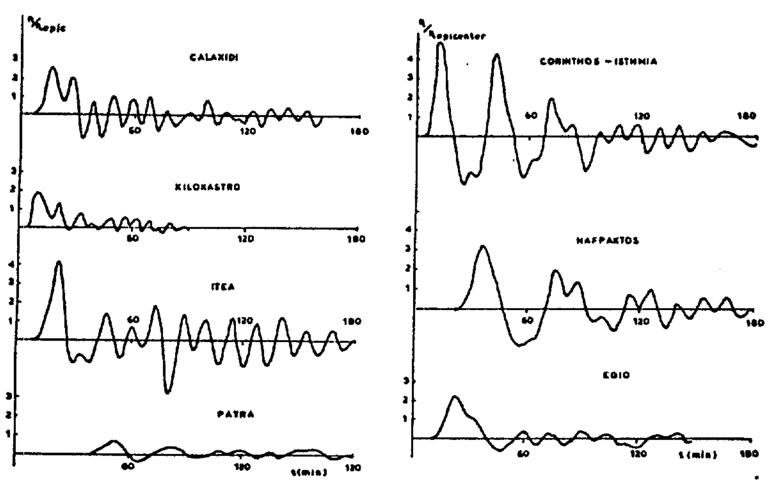

figure 6

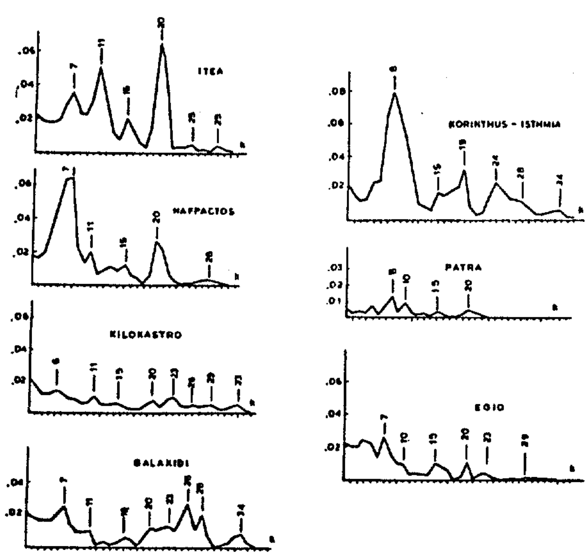

figure 7

SOME TSUNAMI MEMORIES

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ABSTRACT

The author has been involved in developing and administering tsunami warnings for about a half-century. Although much of these activities has been included in published papers by many authors, there are included here various background material, anecdotes, etc. that may be both interesting and important for historical purposes.

There is a growing interest among geophysicists in documenting for historical purposes interesting events and decisions, particularly personal happenings that ordinarily are not described in published papers. It is particularly appropriate for this to be done by senior scientists. At the half-century mark in my oceanographic career, I have already done this for tides (Zetler, 1987) and what follows is a parallel attempt to do something similar for tsunamis.

When I entered the Tides and Currents Division of the US Coast and Geodetic Survey in 1939, I found that there was an ongoing program of noting tsunamis on tide guage records and preserving copies of these in a historical file. There was no indication of who had started this program but it was expected that the file with these data would be continued and, I suppose, there was the thought that hopefully it would have some future use (Heck, 1947). In those days, the waves were called seismic sea waves; that name was discarded later when it was learned that submarine earthquakes generate shock waves in the water column capable of significantly shaking ships at the surface and causing a ship's crew to believe they have grounded on a shoal. I suspect there may still be some "reported shoals" on some nautical charts because of this phenomenon.

On April 1, 1946, an earthquake in the Aleutian Islands generated a tsunami (Bodle, 1946). The large waves, traveling in deep water at the speed of a jet plane, ravaged the Hawaii coasts four and a half hours later. Like all tsunamis in Hawaii's history, these also arrived without any warning; once again it was an act of God!

There were many seismograms and marigrams depicting this event available to scientists in the Coast and Geodetic Survey, fueling a research effort in the agency to learn more about the phenomenon. Elliott Roberts, chief of the Seismology Division, and Charles Green, chief of the Tides and Currents Division, gave their support to these efforts and, indeed, joined in enthusiastically in the effort (Green, 1946). It was not long, less than a year after the earthquake, that discussions began leading to a conclusion that maybe there was a possibility that Hawaii could be warned of an expected arrival of a tsunami in the future. It was an exciting thought!

As in any bureaucratic government agency, the first thought is to prepare a budget proposal requesting Congressional funding of research for the program. As a scientific agency, the C&GS was somewhat of an orphan in the Department of Commerce where the major interest was in business. Making the funding possibility even more dismal was an ongoing economy effort in post-war Congress with research as a primary target for funding cuts. As a matter of fact, the small three man research section I was in, under Walter Zerbe, was hiding behind a strange name, Section of Oceanography in the Tides and Currents Division. Because the new funding possibility was so remote and because Roberts and Green were unwilling to wait a year or two for the normal budgetary procedures, a decision was made to go ahead using our very sparse appropriation as effectively as possible. The effort became a US government program without Congressional authorization or funding. It was a brave, and perhaps foolhardy, decision on the part of these men because they knew that inadequate funding was bound to cause weak links in the system. They knew only too well that if the system failed after Hawaii had been promised a warning system, Congressional inquiries were inevitable and their careers could be significantly affected.

The plan for a warning system involved a network of seismological stations connected by rapid communication facilities to a center in Hawaii. Each station would report arrival times of the "p wave", the primary wave going directly through the earth from the epicenter to the seismograph, and the "s wave", the wave going along the surface of the earth. The time difference between these two arrivals is a measure of the distance of the epicenter from the seismograph station. At the warning center, a seismologist would make appropriate arcs on a globe, each centered on the appropriate seismological station; the intersection of all the arcs would locate the epicenter. The entire process could be completed in an hour or less.

The seismologist at the center would have an approximation of the magnitude of the earthquake from the reports but he would have no way of determining whether the quake had moved the sea floor vertically causing the movement of a large column of water either up or down and therefore likely to generate a tsunami, or if the earthquake consisted of vertical shearing, in which case no wave is apt to be generated. A network of tide stations, each with good communications to the warning center, would be needed. The center would, upon locating the epicenter, request tide gauge reports beginning at the anticipated arrival time at each nearby station. If the earthquake event was near Hawaii, the system could not function in time. However, it could be effective for destructive waves originating in the Aleutians, Kamchatka, Japan, Chile and other remote areas that had been the sources of destructive waves in the past.

There were four immediate requirements that would be necessary to implement a warning system:

1) Seismographs at that time were recording photographically with the film developed every 24 hours. A change to an inked (visible) record was needed and, in addition, an alarm system had to be added to sound an alarm whenever large oscillations exceeded pre-set limits, indicating a large earthquake.

2) Extremely rapid communication arrangements were needed between the warning center and each seismograph and tide station. Arrangements were negotiated with existing military and civilian communication systems with extremely high priorities during an alert and lower priorities for monthly dummy tests in both directions.

3) It was necessary to develop a quick and accurate method for estimating tsunami travel times from an epicenter to

Hawaii. The best available source for information on tsunami speeds was a recent text, "The Oceans" by Sverdrup et al (1942); it gave the formula for the speed of a shallow-water wave (wave length much greater than the depth) as $\sqrt{gd}$ where g is gravity and d the depth. However, Sverdrup went on to say that there was found to be a biased error when calculations based on mean depth along a path were compared with observations. We experimented with this formula and found that the problem was in taking a mean depth for the path; when the formula was used for small sections along the path (120 nautical miles or less) in each of which the depth was reasonably constant and the travel times then added, the total matched historic observations quite well. Our file of earlier tsunami records was very useful here. With this method now available and trusted, we set about constructing a travel time chart for tsunamis from any epicenter in the Pacific to Honolulu. The result was a chart of the Pacific centered in Hawaii with somewhat circular curves for each half-hour from Honolulu to the various Pacific continents. An article (Zetler, 1947) was published including a copy of the chart, a discussion of the method of construction and a comparison of observed and computed travel times to Honolulu and Hilo for all tsunamis recorded at these locations in the past.

4) Finally, we had a problem in that radio communications to some island tide stations were maintained on a limited time basis during the day whereas we needed 24-hour service. It was necessary to invent a "seismic sea wave detector" that would alert the tide observer that a tsunami had arrived. Hopefully, he would then get the radio operator to send an appropriate report on this to the warning center. We built a mechanical band-pass filter that would not respond to tidal rise or fall at the low frequency end of the wave spectrum, nor would it respond to wind waves at the high frequency end of the spectrum. In between, it would respond to waves with periods of roughly one minute to an hour and was designed to sound an alarm if the amplitude of a wave in this frequency band exceeded a preset limit (Zerbe, 1948). Some tide records show seiche in this frequency band and marigrams for such stations had to be scanned carefully to establish an alarm limit sufficiently high to eliminate false alarms due to seiche. Furthermore, a tide observer had to be instructed on how to differentiate on a marigram between seiche (ordinary continuous oscillations) and tsunamis (sudden beginnings of oscillations).

In 1948 the US Coast and Geodetic Survey informed Hawaii that a seismic sea wave warning system for Hawaii was operational. The Civil Defense authorities were given responsibility for warning the people of an alarm, and educational programs were initiated for informing people as to the nature of tsunamis and what they should do when warned (Roberts, 1950 and Zerbe, 1953). People had to be taught that the first wave was usually not the largest and therefore they were not to run towards the sea when the first wave receded but instead to move quickly to higher ground; ships were to head out to sea where tsunamis represented no hazard and, indeed, frequently were not even noticed.

Funding for the warning system remained non-existant for a few years despite annual budget proposals. After the system worked well for the tsunami from Kamchatka in November 1952 (C&GS, 1953), small but inadequate funding was initiated in Congressional appropriations. A colleague, Harris B. Stewart, met a senator from Hawaii at a social function and explained the importance of the warning system to him. The senator offered to write to the Secretary of Commerce suggesting increased budgetary support and asked Stewart to draft an appropriate letter for that purpose. When the letter arrived, it was routed to Stewart and me to prepare the Secretary's reply. There were several such exchanges with Stewart and me writing both ends of the correspondence. Eventually it did help our funding a bit. I think I worried more than Stewart did about the possible consequences if the Secretary ever found out our dual role in the exchange.

The system worked well for a tsunami from the Aleutians in March 1957 (Salsman, 1959) and again for a tsunami from Chile in May 1960 (Symons and Zetler, 1960). However, when the warning authorities in Japan were informed that the tsunami was only moderate in size at Hawaii, they assumed it would be even smaller by the time that it reached Japan. Accordingly, they cancelled the warning. The wave reached Japan almost 24 hours after its genesis near Chile with destructive force, causing loss of life and property. Apparently, they were unaware of a study by Van Dorn, originally in a classified unpublished report, showing that after a wave has traveled one fourth the circumference of the earth, it converges and becomes larger (Van Dorn, 1974). If you think of a tsunami originating at the North Pole of an earth completely covered by a deep layer of ocean, it will diverge (and get smaller) until it reaches the equator; thereafter it will converge and become larger. Japan is about three eighths of the circumference of the earth from Chile; the tsunami was also large and destructive on the Kamchatka coast.

At the 1959 meeting of the International Union of Geodesy and Geophysics (IUGG) in Helsinki, an IUGG Tsunami Committee was organized with R. Takahashi as chairman. Soon thereafter he invited me to join the committee and to attend the first meeting of the committee in Honolulu in 1961. I went there angry, with a chip on my shoulder. The agenda for the meeting included a discussion of tsunami travel time charts and I had read a paper on this subject recently in a Japanese journal in which the author started out by dismissing "Zetler's method" because it took advantage of the relatively smooth distribution of depths in the Pacific and therefore it was unacceptable as a general method for preparing tsunami travel time charts. I was concerned only with what worked for the Pacific Ocean. In my paper on creating this first chart (to Honolulu), I had carefully compared all observed times with computed times and found a good match. By 1961, I had either made or supervised the making of about twenty additional charts for various tide stations in the warning ystem and, for each, a comparable comparison of observed and computed travel times had been satisfactory. My general feeling was that we had a need for charts in the Pacific and here the method worked. Basically, "if it ain't broke, don't fix it"!

On the second (and last) day of the meeting, the subject had not yet been discussed. I raised the point and spoke rather strongly in defense of the method we were using. At the end, Takahashi smiled and bowed as did Dr. Wadati, head of the Japanese Meteorological Agency, an influential member of the committee. Then, to my great surprise, we went on to another topic without any discussion of my presentation. I left for home after the meeting rather smug as to how well I must have presented my case. That feeling dissipated quickly a few months later when I happened to read the novel, "The Ugly American". In the book I found out that when a Japanese gentleman disagrees with a colleague, he will smile, bow, and change the subject. There was no doubt in my mind that Takahashi and Wadati were gentlemen!

Early in this paper I noted an ongoing C&GS program of filing copies of marigrams showing tsunamis. When underwater H-bomb tests started in the 1950's, we were informed that marigrams showing tsunamis from the explosions would be classified. This would be disastrous to our goal of compiling long tide records for analysis and historical purposes. Accordingly, an unofficial policy was adopted not to notice these occurrences on the marigrams. At the same time, Japanese agencies published comparable records freely. I remember a foreign scientist asking me about such records at an international meeting. I appreciated greatly his prefacing his question with the warning, "Before you answer this question, remember I am from East Germany". He, too, was a gentlemen!

Hawaii's great concern with tsunami and the warning system brought me into close contact with Doak Cox of the University of Hawaii. He was truly "Mr. Tsunami" in Hawaii and was active in all aspects of tsunami research and in civil defense aspects of warning facilities and arrangments. He was a key person in the IUGG Tsunami Committee with the chairman depending greatly on his advice. From time to time, he and I discussed our mutual interest and the possibility of merging some of the tsunami research activities of the Coast and Geodetic Survey and the University of Hawaii. Finally, we agreed to try to organize a Joint Tsunami Research Effort (JTRE) at the Institute for Geophysics at the University. The Coast Survey was expanding its research efforts at the time and I had little difficulty in gaining support. However, for a while, Admiral Karo insisted on JTRE being located at the warning center in Ewa Beach while I was insisting that an appropriate academic atmosphere was possible only on the campus of the University. By enlisting the aid of key research people in the Coast Survey, I finally prevailed and a contractual agreement along these lines was signed. I then recruited Gaylord Miller when he completed his Ph.D. at Scripps Institution of Oceanography to head JTRE. When the Coast Survey and the Weather Bureau merged into ESSA (Environmental Science Services Administration), I was appointed Director, Physical Oceanography Lab, Atlantic Oceanographic and Atmospheric Laboratories (AOML) and JTRE came under my direct supervision. In order to provide Miller with a reasonably viable team, I transferred several vacancies in my lab to JTRE; Miller used these very effectively in recruiting new personnel. Miller and I had a very close rapport (we had worked closely together during my sabbatical at SIO in 1962-63) and he very quickly established JTRE as a major international organization in the field with scientists from many countries visiting JTRE on sabbaticals. Later in the decade, ESSA established a research laboratory comparable to AOML on the west coast, Pacific Marine Environmental Laboratory (PMEL). An administrative decision then moved JTRE into PMEL on the basis that a lab in Hawaii should be assigned to a facility located on the Pacific rather than to my location in Miami. It was upsetting to both Miller and me, but nevertheless we maintained a close informal relationship until his untimely death in 1976. Gaylord Miller was one of the brightest and most likable people I have ever known and I still mourn for him.

In 1964 a representative of the Matson Line (passenger ships then sailing between Hawaii and California) requested that I prepare a paper for the company magazine on the warning system as related to ships. I wrote a comprehensive paper on the subject with the title, "The Seismic Sea Wave Warning System, An Aid to Mariners". Subsequently I received an apologetic note from the Matson Lines saying that the paper was fine but it would not be used since company policy ruled out using anything that might alarm the passengers. I filed the paper and it remained there for about a year until I received a call from the editor of the Mariners Weather Log requesting essentially the same paper. He was quite surprised when I told him I needed one day to comply with his request, to change the final paragraph to include the recent establishment of ESSA. The paper (Zetler, 1965) was later included in a book collecting papers on marine science written in non-technical language (Gordon, 1970). I was pleased to have my name linked as an author with Genesis, Plato, Pliny, Leonardo Da Vinci and others.

In the early 1960's, Leonard Murphy and I shared managerial responsibility for the warning system, he in seismology and I in oceanography. I was always looking for appropriate locations for additional tide stations and the Gulf of Alaska had a high priority in this search. I found one island in a very desirable location but the Coast Pilot for the area said little about it. I requested a reconnaisance of the island by a Coast Survey ship scheduled to do hydrography in the general area. When the ship approached the island, a shot was fired over its bow and a booming voice announced that it was a classified area of the US Navy and the ship was to turn around. Needless to say, we did not get that location.

In the very late 1960's, Robert White, Administrator of NOAA (National Oceanographic and Meteorological

Administration), issued an order placing all NOAA real-time warning systems under the Weather Service. As I recall the order, the Tsunami Warning System was not specifically named but, nevertheless, here was a system transferred to an agency which had no seismologists and no oceanographers trained in tsunamis. Obviously, a few people would be transferred, but I was both appalled and disheartened at the change. I knew a personal appeal was pointless as I had recently cancelled my lab's participation in BOMEX (Barbados Oceanographic and Meteorological Experiment), a huge NOAA air-sea interaction program. I had done this after being told that ship priorities would preclude any ship from taking time to salvage a buoyed line of our sensors if it should break loose from its bottom moorings. Because my lab's annual appropriation for equipment was $6,000, I had not changed my decision despite strong pressure from NOAA including a personal call from Richard Hallgren, White's top assistant. Incidentally, my estimate of oceanography's low priority in NOAA was seconded by a lecturer on BOMEX at NOAA's Fisheries Lab in La Jolla a few years later. He started his talk by writing "BoMEX" on the board and saying "with a lower case o, as in NoAA".

I decided my best choice of reversing the order was by enlisting the aid of William Hess, Director of NOAA's Environmental Research Labs. I was getting ready at this time to deploy two free-fall pelagic tide gauges in the center of the Caribbean as part of the Unesco CICAR (Cooperative Investigations of the Caribbean and Adjacent Regions) program in which I was international subject leader for tides. I knew that Bob Munson, commanding officer of the ship from which the gauges would be launched, had been the Honolulu District Officer for several years and therefore he was very familiar with the tsunami warning system. I guessed that he would be equally upset by the order placing the system in the Weather Service and therefore, if I could persuade Bill Hess to join us on the Caribbean cruise, Munson would join me in requesting Hess's aid in appealing the change. I was gratified when Hess accepted my invitation and Munson and I did besiege him with our protests. Finally, he cut us short by telling us he agreed with us but he had a limited number of items he could take to White and he did not give this a sufficiently high priority to include it in his agenda. There was nothing more I could do!

Even though at that time I had no formal role in the warning system, I had kept in touch with the people involved. At that time I was chairman of the IUGG Tsunami Committee and was a member of the US delegation to IOC (International Oceanographic Commission) meetings on tsunami warning systems. At the 1971 IUGG meeting in Moscow, I resigned from my chairmanship of the Tsunami Committee. A year later when I became eligible to retire from NOAA, I accepted an invitation by Walter Munk to join him at SIO.

I've had a few tsunami experiences since, several very pleasant and one very painful. In 1977, after Gaylord Miller's death, I was invited to a NOAA meeting in Honolulu at which it was decided to terminate JTRE and move some of the personnel to PMEL in Seattle. In 1981 I had the honor of being the only foreign speaker at the opening ceremonies for an IUGG symposium in Sendai, Japan. Again, in 1985, I was invited to give the principal address at an IUGG tsunami meeting in Victoria, Canada. In the same year I visited the Tsunami Warning System Center in Hawaii and was delighted to learn that our long-time dream of real-time transmission of tide heights by satellites to Hawaii had finally been achieved.

As my professional career reaches the half-century mark in 1988, I look back on my participation in tides research and in tsunami warnings with a good deal of satisfaction. In a small way, I, too, have left behind me some "foot-prints on the sands of time".

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PLANNING FOR RISK: COMPREHENSIVE PLANNING FOR TSUNAMI HAZARD AREAS

SUMMARY

The project was conducted by Urban Regional Research (Jane Preuss, Principal Investigator), INA Civil Engineers of Tokyo, Japan with Professors Shuto and Horikawa as advisors. In addition, Salvador Farreras of CICESE and Antonio Sánchez of Baja California, Mexico were responsible for analysis of the Mexican earthquake.

The Japanese team first applied a hydrodynamical model to the case study area of Kodiak, Alaska to refine capabilities to project exposure to the hazard. Subsequently, prototype vulnerability assessments are conducted for a community susceptible to a high velocity event (Kodiak, Alaska) and a low velocity event (Lázaro Cárdenas, Mexico).

Based on an improved understanding of vulnerability to both the direct and indirect forces, an urban planning methodology emphasizing risk reduction through mitigation and preparedness is developed. Assumed vulnerability is from direct water forces (hydrodynamic and buoyant forces) and from secondary forces such as loss of ground support (erosion and ground failure), debris impact (boats, vehicles, dislodged structures), and fire. In addition, indirect effects such as economic/revenue loss are profound. Planning methodologies are use oriented (commercial, industrial, general marine, and resort). In addition, life safety preparedness programs are suggested.

The implementation program is incremental in orientation. It is not intended that each community will adapt the "whole package" rather that they will select individual recommendations responding to the highest vulnerability to secondary impacts, e.g. erosion and fire. It is also intended that these recommendations be equally applicable to reduction of hazards from coastal winter storms, floods, and port fires.

Copies of the report are available for $25.00 U.S. from:
Urban Regional Research
1809 7th Avenue; Suite 1000
Seattle, Washington, 98101.

Suite 1000
Tower Building
1809 Seventh Avenue
Seattle, WA 98101
(206) 624-1669

ITS:WS:ITSU XII

FIRST ANNOUNCEMENT

INTERNATIONAL TSUNAMI MEETINGS

Three International Tsunami Meetings, ITS'89, WORKSHOP and ITSU-XII will be held consecutively in Novosibirsk Science Center of the Siberian Division of the USSR Academy of Sciences in I989.

ITS'89 — The International Tsunami Symposium of the Tsunami Comission of the International Union of Geodesy and Geophysics, July 3I - August 3, I989.

WORKSHOP — The International Workshop on the Technical Aspects of Tsunami Warning System, Tsunami Analysis, Preparedness, Communication and Instrumentation, August 4-5, I989.

ITSU-XII — The Twelfth Session of the International Coordinating Group for the Tsunami Warning System in the Pacific, August 7-IO, I989.

The second announcement and call for papers to be issued in October I988 will contain full details about the submission of papers, registration fee, social events, etc. It will be distributed among those who will complete and return the form below by October I, I988. All correspondence should be kindly directed to:

TSUNAMI 89 ORGANIZING COMMITTEE

Computing Center, Novosibirsk, 630090, USSR.

FULL NAME ______________________________

POSITION ______________________________

INSTITUTION ______________________________

MAILING ADRESS ______________________________

______________________________ PHONE ______________

I plan to participate in [ITS'89 | WS | ITSU-XII]

At ITS'89 I plan to present [I | 2 | 3] papers

Accompaning persons [No | I | 2 | 3]

I would like to participate in postevent tour (Baikal Lake, 5-6 days about 300$ per person) [Yes | No]

ВЦ Зак.375-100-88г.

APPLICATION FOR MEMBERSHIP

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

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

☐ Student ☐ Member ☐ Institutional Member

Name ______________________ Signature ______________________
Address ______________________ Phone No. ______________________
____
Zip Code ______________________ Country ______________________
Employed by ____
Address ____
Title of your position ____

FEE: Student $5.00 Member $25.00 Institution $100.00
Fee includes a subscription to the society journal: SCIENCE OF TSUNAMI HAZARDS.

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