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SOME OPPORTUNITITES OF THE LANDSLIDE TSUNAMI HYPOTHESIS

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ABSTRACT

Tsunami sources are intimately linked to geological events. Earthquakes and landslides are shown to be part of a continuum of complicated geological phenomena. Advances in landslide tsunami research will remain coupled with marine geology research. The landslide tsunami hypothesis is shown to have originated in the scientific literature in the early 1900s. Tsunami science has been slow to embrace the hypothesis in part because of the tremendous uncertainity that it introduces into tsunami gneration. The 1998 Papua New Guyinea event sparked much controbersy regarding the landslide tsunami hypothesis despite a preponderance of the evidence in favor of one simple and consistent explanation of the tsunami source. Part of the difficulty was the unanticipated distinction between slide and slump tsunami sources. Significant controversies still exist over other aspects of the Papua New Guinea event. The landslide hypothesis will become widely acceepted once direct measurements of underwater landslide events are made. These measurements will likely be integrated into a local tsunami warning system.

INTRODUCTION

For the sake of argument, let us define a tsunami as the water waves resulting from an identifiable geological event, which may involve an earthquake, a landslide, volcanic activity, a gas diapirism, etc. Further, let the geological event occur in any body of water, whether ocean, river, or lake. A geological event can be a tsunami source by its ability to generate water waves through a particular and distinct mechanical action. As we shall see, these definitions accept the possibility that a single nearshore earthquake can unleash numerous tsunami sources. A single earthquake can generate more than one tsunami if rupture involves localized fault mechanics. For example, the strike-slip 1999 Izmit earthquake in Turkey (Yalçiner *et al.*, 1999) generated tsunamis at several submerged fault step-overs, where the subsidence within each step-over is the salient tsunami source mechanism. The remainder of the numerous potential tsunami sources are associated with landslides triggered by the earthquake. On land, a magnitude 7 earthquake can trigger thousands of landslides, with smaller events typically being more frequent (Wilson and Keefer, 1985; Kramer, 1996). Visual observation of the sea floor near the 1998 Papua New Guinea earthquake epicenter (Tappin *et al.*, 2001) revealed many slides and rock falls less than 1 m thick and 100 m long, supporting the idea that there could have been thousands of submarine mass failures. Given the 1-2 km depth of these landslides, most of them would not be tsunamigenic, although they could have been if they had occurred in a shallow nearshore environment. Some tsunami researchers refer to certain atmosphere induced water waves as tsunamis. Atmosphere induced long waves are perhaps better referred to as seiches to distinguish these phenomena from geological events.

The landslide tsunami hypothesis can often be proven whenever the landslide source is subaerial (Miller, 1960) or partially exposed along a harbor front (Plafker *et al.*, 1969; Bjerrum, 1971; Seed *et al.*, 1988; Synolakis *et al.*, 2000). Therefore, it is not surprising to find early speculation regarding the potential tsunamigenicity of underwater landslides (Milne, 1898; Montessus de Ballore, 1907; Gutenberg, 1939). The first experimental work on landslide tsunamis was apparently performed by Wiegel (1955), although the impetus for the research was nuclear bomb tests in the South Pacific as opposed to geological events (Raichlen, pers. comm.). To put the case for the landslide tsunami hypothesis succinctly, landslide tsunamis have amplitudes proportional to their vertical center of mass displacement (Murty, 1979; Watts, 1998, 2000). Underwater landslides can have vertical displacements of up to several kilometers, contrary to coseismic displacement during earthquakes which rarely surpasses 5 m (Geist, 1998). Without addressing the frequency of occurrence of such catastrophic events (see Watts and Borrero, 2001), the maximum tsunami amplitude from the largest possible landslide on earth is dictated solely by the depth of the oceans. Landslide tsunamis therefore pose one of the greatest tsunami hazards.

These basic considerations are often obscured by confusion regarding the mechanics of tsunami generation. Such confusion, when it manifests itself, may be derived in part from the inherent complications of geological phenomena: there is some potential common ground whereby earthquakes and landslides resemble each other. Earthquakes have slip surfaces and landslides have failure planes that are both crustal dislocations. A subsiding block delimited by two coeval normal faults (Yalçiner *et al*., 1999) is not unlike a landslide in that the block may be completely detached from the crust and moving coherently down due to gravity. A giant slump extending through lithified sediment (i.e., soft rock) to the subduction zone (von Huene *et al*., 2001) is not unlike an earthquake in that limited horizontal motion occurs along a slip surface. This qualitative comparison may be extended to certain quantitative measures that show additional similarities between earthquakes and landslides. For example, landslides routinely release as much potential energy as earthquakes release in elastic energy (Tappin *et al*., 2001; Ruff, 2001). In addition, both earthquakes and landslides experience many more small events than large events. There is sufficient geological commonalty between earthquakes and landslides to potentially generate confusion.

On the other hand, there are many events that can be squarely considered as either earthquakes or landslides. The delineations between earthquakes and landslides, for those events where delineations make sense, may be the cause, extent, and duration of rupture (or failure). To be concrete, a typical earthquake with elastic energy release lasting 10 s is manifestly different from a typical landslide with potential energy release lasting 600 s. When the delineation between earthquakes and landslides is clear, landslides can be much more efficient tsunami generators than earthquakes (Wiegel, 1955; Watts, 2000; Ruff, 2001). This suggests that the mechanics of wave generation can remain two distinct asymptotic limits (Hammack, 1973; Watts, 1998): for example, there may not be an earthquake tsunami analogy to the 1994 Skagway, Alaska landslide tsunami. This proposition should not be surprising because the term underwater landslide encompasses all submerged rock slides, reef failures, and myriad forms of sediment failure (Hampton *et al*., 1996; Turner and Schuster, 1996). Underwater slides are identified by thin, translational failures that travel long distances, while underwater slumps are defined to undergo thick, rotational failures with minimal displacement (Prior and Coleman, 1979; Edgers and

Karlsrud, 1982; Schwab *et al*., 1993). Broadly defined, approximately half of all underwater landslides appear to be slides, whereas the other half of all underwater landslides appear to be slumps (Schwab *et al*., 1993). For example, most of the local tsunamis within Prince William Sound, Alaska in 1964 were generated by underwater slides (Plafker *et al*., 1969), whereas the 1998 Papua New Guinea tsunami was generated by an underwater slump (Tappin *et al*., 2001). Because center of mass motion governs landslide tsunami generation (Watts, 1998; Watts *et al*., 2000, 2001a, 2001b), the tsunamigenicity of underwater landslides is intimately related to the dynamics of mass failure, which is controlled in turn by the local geology, including classifications such as a slide or a slump (Tappin *et al*., 1999). Therefore, geology, or earth sciences, contributes to tsunami research as a scientific discipline that describes, organizes, interprets, and explains tsunami generation.

The use of depth-averaged landslide tsunami generation models may also be propagating errors and misconceptions regarding the landslide tsunami hypothesis (Watts *et al*., 2000). Despite the groundbreaking numerical work on landslide tsunamis by Mader (1984) and Iwasaki (1987), the most influential work remains the sequence of papers by Jiang and LeBlond (1992, 1993, 1994). The latter work considers translational slides that produce small waves propagating in the opposite direction from landslide motion. This observation is most likely a by-product of depth averaging. The 1964 Alaskan events (Plafker *et al*., 1969), the 1994 Skagway event (Synolakis *et al*., 2000), and the 1998 Papua New Guinea event (Tappin *et al*., 2001) all prove that significant wave energy propagates back towards shore, as observed by Heinrich (1992) and Watts (1997) during laboratory experiments. The models of Jiang and LeBlond (1992, 1993, 1994) also concentrate on reproducing landslide deformation instead of landslide center of mass motion. This ability renders their models ideal for studies of landslide deposition, but raises serious doubts about their applicability to tsunami generation. Once again, Watts *et al*. (2000, 2001a, 2001b) show that center of mass motion is a much more important consideration for tsunami generation than landslide deformation. Last of all, the models of Jiang and LeBlond (1992, 1993, 1994) may not accurately reproduce tsunami amplitude (Murty, 2001). The tsunami amplitude can instead be estimated from available analytical approximations (Striem and Miloh, 1976; Murty, 1979; Pelinovsky and Poplavsky, 1996; Grilli and Watts, 1999; Goldfinger *et al*., 2000; Bohannon and Gardner, 2001; McAdoo and Watts, 2001; Murty, 2001; Watts *et al*., 2001a). Landslide tsunami generation should not be depth averaged whenever possible.

COMPLICATIONS OF LANDSLIDE TSUNAMIS

This section examines in a heuristic manner the fascinating complications introduced by the landslide tsunami hypothesis. The tsunami generation scenarios considered here are rough and ready interpretations of the possible consequences following a nearshore earthquake. Before the threat of landslide tsunamis became widely studied, the number of tsunami generation scenarios to consider consisted of something like:

Mild EQ	No transoceanic tsunami
Strong EQ	No transoceanic tsunami
Strong EQ	Transoceanic tsunami

These three scenarios form the basis of our valuable tsunami warning centers. Because of the variety in earthquake focal mechanisms and magnitudes and hypocenter depths, a modest fraction (perhaps 20% according to the Russian online tsunami catalogue) of large earthquakes produce significant transoceanic tsunamis. Seismographs provide preliminary earthquake magnitudes and hypocenters with which an assessment regarding transoceanic tsunamis needs to be made. Telemetered (or online) tide gauge records now assist the warning centers in judging tsunami amplitude in real time, along with historical records of events from the same region, and possibly some deep ocean pressure sensors. Local tsunamis, as opposed to the transoceanic events, were assumed to follow an earthquake that would serve as a warning to seek safe elevations or inland shelter.

The situation gets quite complicated when landslide tsunami scenarios are added into the mix. For example, the 1999 Fatu Hiva tsunami was generated by a spontaneous subaerial landslide and therefore there was no warning of tsunami arrival at a seaside village 5 km away (USC tsunami web site). Also, when there is an earthquake or some sediment loading, sediment slopes can fail due to pore water migration up to several years after the perturbation (Biscontin *et al.*, 2001). Last of all, a magnitude 7 earthquake can be presumed to generate thousands of underwater landslides, the large majority of which will not be tsunamigenic, but a few of which may generate catastrophic tsunamis (Wilson and Keefer, 1985; Imamura *et al.*, 1995; Imamura and Gica, 1996; Kramer, 1996; Fryer *et al.*, 2001; Tappin *et al.*, 2001). Consequently, we face a situation roughly described by the following eleven tsunami generation scenarios:

No EQ	Landsliding without tsunami
No EQ	Landsliding with tsunami
Mild EQ, no tsunami	Landsliding without tsunami
Mild EQ, no tsunami	Immediate landslide tsunami
Mild EQ, no tsunami	Delayed landslide tsunami
Strong EQ, no tsunami	Massive landsliding without tsunami
Strong EQ, no tsunami	Landsliding with immediate tsunami
Strong EQ, no tsunami	Landsliding with delayed tsunami
Strong EQ, tsunami	Massive landsliding without tsunami
Strong EQ, tsunami	Landsliding with immediate tsunami
Strong EQ, tsunami	Landsliding with delayed tsunami

The scenarios are constructed to make several points and as such they are not meant to be taken literally. First of all, landsliding is recognized as being ubiquitous in these scenarios. Landsliding is commonly triggered by any earthquake with magnitude greater than 5 based on terrestrial observations (Wilson and Keefer, 1985; Kramer, 1996). Second, earthquakes and landslides are dealt with as essentially independent tsunami sources in a combinatorial manner. This is the prime reason for the increase in scenarios. Last of all, the distinction between local and transoceanic tsunamis is blurred by these scenarios. On the one hand, a massive local tsunami is not necessarily indicative of a transoceanic tsunami (Tanioka, 1999). On the other hand, a magnitude 7.1 earthquake can still result in a massive transoceanic tsunami, as in the 1946 Unimak, Alaska event (Fryer *et al.*, 2001).

The reader will notice that volcanic events (collapses, lahars, pyroclastic flows, etc.) and subaerial landslides have been left out of the previous discussion. The full range of tsunami generation mechanisms is quite broad, and the potential combinations of tsunami sources is therefore extremely complicated, even if one accepts the succinct classification of scenarios described above. Identifying distinct tsunami sources associated with a single geological event has become a priority in tsunami science. Tsunami catalogues will no longer be able to attribute water inundation heights to a single tsunami source mechanism. Inversion of wave data from sparsely distributed ocean bottom pressure sensors or tide gauge stations will need to consider multiple tsunami sources based in part on geological evidence (Fryer *et al.*, 2001). Geologists have a crucial role to play in multidisciplinary tsunami research by distinguishing and interpreting potential tsunami sources (Tappin *et al.*, 1999; Zitellini *et al.*, 2001). It is this intimate connection between tsunami sources and geology that currently suggests defining tsunamis as originating from geological events.

Not all landslides are created equal in terms of tsunamigenicity. The range in possible landslide tsunami amplitudes goes from zero up to the vertical landslide displacement, which may be a significant fraction of the maximum local water depth (Murty, 1979; Watts, 1998, 2000). Tsunami amplitude is therefore an important quantity to resolve for underwater landslides. In general, tsunami amplitude will depend most on the landslide volume and its mean water depth (Murty, 2001; Watts *et al*., 2001a), both essentially geological quantities. Many landslide tsunamis produce highly localized waves, and the common perception is that landslide tsunamis must remain local events. However, there is no fundamental reason why large landslides cannot also produce transoceanic tsunamis along rays emanating from the axis of failure (Ben-Menahem and Rosenman, 1972; Iwasaki, 1997; Fryer and Watts, 2000; Fryer *et al*., 2001). Landslide sediment can control both the size of failure and the landslide motion as demonstrated by the 1998 Papua New Guinea event (Tappin *et al*., 1999, 2001; Watts *et al*., 2001b) as well as Monte-Carlo predictions (Watts and Borrero, 2001) and outbuilding delta simulations (Syvitski and Hutton, 2001). To this day, there is a multiplicity of landslide classifications based on failure morphology, sediment type, landslide dynamics, etc. where the reader will appreciate that these classifications are not unique and can be interrelated (Hampton *et al*., 1996; Turner and Schuster, 1996; Keating and McGuire, 2000; McAdoo *et al*., 2000; Watts *et al*., 2001a). Underwater slides and underwater slumps, regardless of how they are defined, serve as useful end members that bound the range of possible tsunami features (basically amplitude and wavelength). However, the differences between tsunami features generated by otherwise identical slides and slumps can reach up to a factor of five (Watts *et al*., 2001a, 2001b).

The wavelengths of landslide tsunamis also vary significantly and have an important bearing on the recognition of such events. Because of the relatively short duration of most earthquakes, the wavelength of an earthquake tsunami is governed by the length of rupture (Hammack, 1973) which is often greater than 40 km for tsunamigenic events (Geist, 1998). The tsunami period follows from the wavelength and the typical water depth of coseismic displacement. On the other hand, because of the relatively long duration of most landslides, the tsunami period follows from the duration of landslide acceleration (Watts, 1998). The tsunami wavelength then follows from the period and the mean water depth near failure. Note that the order in which wavelength and period are calculated is reversed because of the differing wavemaker regimes. The fundamental period is nearly preserved during tsunami propagation, while the wavelength varies with bathymetry. Tsunamigenic landslides off Southern California can be expected to have periods of 3-20 minutes, whereas transoceanic tsunamis often have periods greater than one hour (Watts *et al.*, 2001c). Therefore, landslide tsunamis can often be recognized by their relatively short period, even if they are transoceanic events (Fryer and Watts, 2000; Fryer *et al.*, 2001). The typically shorter period of landslide tsunamis leads to an observational complication hitherto overlooked for transoceanic events of longer period: the leading elevation wave in the far-field behaves like an Airy wave of infinite wavelength (Mei, 1983; Watts, 2000). Consequently, this Airy wave can come and go without any significant interaction with the local bathymetry, somewhat like a very small amplitude tidal fluctuation. Eyewitness reports of leading depression N-waves prior to tsunami attack may have more to do with **not observing** the rise and fall of the leading elevation Airy wave than any other fundamental wave mechanics.

One of several obstacles to widespread acceptance of the landslide tsunami hypothesis may be precisely the (scientific and tsunami warning and psychological) uncertainty that it introduces. This uncertainty includes the unfamiliar terminology and ideas borrowed from geology or soil mechanics and incorporated into tsunami science. This uncertainty includes the occurrence of tsunamigenic landslides, the size and shape of tsunamigenic landslides, the initial acceleration of the landslides, and the tsunami amplitude generated by the landslides. This uncertainty includes the new models, new institutions and new financial structures required for the research. Some people may inherently wish to fight the landslide tsunami hypothesis because they would prefer to live without such uncertainty. Scientists who subscribe to this view are ignoring the many fascinating (if sometimes foreign) research issues available to tackle. Landslide hazards motivate the prediction of landslide tsunamis. And, the desire to predict hazards opens up a multitude of research and collaboration opportunities (for example, see http://*rccg03.usc.edu/la2000/*). Given the new opportunities made available, acceptance of the landslide tsunami hypothesis may represent a coming of age for tsunami science. Some researchers even see the emergence of a new scientific discipline, landslide hazards, out of the current multidisciplinary efforts. Regardless, change is afoot.

UNCERTAINTY AND THE PAPUA NEW GUINEA EVENT

The seminal event for landslide tsunami research to date is the 1998 Papua New Guinea tsunami. First of all, the staggering and violent loss of life stimulated international attention and concern (Kawata *et al.*, 1999). Second, the mismatch between earthquake magnitude and tsunami amplitude is surpassed in recent history only by the 1896 Sanriku, Japan and 1946 Unimak, Alaska events (Russian web site). Third, the eyewitness observations could not be reproduced by any earthquake tsunami source based on reasonable parameters for rupture along the subduction zone (Titov and Gonzalez, 1998). Fourth, the landslide tsunami hypothesis is bolstered by interpretation of marine surveys (Tappin *et al*., 1999, 2001). Fifth, the eyewitness observations appear to be reproduced by numerical simulations using a single slump tsunami source (Heinrich *et al*., 2000; Tappin *et al*., 2001; Watts *et al*., 2001b). The specific definition of a slump tsunami source was achieved with scaling analyses and numerical models (Watts, 1998, 2000; Grilli and Watts, 1999, 2001) that were concurrently available to mesh with the latest in sea floor mapping technologies. The merger between modern engineering models and modern marine surveys may be the single most important scientific outcome of the Papua New Guinea research, and the process has only just begun (Day *et al*., 2000; Synolakis *et al*., 2000; McAdoo and Watts, 2001; von Huene *et al*., 2001; Fryer *et al*., 2001; Locat *et al*., 2001; Watts *et al*., 2001c; Zitellini *et al*., 2001).

Some experience acquired during the early phases of the 1998 Papua New Guinea research and debate may serve the larger tsunami community. This experience is foremost an affirmation of the multidisciplinary nature of modern tsunami research in general, and of tsunami sources in particular. However, the experience has also involved significant uncertainty. One can in general collect evidence for a single tsunami event, including

An apparently normal subduction zone earthquake
Arrival foremost of a leading depression N-wave
Exceedingly large tsunami amplitudes above sea level
Limited and focused longshore runup distribution
Eyewitness accounts of a peculiar time of arrival
Multibeam bathymetry of the regional sea bed
Identification of a recent and large underwater slump
Sea floor photos of fissures and deformation on the slump
Seismic survey records clearly showing a large slump
Acoustic records of failure at tsunami generation time
Seismic records of failure at tsunami generation time
Slump simulations that satisfy most available evidence

and yet arguments regarding the tsunami source continue unabated (Geist, 2000, 2001; Okal and Synolakis, 2001). The most common reasoning among skeptics trying to refute the landslide tsunami hypothesis is to discredit each line of evidence one at a time. For example, the acoustic records may indicate a time of mass failure (Okal, 2000) and the marine survey records may prove the existence of a slump (Tappin *et al*., 2001), but one can only prove that the two are related with a sophisticated landslide detection system. Such a landslide detection system may be implemented as part of a local tsunami warning system, but no such systems have apparently been deployed aside from civilian uses of the military SOSUS arrays (Walker and Bernard, 1993; Caplan-Auerbach *et al*., 2001).

The nature of opposition to the landslide tsunami hypothesis is revealed by examining the Papua New Guinea tsunami literature in a historical context. The work presented by Tappin *et al*. (1999) was an interdisciplinary collaboration of all tsunami scientists onboard the Kairei cruise KR98-13. Despite this apparent consensus, there was considerable opposition to the landslide tsunami hypothesis at the nightly scientific meetings on the Kairei. The opinions at these scientific meetings are recorded by Satake and Tanioka (1999) and Matsuyama *et al*. (1999) in favor of the earthquake hypothesis, and by Tappin *et al*. (2001) in favor of the slump hypothesis. The opposition can be related in part to unfamiliarity: there was no landslide tsunami generation model on board the Kairei, and landslide tsunami generation estimates were computed by hand from newly available literature (Watts, 1998; Grilli and Watts, 1999; Watts, 2000). Given the absence of a well accepted landslide tsunami model and the novelty of such interdisciplinary tsunami research, the poor reception accorded the landslide tsunami hypothesis almost seems inevitable in retrospect. In the far-field, there is no debate that the main shock generated an earthquake tsunami measured in Japan (Tanioka, 1999; Tappin *et al*., 2001).

The landslide tsunami literature deals almost exclusively with tsunami generation by thin, translational slides as opposed to thick, rotational slumps. The Papua New Guinea mass failure is a slump (Tappin *et al*., 1999). Slumps are often comprised of cohesive sediments that travel a small fraction of their initial lengths. In 1998, there was little appreciation of the stark impact on tsunami features between the center of mass motion of a slide *versus* that of a slump: for the same size and density landslide, tsunami amplitudes and wavelengths can differ by up to a factor of five depending on the center of mass motion (Watts *et al*., 2001a, 2001b). There are two slump generated tsunami studies of the 1992 Flores Island tsunami prior to the 1998 Papua New Guinea event (Imamura *et al*., 1995; Imamura and Gica, 1996). Once the Papua New Guinea slump was identified, analytical approximations were adopted on the Kairei to finesse this difficulty, with only moderate success (Tappin *et al*., 1999). The simulation work described in Tappin *et al*. (2001) and Watts *et al*. (2001b) was first attempted after the Kairei cruise in response to the need for numerical simulations with accurate slump motion. Consequently, given the difficulty and the originality of the work, it was impossible for anyone on the Kairei (including the author) to have appreciated the scientific complications awaiting simulations of the 1998 Papua New Guinea tsunami source. Because the Kairei data could explain some of these complications and dispel some of the uncertainty, tsunami research was fundamentally guided by geological interpretation of sea floor data (Tappin *et al.*, 2001).

CONTROVERSIES REMAINING FROM PAPUA NEW GUINEA

Recent geology and simulation work published on the 1998 Papua New Guinea tsunami (Tappin *et al.*, 2001; Watts *et al.*, 2001b) has not resolved a number of finer issues regarding that event, even if one believes wholeheartedly and unwaveringly in the slump source. A subset of these potential controversies are reviewed here, with some arguments revealed despite the absence of data and analyses. The point in reviewing these potential controversies is to demonstrate by way of several examples that geology is complicated. And, as water waves generated by geological events, tsunami sources inherit a large fraction of the complications. This section therefore supports the claim that marine geology is in fact a necessary science for tsunami research, in the sense of inclusion among current tsunami disciplines. A history of tsunami research might argue in favor of seismology as the fundamental science of tsunami generation, but seismology may be in the midst of being overtaken by its rightful heir, geology. This is not to say that seismology will disappear from tsunami science, only that there are needs in tsunami science that seismology cannot address.

Shallow Water Bathymetry

The focusing of waves onto Sissano Lagoon has been attributed to a hemispherical shallow shelf leading from the sand spit to the location of slumping (Tappin *et al.*, 1999; Matsuyama *et al.*, 1999; Heinrich *et al.*, 2000). Despite this shelf, the directivity of landslide tsunami energy should have produced maximum runup around Malol instead of Sissano Lagoon (Iwasaki, 1997), but wave energy was refracted away from Malol and onto the shelf by the deep water of Yalingi canyon (Tappin *et al.*, 2001). Two-dimensional analytical work suggests that the shallowest bathymetry appears to have the most impact on local tsunami runup, especially the final beach angle (Kanoglu and Synolakis, 1998). Because the Kairei bathymetry stops at 200 m depth and the nautical chart has sparse soundings, can any of the Papua New Guinea numerical simulations claim to be accurate?

Existing numerical simulations (e.g., Tappin *et al*., 2001) manage to reproduce the observed tsunami focusing without any specific consideration of beach angles. One explanation of this apparent contradiction is that three-dimensional wave focusing may depend most on a particular range of water depths and distances (say from 100 to 1500 m and >5 km from shore), and thereafter refraction can no longer impact wave height considerably. Another explanation is that the sensitivity implied by analytical runup work may be in error because it assumes that depth-averaged equations are valid throughout runup -- a broken wave that approaches the shoreline as a bore may be insensitive to beach angle. Yet another explanation, although much less likely, is that the bathymetry files may have captured the correct beach angles more or less by accident. The shallow water bathymetry is needed in order to test these very different explanations. A shallow water survey would be interpreted by marine geologists.

Spray and Booms

There was a loud boom heard at Sissano Lagoon and what could be spray observed on the horizon a short time before the 1998 Papua New Guinea tsunami struck (Davies, 1998). It seems fair to speculate that the boom was related to the presumed spray because the observations were nearly simultaneous. There are two competing explanations of these curious observations. The first explanation is that a buried gas pocket exploded through the sediment as the leading depression wave, which corresponds to about a one atmosphere drop in overburden in this case, propagated overhead 15-18 minutes after the main shock, depending on the unknown location of the gas pocket. The location of the diapirism may be where bubbles were seen rising to the surface 5-10 km from shore the day before the earthquake and tsunami (Davies, 1998), providing clear evidence that buried gas pockets exist. The second explanation is that the highly nonlinear wave approaching on a shallow shelf formed a plunging breaker, known to produce a loud bang and vertical spray. Both explanations are possible, with no published data or evidence favoring either the geological explanation or the wave mechanics explanation. A shallow water survey could provide evidence of diapirism. More accurate numerical simulations are needed.

Missing Seismicity

Villagers at Malol witnessed the tsunami arriving immediately after a strong pair of aftershocks that occurred approximately 20 minutes after the main shock (Davies, 1998). Witnesses at the nearby village of Arop did not feel the strong aftershocks prior to tsunami arrival. It seems logical to conclude that the tsunami arrived at Arop before the aftershocks. However, numerical simulations uniformly predict that the tsunami arrived at Arop several minutes after Malol and therefore following the aftershocks, assuming the tsunami source is on the continental slope in front of Sissano Lagoon. This result is not surprising: the deep water of Yalingi canyon allows the wave to arrive earlier at Malol than at Arop, which is fronted by the shallow shelf. The horseshoe configuration of waves observed converging on Arop also support the tsunami arriving there last among adjacent villages (Davies, 1998). This last observation is also consistent with numerical simulations. If the tsunami arrived at Arop later than at other nearby villages, why did residents of Arop not report feeling the aftershocks? One possible answer is that Arop is situated on the sand spit in front of Sissano Lagoon (Tappin *et al*., 2001; Watts *et al*., 2001b), which experienced significant evidence of liquefaction (Davies, 1998; McSaveny *et al*., 2000). The village of Malol apparently had more solid foundations. Regardless of the veracity of this explanation, it is clear that eyewitness accounts need to be interpreted in the context of the local geology (McSaveny *et al*., 2000).

Steep Slopes

Off northern Papua New Guinea, the largest slope that failed in 1998 was initially around 15 degrees, which is reasonably steep for a submarine slope. Nevertheless, the amphitheater itself was surrounded by vertical rock cliffs and intact sediment masses that were observed at up to 40-50 degrees. In general, marine geologists observe that shallow slopes are just as likely to fail as steep slopes (McAdoo *et al*., 2000; O'Grady *et al*., 2001). One of the least inclined slopes produced the largest failure off Papua New Guinea in part because the accumulated sediment mass existed. While gravity drove the slump following failure, one can readily show that gravity could not have initiated failure on its own (Bardet, 1997). Ground motion attributed to the main shock can be ruled out because of the close proximity of the intact steeper slopes and because the slump failed about 11 minutes after the main shock. Pore water migration can probably be ruled out because most sediment masses appeared (visually and from coring) to consist of biogenic mud with low porosity (Tappin *et al*., 2001). A sediment comprised of stiff clay with relatively strong shear strength tends to produce larger slumps and tsunamis when failure does occur (Syvitski and Hutton, 2001; Watts and Borrero, 2001). According to the equations of slope stability (Turner and Schuster, 1996), water pressure is therefore necessary to initiate failure, but which water and from where? Marine geology investigations reveal significant water flows through faults beneath and within marine sediments (Sibson, 1981a, 1981b; Moore *et al*., 1986; Orange *et* *al*., 1999; von Huene *et al*., 2001; Tappin *et al*., 2001). Tsunamigenic underwater landslides almost certainly result from a confluence of tectonic, geological, sedimentary, and oceanic processes. The continental slope is but one relatively minor quantity that enters into a slope stability calculation (Watts and Borrero, 2001). Certain key indicators of failure such as stiff clays, mid-slope faulting, and chemosynthetic organisms have already been identified thanks to extensive marine surveys off Papua New Guinea, Costa Rica, Oregon, California, etc. (Moore *et al*., 1986; Orange *et al*., 1999; McAdoo *et al*., 2000; von Huene *et al*., 2001; Tappin *et al*., 2001). It remains to be seen how often and where these key indicators overlap to produce a tangible threat of tsunamigenic landslides.

PHILOSOPHY OF THE LANDSLIDE TSUNAMI HYPOTHESIS

The Papua New Guinea experience of complications, uncertainty, and controversy is by no means an isolated event. For example, those who study the landslide tsunami hypothesis are revisiting historical records with the intent of explaining candidate landslide tsunami events (Day *et al*., 2000; Fryer and Watts, 2000; Fryer *et al*., 2001). These researchers must regularly invoke the philosophy of science and paraphrase Occam's razor: which explanation fits all of the available evidence with the simplest available theory? In the course of discussing one such historical analysis, a respected seismologist dismissed

> Large wave amplitudes relative to moment magnitude
> Very localized region of tsunami runup
> Extraordinarily long duration earthquake
> 1:1 aspect ratio of the aftershock region

as not being evidence of a landslide earthquake (i.e., mass failure induced seismic radiation) and a landslide tsunami. These facts can indeed be explained by tectonic processes, although somewhat improbable or abnormal. The facts can also be explained by a normal underwater landslide. The most reasonable answer according to Occam's razor, which is only one philosophical measure, appears to favor the landslide source over the earthquake source. Scientists are supposed to recognize the viability of a theory directly from the preponderance of misfitting evidence that it can explain. Why is the landslide tsunami hypothesis then so often rejected in spite of the available evidence?

Most often, after reviewing and acknowledging the misfitting evidence, the landslide tsunami hypothesis is soundly rejected on the grounds of not being an "available theory", which then yields the seismic explanation by default. This is an obvious Catch 22. By denying that the landslide tsunami hypothesis is an "available theory", tsunami scientists are then able to conclude that the misfitting evidence never points to a landslide tsunami, thereby avoiding the complications, uncertainty, and controversy as well as the opportunities of the hypothesis. The argument is futile, just as it is false. The 1955 Lituya Bay and 1964 Alaskan events prove that not only is the landslide tsunami hypothesis an "available theory", but that it is a necessary and sufficient condition to explain the observed phenomena. The 1998 Papua New Guinea event represents perhaps the best studied of a reasonably long list of mutually supporting events with common features: leading depression wave, exceptional wave amplitude, short period, localized runup peak, strong far-field directionality, etc. (Imamura and Gica, 1996; Iwasaki, 1997; Kawata *et al*., 1999; Tappin *et al*., 2001; Watts *et al*., 2001a; Fryer *et al*., 2001). These explanatory abilities and common features constitute important foundations of the science of landslide tsunamis. The logical conclusion is that opposition to the landslide tsunami hypothesis is illogical. While this may be true, the situation is actually much more complicated, as pointed out in previous sections above.

Seismologists probably retain their hegemony over tsunami generation by virtue of the tools of the trade: seismographs are deterministic and often remote instruments. Any reasonable seismic reconstruction of the measured records must perforce be a plausible and viable seismic scenario. When expressed as distant inverse problems with sparse measurements, seismic scenarios can finesse the geological question: what is really going on in the earth? Therefore, any scientist can choose to play devil's advocate for the older or more isolated or more distant tsunami events (e.g., Tanioka, 1999 or Kikuchi *et al*., 1999). But to what end? As of now, earthquake tsunami sources enjoy a familiarity, immediacy, and positivism by virtue of the available measurement tools. Landslide tsunami sources do not yet enjoy widespread use of equally powerful tools (viz., Walker and Bernard, 1993). This situation makes defending the landslide tsunami hypothesis difficult as of now. For example, a marine geologist has the ability to interpret and to evaluate tsunami scenarios based on marine survey evidence, often in conjunction with modeling efforts (Goldfinger *et al*., 2000; Tappin *et al*., 1999, 2001; Watts *et al*., 2001c; Zitellini *et al*., 2001). However, even direct marine geology evidence of tsunami generation remains somewhat interpretive, primarily with regard to the time and rate of faulting or landsliding (Tappin *et al*., 2001). Cable breaks sometimes provide localized spatial or temporal information on landsliding, but not necessarily on tsunami generation (Heezen and Ewing, 1952; Kuenen, 1952; Houtz, 1962; Bjerrum, 1971, Yehle and Lemke, 1972; Hasegawa and Kanamori, 1987). The current situation favoring seismologists is bound to change in the face of recent research interests and ongoing technological innovation. The landslide tsunami hypothesis will be proven time and again when scientists deliberately measure for such events with a specialized tool, such as a hydrophone array or a single force seismic inversion (Eisler and Kanamori, 1987; Hasegawa and Kanamori, 1987; Kawakatsu, 1989; Okal, 2000; Caplan-Auerbach, 2001). The level of proof available is inherent to the tools of the various trades. The lack of well developed and reliable underwater landslide measurement tools means that the landslide tsunami hypothesis will continue to be challenged. How long can this situation last?

In the absence of conclusive measurement tools, those who study landslide tsunamis need to be cautious when advancing an event as having a landslide source. Otherwise, some scientists will have sufficient ammunition to say that proponents see nothing but landslide tsunamis. However, skeptical scientists also need to be cautious in their criticism of landslide tsunami sources. There are indeed landslides almost everywhere, whether on land or underwater, whether large or small, whether tsunamigenic or not. And enormous events such as catastrophic volcano collapses must do something big to the ocean surface (Moore *et al*., 1986; Murty, 2001). The landslide tsunami hypothesis cannot be wished away despite the current absence of definitive measurements. There have to be some landslide tsunamis happening every so often, and it is only a matter of time before we see these events for what they are. The hypothesis is here to stay.

LOCAL TSUNAMI WARNING SYSTEMS

As noted in an earlier section, the landslide tsunami hypothesis poses particular challenges for tsunami warning. To begin with, there is not necessarily any felt earthquake. Landslide tsunamis usually present a leading depression N-wave that would provide several minutes warning if seen by a coastal population. Moreover, the tsunami may have the form of a bore as it approaches the shoreline, which would likely be seen or heard, providing two further forms of tsunami warning. A coastal population educated in tsunami hazards can seek safe distances and safe elevations because of these natural forms of tsunami warning. However, a local tsunami warning system able to detect and locate underwater landslides could provide tens of minutes with which to enact an emergency response. While the warning times may be short relative to current transoceanic warnings, any such warning provides an opportunity to prepare for tsunami attack (e.g., closing oil transfer valves or walking to safety) and to enact a safe emergency response (e.g., when to send in emergency response personnel or where to station fire boats). Geological event specific evacuations become possible, impacting fewer coastal residents and reducing the chances of false alarms. Local tsunami warning systems force tsunami scientists to share their best information with emergency coordinators who need to know about all relevant tsunami sources in real time.

Shoreline or open water wave height measurements provide important information with which to issue tsunami warnings. However, the tsunami sources must be known in advance if one wishes to use this information effectively. Otherwise, the multitude of potential tsunami sources prevents an accurate warning based on a smaller number of discrete wave height measurements. In order for the inverse problem to be determinate, there have to be more measurement devices than significant tsunami sources. For example, one ocean floor pressure gauge cannot readily distinguish between an earthquake tsunami and three landslide tsunamis (Fryer *et al.*, 2001). A Green's function approach will not substantially improve tsunami warning accuracy unless all tsunami sources are superposed. The tsunami community should consider that equipment such as hydrophone arrays and techniques such as single force seismic inversion may ultimately be needed to identify and warn of landslide tsunamis directly. Otherwise, we are like seismologists without seismographs. Once we are set up to observe and measure underwater landslides directly, then we will have returned a significant amount of scientific explanations, certainty, and consensus to tsunamis science. Local tsunami warning systems will become commonplace, and tsunami warnings will be reliable. The landslide tsunami hypothesis will be proven.

As a hypothetical consideration, local tsunami warning systems might be set up to ask the following five questions:

> Did we detect a potential earthquake tsunami source?
> Did we detect a potential landslide tsunami source?
> What nearby locations need to receive a tsunami warning?
> Can we get several early measures of wave amplitude?
> What distant locations need to receive a tsunami warning?

These five questions form a cycle that may iterate until no more tsunami sources are detected and no more tsunami hazards exist. Iteration is necessary given the delays that can exist between earthquakes and mass failure (and *vice versa*). Every discipline now taking part in tsunami science has a role to play in answering one or more of these questions. And, we clearly need new measurement equipment in order to produce answers: ocean bottom pressure sensors **and** hydrophones in the SOFAR channel **and** coastal tsunami gauges. There is currently no magic bullet when it comes to figuring out what took place on the ocean floor. All forms of detection provide unique and valuable insight into the tsunami sources of a geological event.

The detection of underwater landslides argues for prediction of tsunami scenarios that include possible landslide tsunamis (see http://*www.tsunamicommunity.org* or Watts *et al*., 2001c). The scenarios are needed, among other reasons, to plan emergency responses to local tsunamis. This is especially important given the range of potential landslide tsunami amplitudes and local tsunami arrival times. Based on scenario results, the time of the maximum tsunami amplitude indicates the approximate time of the greatest tsunami hazard, which can vary substantially from place to place. Tsunami warnings should be issued in advance of this time. The last time of tsunami wetting is an important measure of overland inundation because it determines when it is finally safe to send in emergency response crews without exposing them to subsequent tsunami attack. Such a quantity is not usually provided as part of tsunami inundation studies. Landslide tsunami scenarios can provide valuable testing of local tsunami warning protocols and training of emergency response personnel.

CONCLUSIONS

A single geological event can give rise to many distinct tsunami sources. Marine geology will continue to play an important role in interdisciplinary tsunami research as more tsunami sources are recognized as originating from complicated geological events. Tsunamigenic underwater landslides almost certainly result from a confluence of tectonic, geological, sedimentary, and oceanic processes. More scientific explanations, certainty, and consensus will return to tsunami research as new underwater landslide measurement tools are deployed and local tsunami warning systems are devised. The landslide tsunami hypothesis is here to stay.

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A NON-LINEAR NUMERICAL MODEL FOR STRATIFIED TSUNAMI WAVES AND ITS APPLICATION

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ABSTRACT

A non-linear numerical model is developed for the computation of water level and discharge for the propagation of a unidirectional two-layered tsunami wave. Four governing equations, two for each layer, are derived from Euler's equations of motion and continuity, assuming a long wave approximation, negligible friction and no interfacial mixing. A numerical model is developed using a staggered Leap-Frog scheme. The developed non-linear model is compared with an existing validated linear model developed earlier by the author for different non-dimensional wave amplitudes. The significance of non-linear terms is discussed. It is found that for simulations of the interface wave amplitude, the effect of non-linear terms is not significant. However, for the simulation of the top surface, the effect of non-linear terms is significant for higher wave amplitudes, and insignificant for lower wave amplitudes. Developed non-linear numerical model is used for the case of a progressive internal wave in an inclined bay. It is found that the effect of an adverse bottom slipe towards the direction of wave propagation is to amplify the wave. This amplification depends on the steepness of slope as well as the ratio of densities of upper layer fluid to lower layer fluid (α). Amplification increases with slope. For higher values of α, amplification of the top and interface surface decreases, which is reasonable. It is also found that even for a 4 percent density difference between upper layer and lower layer, amplification of the top surface will be twenty times higher than amplification in the non-stratified case. The model can be applied confidently to simulate the basic features of different practical problems, similar to those investigated in this study.

INTRODUCTION

Tsunamis are generated due to disturbances of free surface caused not only by seismic fault motion, but also by landslides and volcanic eruptions (Imamura and Imteaz, 1995). Tsunamis are categorized as a long wave and as such, long wave theory has been applied for the governing equations of tsunami propagation considering a single layer (i.e. equal density throughout the depth). But even with respect to the density gradient in deep sea, it is necessary to consider the stratified layers. An exchange between fresh water and saline water is known to limit the amount of mixing that can occur at the mouth of an estuary. In the case of landslide generated flows it is imperative to consider the mudflow as stratified. Two-layered long waves or flows in cases where underwater landslides have generating tsunamis has been studied by Hampton (1972), Parker (1982) and Harbitz (1991).

Analytical and experimental studies generally involve the flow of current over a horizontal bottom, which neglects the change in terrain over which atmospheric currents move and the change in height associated with avalanches. Studies on two-layered flow which consider current flowing over a non-horizontal bottom has been investigated by Benjamin (1968), Britter & Linden (1980) and Lin Po-Ching (1990). However, the use of small-scale physical model to design large structures may not reproduce the relative effects of viscosity, Reynolds number and some of the other design parameters.

Simulating the behaviour of two-layered flow has also been attempted numerically, however, for simplicity non-linear terms are often ignored. Some examples of linear numerical model can be found in Akiyama et al. (1990), Kranenburg (1993) and Imamura & Imteaz (1995). But accurate results can not be expected until simulations account for non-linear terms.

Jiang & Leblond (1992) developed a numerical model coupling a submarine landslide and the surface waves it generated. They have assumed the landslide as laminar flow of an incompressible viscous fluid and the water motion as irrotational. Long wave approximations were adopted for both water waves and mudslides. They have shown that three main waves are generated by a landslide. The first wave is a crest which propagates away from the mudslide site into deeper water. This crest is followed by a trough in the form of a forced wave which propagates with the speed of the mudslide front and the third wave is a relatively small trough which propagates shoreward. They have also found that two major parameters dominate the interaction between the slide and waves it produces: the density of sliding material and the depth of water at the mudslide site. The findings are similar to those of Imamura & Imteaz (1995).

Considering non-linear terms an adequate numerical model for simulating two-layer flow on non-horizontal bottom is attempted. Moreover, the model was extended to actual or field condition.

NOTATIONS

ρ = Density of fluid
M = Discharge per unit width of flow
η = Water surface elevation above still water level
α = Density ratio of upper layer fluid to lower layer fluid
h = Still water depth
x = Distance in downstream direction
t = Axis representing time

g = Acceleration due to gravity
i = Spatial node points in a finite difference scheme
n = Temporal node points in a finite difference scheme
β = Depths ratio of lower layer to upper layer
k = Wave number
L = Wave length
a = wave amplitude

THEORETICAL BACKGROUND

A mathematical model for two-layer flow in a wide channel with non-horizontal bottom was set up assuming a hydrostatic pressure distribution, negligible friction and negligible interfacial mixing. Also uniform density and velocity distributions in each layer was assumed. Considering a two dimensional case as shown in Figure 1, using Euler's equation of motion and continuity for each layer, by integrating the equations for specified limit of each layer and applying long wave approximation (i.e. vertical accelerations are negligible) and boundary conditions, following integrated governing equations were derived. The derivations are explained in details by Imteaz, M.A. (1994).

For upper layer-
Mass conservation equation,

$$\frac{\partial M_1}{\partial x}+\frac{\partial(\eta_1-\eta_2)}{\partial t}=0 \tag{1}$$

and momentum equation,

$$\frac{\partial M_1}{\partial t}+\frac{\partial(M_1^2/D_1)}{\partial x}+gD_1\frac{\partial \eta_1}{\partial x}=0 \tag{2}$$

For lower layer-

Mass conservation equation, $\frac{\partial M_2}{\partial x}+\frac{\partial \eta_2}{\partial t}=0$ (3)

and momentum equation,

$$\frac{\partial M_2}{\partial t}+\frac{\partial(M_2^2/D_2)}{\partial x}+gD_2\left\{\alpha\left(\frac{\partial \eta_1}{\partial x}+\frac{\partial h_1}{\partial x}-\frac{\partial \eta_2}{\partial x}\right)+\frac{\partial \eta_2}{\partial x}-\frac{\partial h_1}{\partial x}\right\}=0 \tag{4}$$

Where,
η_1 = Water surface elevation above still water level of layer '1'
η_2 = Water surface elevation above still water level of layer '2'
$D_1 = \eta_1 + h_1 - \eta_2$
$D_2 = h_2 + \eta_2$
h_1 = Still water depth of layer '1'
h_2 = Still water depth of layer '2'
$\alpha = \rho_1/\rho_2$

$$M_1=\int_{-h_1+\eta_2}^{\eta_1} u_1\,dy,\quad M_2=\int_{-h_1-h_2}^{-h_1+\eta_2} u_2\,dy$$

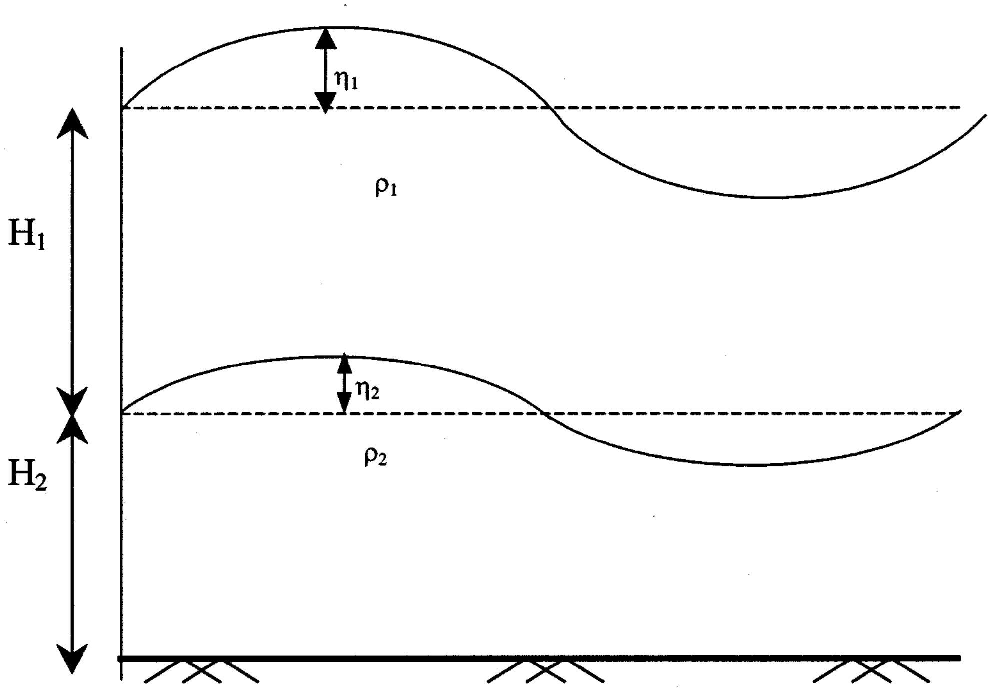


Figure 1 Schematic diagram of a two-layer profile

NUMERICAL MODEL

It is very difficult to solve the governing non-linear equations analytically, however, a finite difference scheme using a staggered leap-frog scheme can provide a numerical solution. This scheme has been used previously for the solution of linearized governing equations with good results (Imamura & Imteaz, 1995). This scheme is one of the explicit central difference scheme with the truncation error of second order. The staggered scheme considers that the computation points for the water surface elevation (η) does not coincides with the computation points for discharge (M). There are half step differences (½Δt for temporal and ½Δx for spatial) between computation points of two variables, η and M (as shown in Figure 2). Finite difference equations for this scheme are as follows.

Mass conservation equation for the upper layer,

$$\frac{\eta_{1,i}^{n+1/2} - \eta_{1,i}^{n-1/2} - \eta_{2,i}^{n+1/2} + \eta_{2,i}^{n-1/2}}{\Delta t} + \frac{M_{1,i+1/2}^{n} - M_{1,i-1/2}^{n}}{\Delta x} = 0 \qquad (5)$$

Mass conservation equation for the lower layer,

$$\frac{\eta_{2,i}^{n+1/2} - \eta_{2,i}^{n-1/2}}{\Delta t} + \frac{M_{2,i+1/2}^{n} - M_{2,i-1/2}^{n}}{\Delta x} = 0 \qquad (6)$$

Momentum equation for the upper layer,

$$\frac{M_{1,i+1/2}^{n} - M_{1,i+1/2}^{n-1}}{\Delta t} + g\frac{D_{1,i+1}^{n-1/2} + D_{1,i}^{n-1/2}}{2}\,\frac{\eta_{1,i+1}^{n-1/2} - \eta_{1,i}^{n-1/2}}{\Delta x} +$$

$$\frac{\dfrac{(M^{n-1}_{1,i+1/2})^2}{(D^{n-1/2}_{1,i+1}+D^{n-1/2}_{1,i}+D^{n-3/2}_{1,i+1}+D^{n-3/2}_{1,i})/4}-\dfrac{(M^{n-1}_{1,i-1/2})^2}{(D^{n-1/2}_{1,i}+D^{n-1/2}_{1,i-1}+D^{n-3/2}_{1,i}+D^{n-3/2}_{1,i-1})/4}}{\Delta x}=0 \qquad (7)$$

Momentum equation for the lower layer,

$$\frac{M^{n}_{2,i+1/2}-M^{n-1}_{2,i+1/2}}{\Delta t}+g\frac{(D^{n-1/2}_{2,i+1}+D^{n-1/2}_{2,i})}{2}$$

$$\frac{\alpha(\eta^{n-1/2}_{1,i+1}-\eta^{n-1/2}_{1,i})+(\alpha-1)(h_{1,i+1}-h_{1,i})+(1-\alpha)(\eta^{n-1/2}_{2,i+1}-\eta^{n-1/2}_{2,i})}{\Delta x}+$$

$$\frac{\dfrac{(M^{n-1}_{2,i+1/2})^2}{(D^{n-1/2}_{2,i+1}+D^{n-1/2}_{2,i}+D^{n-3/2}_{2,i+1}+D^{n-3/2}_{2,i})/4}-\dfrac{(M^{n-1}_{2,i-1/2})^2}{(D^{n-1/2}_{2,i}+D^{n-1/2}_{2,i-1}+D^{n-3/2}_{2,i}+D^{n-3/2}_{2,i-1})/4}}{\Delta x}=0 \qquad (8)$$

Where 'n' denotes the temporal grid points and 'i' denotes the spatial grid points as shown in Figure 2. To calculate 'D' values at the computation point of 'M', the average of four surrounding 'D' values is required.

In the spatial direction all η_1, η_2 values at step 'n-1/2' and all M_1, M_2 values at step '(n-1)' are assigned as initial conditions. For all later time steps at left and right boundaries, values of M_1 and M_2 are calculated by a characteristic method, which uses the values of previous time steps and wave celerity. The finite difference momentum equations for the upper and lower layer allows M_1 and M_2 values at step 'n' to be calculated. Then using the latest values of M_2 and deduced finite difference continuity equation for lower layer all values of η_2 at step '(n+1/2)' can be calculated. Then using the latest values of η_2, M_1 and deduced finite difference continuity equation for upper layer, all values of η_1 at step '(n+1/2)' are calculated. Similarly using new values of η_1, η_2, M_1, M_2 as initial conditions, calculations can be proceeded in time direction up to desired step. As initial condition (i.e. at $t = 0$) all η_1 and M_1 values are taken as zero. For interface, initial condition was found by substituting $t = 0$ in the known expression of η_2, which gives,

$$\eta_2 = a_2 \operatorname{Sin}(kx) = a_2 \operatorname{Sin}(\frac{2\pi}{L}x)$$

A linear relationship between the water level and discharge is used to determine an expression for M_2 i.e.,

$$M_2 = \sqrt{(g/H_2)}\,\eta_2(H_2+\eta_2)$$

COMPARISON WITH LINEAR MODEL

Simulations of the numerical model were performed for different values of initial interface wave amplitude, a_2. For these simulations, α=0.2, β=1.0, ΔX=10 m, ΔT=0.2 sec, h_1=25 m, h_2=25 m have been chosen and kept constant for different simulations. Simulations were performed separately for a_2= 2 m, 6 m and 8 m with periods of 4 sec, 6 sec and 8 sec. Results were compared with known linear model results using the same input parameters. Results are shown graphically in Figures 3, 4 and 5 for a_2 = 2 m, 6 m and 8 m respectively. From the figures it is seen that for a_2 = 2 m (a_2/h_2=0.08) linear and non-linear model results are almost identical. This indicates the insignificance of non-linear terms and confirms that for $\eta/h < 0.1$, the effect of non-linear terms is negligible.

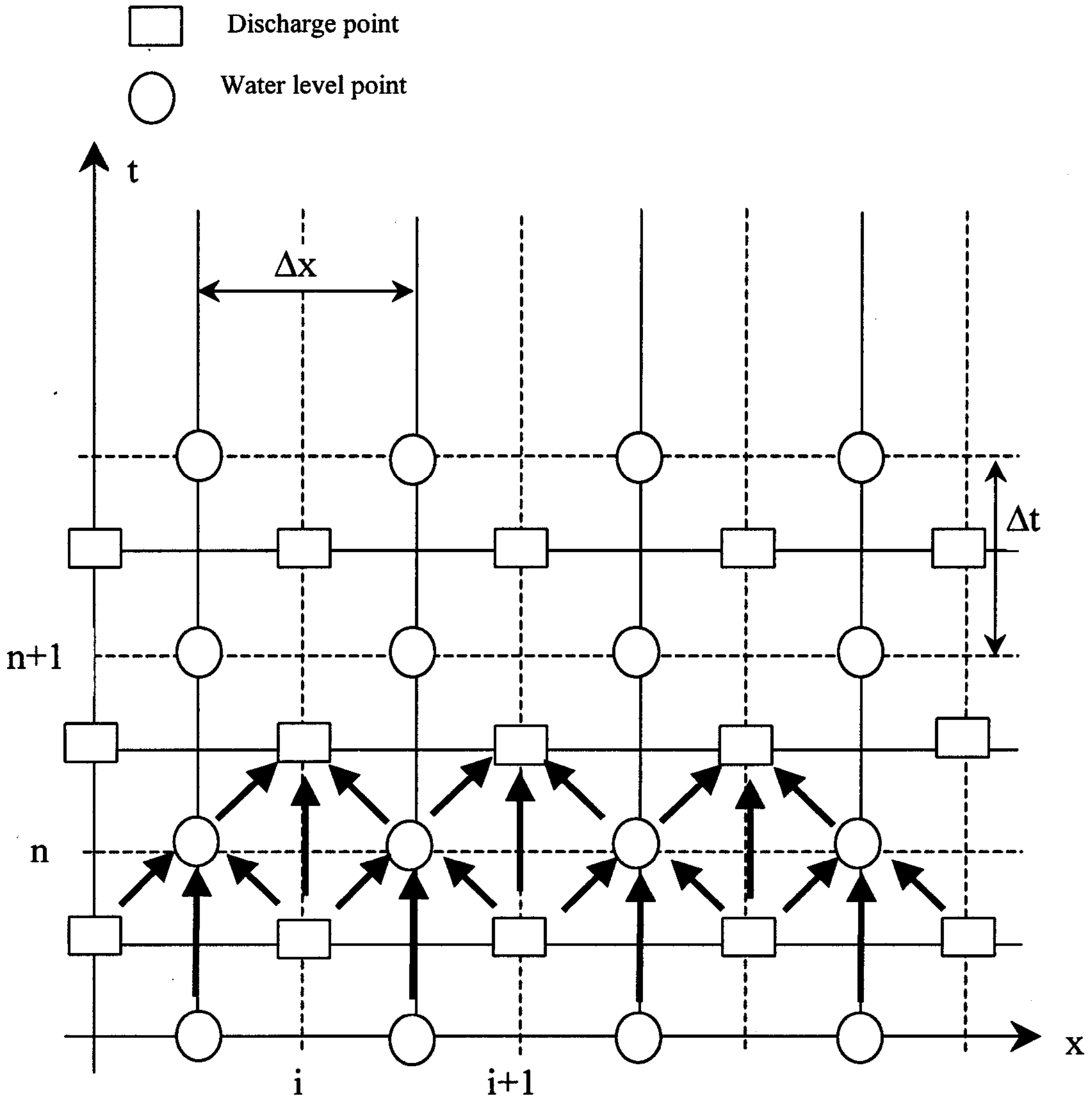
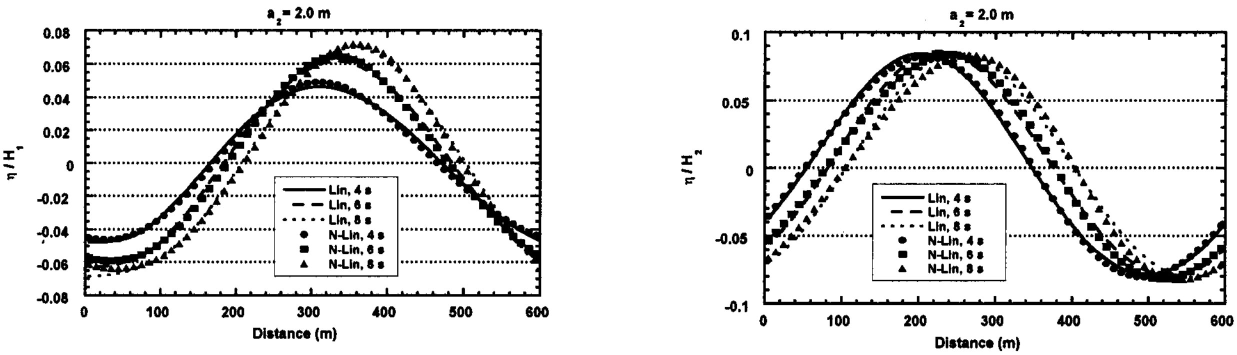


Figure 2 Schematic diagram of the staggered leap-frog scheme

For simulations with $a_2 = 6$ m and 8 m, it is found that there is a significant difference between linear and non-linear model, proving that for $\eta/h > 0.1$ non-linear terms are significant. For the modelled interface differences arise as a result of the wave celerity. Due to inclusion of η in the wave celerity, $\sqrt{\{g(h+\eta)\}}$, non-linear wave celerity is greater than the linear wave celerity. For this reason it is seen that non-linear waves propagate with greater speed and become steeper as time progresses. The above mentioned reasoning also accounts for the marked difference between linear and non-linear top surface levels, η_1. It is observed that this difference increases with the value of interface amplitude, a_2. For higher values of a_2, values of η_1 will increase. As can be seen any increase a_2 of values will cause increase of η_1 values, (i.e increase of η_1/h_2 values) and for higher η_1/h_2 non-linear terms become more significant.

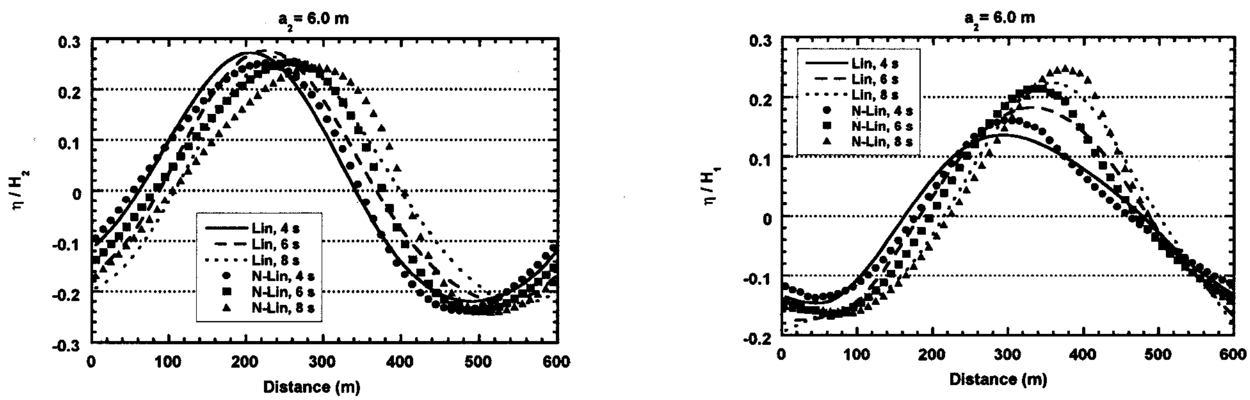
APPLICATION OF THE MODEL

A progressive internal wave into an inclined bay, which is related with a sudden exchange of water in the bay was considered as an application for the numerical model. This condition may occur due to an internal tide wave into the bay. For this condition a sloping bottom is assumed with a vertical wall at the downstream end. As an upstream boundary condition, the interface water surface is assumed to be known and a function of time. A downstream boundary condition involves setting the discharge of both layers, M_1 and M_2 as zero.



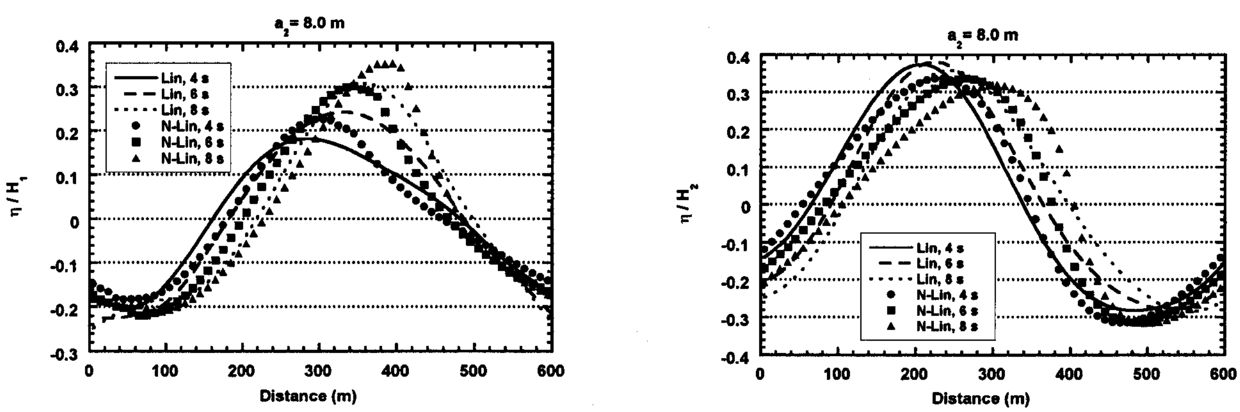
(a) (b)

Figure 3 Comparison of linear and non-linear model for (a) top surface and (b) interface (a_2=2 m)



(a) (b)

Figure 4 Comparison of linear and non-linear model for (a) top surface and (b) interface (a_2=6 m)

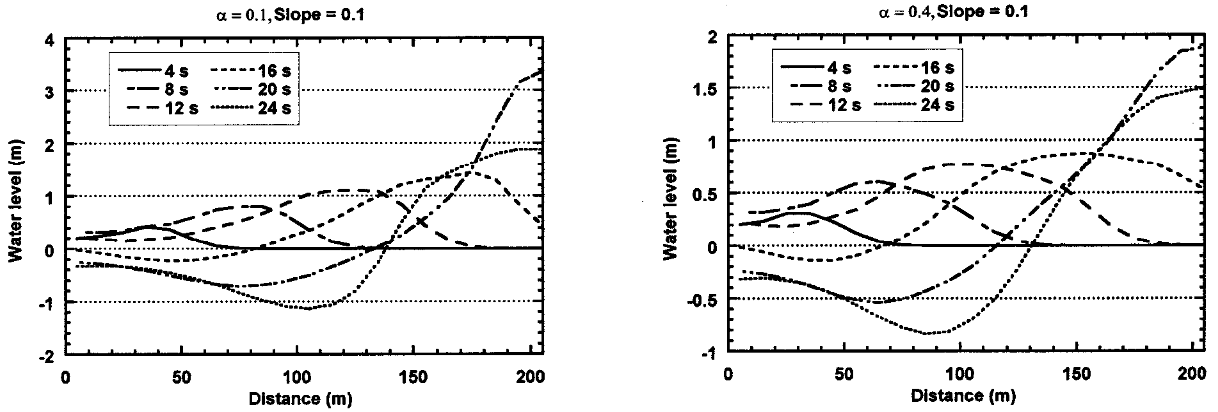


(a) (b)

Figure 5 Comparison of linear and non-linear model for (a) top surface and (b) interface (a_2=8 m)

In general it was found that a sloping bottom amplifies wave as it propagates. So for longer travel distance amplification will be higher. Simulations were performed for different slopes and different 'α' values. To show the effect of 'α', model results for α = 0.1, 0.4, keeping other parameters constant have shown in Figure 6(a) and 6(b). In these cases slope was assumed as 0.1, computational length 200 m, u/s h_2= 25 m, at d/s h_2= 5 m and h_1= 10 m (constant). From these figures it is clear that when 'α' increases the amplification of top surface decreases and vice versa. For interface amplification decreases as 'α' increases, which is opposite as discussed by Imamura & Imteaz (1995) for the case of free transmission (no vertical wall at d/s end). This is due to presence of vertical wall and rigorous interaction at d/s end by the reflected upper layer wave. To see the effect of stratification, model was again simulated for α=1.0 (i.e. same density for both the layers) keeping other conditions same as in Figure 6(a) and 6(b). From Figure 7(a) it is found that amplification of top surface is very low for this case, which is not real, as in reality there is always some stratification. Figure 8 shows the ratio of top surface amplitudes of stratified case to non-stratified case for several density difference in percent. It is shown that even for a 4% density difference between upper layer and lower layer, amplification of top surface will be twenty times higher than the amplification in non-stratified case. This warns serious error in calculation if uniform water density is considered for the case of Tsunami and long waves.

To show the effect of slope change on amplification, simulation was carried out for two different slopes, keeping other parameters (ΔX, ΔT, α, u/s and d/s h_1, u/s h_2) constant. At d/s h_2 can not be constant to allow the variation of slope and to keep the h_2 at u/s, computational length constant. Computational length should be keep constant, the amplification at the d/s end is the cumulative effect of slope used, so change in computational length will affect the result. For the same α Figure 6(a) shows the amplification of top surface for slope equals to 0.1 and Figure 7(b) shows for slope equals to 0.075. It is found from these figures that as slope decreases amplification of both top surface and interface also decreases which is reasonable. Presented results are not indicating the sole effect of slope variation, because there is a change of depth 'h_2' at right side boundary. At right side 'h_2' has to increase to provide decreasing slope in computation. But 'h_2/h_1' ratio also affects the amplification of top surface. In the above-mentioned cases, change of h_2/h_1 at right side boundary causes opposite the effect of slope changes.



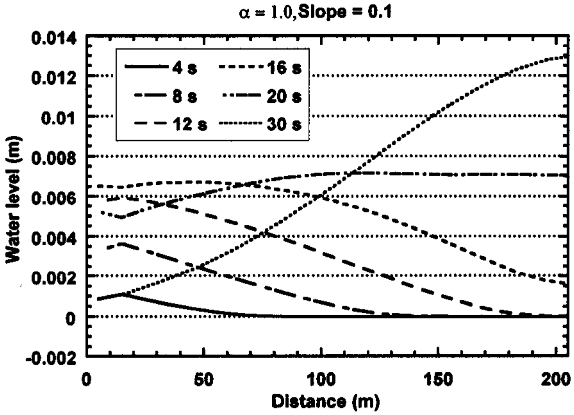
(a) (b)

Figure 6 Amplification of the top surface for (a) α= 0.1 and (b) α=0.4 having slope=0.1

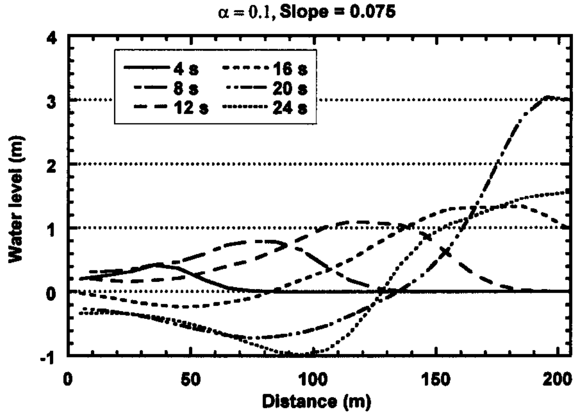
CONCLUSION

A numerical model using a staggered leap-frog scheme has been used to ascertain the effect of including non-linear terms in the governing equations of two-layered flow. Comparison has been made against the linear model of Imteaz, M.A. (1994). Non-linear terms were found to be insignificant for low (< 0.1) η/h_2 ratio, however non-linear terms are highly significant for higher η/h_2 ratio. Therefore in the cases of high η/h_2 ratio and where amplitude amplifies due to slope or vertical wall, non-linear model should be used to simulate the actual condition.

The developed model was applied to the case of rapid exchange of water in the bay, considering non-horizontal bottom, two-layer flow and vertical wall at the shore. In this case constant bottom slope and horizontal interface was assumed, but model can handle for cases of any arbitrary bottom and any arbitrary interface as well as top surface. Effects of sloping bottom and relative density of two layers were investigated. It was found that, presence of adverse slope to the direction of wave propagation causes to amplify the propagating wave as wave propagates, which is reasonable. Again it was found that as 'α' increases amplification of top surface decreases and vice versa, which is reasonable. It must also be noted that the phenomena of shoaling for a single layer flow can not be used as a comparison for the investigated two layer flow.



(a)



(b)

Figure 7 Amplification of the top surface for (a) α= 1, slope=0.1 and (b) α= 0.1, slope=0.075

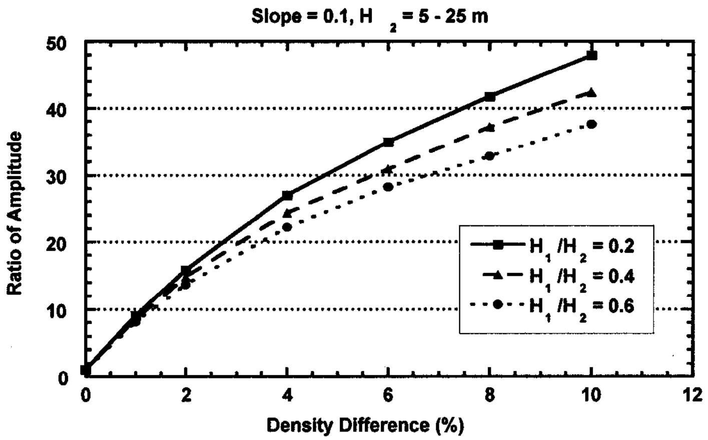


Figure 8 Ratio of top surface amplitudes for stratified case to non-stratified case

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MODELING THE LA PALMA LANDSLIDE TSUNAMI

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ABSTRACT

The tsunami expected from a lateral collapse of the Cumbre Vieja Volcano on La Palma in the Canary Islands was modeled. The flank collapse for a 'worst case" landslide was modeled as a 650 meter high, 20 kilometer radius water wave after 30 kilometers of travel as predicted by physical modeling studies of Fritz at ETH in Zurich, Switzerland.

The modeling was performed using the *SWAN* code which solves the nonlinear long waver equations. The tsunami generation and propagation was modeled using a 10 minute Mercator grid of 600 by 640 cells. The small wavelength and period of the tsunami expected from the landslide source results in an intermediate wave rather than a shallow water tsunami wave. The use of a shallow water model only describes the geometric spreasing of the wave and not the significant dispersion such a short period wave would exhibit. Dispersion would reduce the wave amplitudes to less than one-third of the shallow water amplitudes.

The upper limit shallow water modeling indicates that the east coast of the U.S.A. and the Caribbean would receive tsunami waves less than 3 meters high. The European and African coasts would have waves less than 10 meters high.

Full Navier-Stokes modeling including dispersion and geometric spreading for the Fritz initial wave profile predicts that the maximim wave amplitude off the U.S. east coast would be about a meter. Even with shoaling the wave would not present a significant hazard.

INTRODUCTION

The lateral collapse of the flank of the Cumbre Vieja Volcano on La Palma in the Canary Islands represents the "worst case" La Palma submarine landslide. This La Palma slide involves 500 cubic kilometers of material runing out 60 kilometers at a mean speed of 100 meters/second. Hermann Fritz of the Swiss Federal Institute of Technology (ETH) in Zurich, Switzerland has physically modeled the landslide as one single block and determined that the landslide would generate a 650 meter high wave with a wavelength of 30 to 40 kilometers and period of 3 to 4 minutes. The wave profile he obtained is shown in Figure 1. It was measured after the wave had effectively traveled 30 kilometers or about 0.04 hours from the source.

He states that such a wave is not a shallow water tsunami wave but an intermediate wave that would undergo significant dispersion as it propagated away from the source region. The experimental apparatus created by Fritz and used to physically model the La Palma landslide is described in reference 1.

If one considers only geometric spreading in a constant depth basin then to an upper limit approximation the wave height as a function of distance from the source would be

$$H = H_o \frac{R_o}{R}$$

where R_o is the initial source radius, R is the distance the wave has traveled from the source, and H_o is the initial height of the source. For the La Palma event R_o is 20 kilometers, and H_o is 650 meters. The wave height after 1000 kilometers of travel is 13 meters, after 5000 kilometers is 2.6 meters and after 10,000 kilometers is 1.3 meters.

To obtain another estimate of the effect of geometric spreading throughout the Atlantic Ocean and of the upper limit wave amplitude, shallow water numerical modeling was performed using the actual depths throughout the Atlantic Ocean.

In addition to geometric spreading the short wavelength and short period wave would undergo significant dispersion as it traveled. To a first approximation dispersion would reduce the heights to less than one-third of the above values as shown in reference 2 and 3 and discussed in the section on dispersion modeling.

For comparision, the great Lisbon tsunami was caused by the November 1, 1755 earthquake generated a tsunami which arrived at Lisbon between 40 minutes and one hour after the earthquake as a withdrawing wave that emptied the Lisbon Oeiras Bay to more than a mile out. Then a tsunami wave with an amplitude of about 20 meters arrived followed by two more waves about an hour apart. The tsunami wave had a period of one hour and amplitude of up to 20 meters at Lisbon and along the African and south European coasts, of 4 meters along the English coast, and of 7 meters at Saba in West Indies after 7 hours of travel. To reproduce the observed tsunami wave characteristics a source source 300 kilometers in radius with a drop of 30 meters is required to be located in the region of the 1969 earthquake near the Gorringe bank as described in reference 4.

The May 23, 1960 Chile earthquake had a magnitude of 8.5 and generated a tsunami with a 25 minute period from a source about 150 kilometers wide and 800 kilometers long or 120,000 sq kilometers. The March 26, 1964 Alaskan earthquake near Prince William sound had a magnitude of 8.4 and generated a tsunami wave with a 30 minute period from a source about 300 kilometers wide and 800 kilometers long or 240,000 square kilometers.

To a first approximation the tsunami wave period is determined primarily by the width of the ocean floor displaced by the earthquake and the wave amplitude by the height of the ocean floor displacement.

SHALLOW WATER MODELING

The modeling was performed using the *SWAN* non-linear shallow water code which includes Coriolis and frictional effects. The *SWAN* code is described in Reference 5. The calculations were performed on 866 Mhz Pentium personal computers with 128 megabytes of memory.

The 10 minute Atlantic topography was generated from the 2 minute Mercator Global Marine Gravity topography of the earth of Sandwell and Smith of the Scripps Institute of Oceanography and described in Reference 6. The grid was 600 by 640 cells with the left hand corner at 20 S, 100 W. The grid extended from 20 N to 65 N and from 100 W to 0 W. The time step was 10 seconds.

The source used in the calculations was a 20 kilometer radius of 650 meter high water located near La Palma island. This source is the same height but larger in area than the Fritz wave shown in Figure 1.

The Table 1 locations are shown in Figure 2. The units of the X and Y axis are 10 kilometers. The travel time chart is shown in Figure 3.

The maximum deep water amplitudes at various locations are given in Table 1. The positive wave arrives first followed by several waves with the first wave usually having the maximum amplitude.

As shown in Reference 5 the run-up amplification may be from 2 to 3 times the deep water wave amplitude. Locations with depths less than 1000 meters have some of the run-up amplification included. Dispersion would reduce the wave amplitudes by more than the run-up amplification so the upper limit amplitudes in Table 1 are also upper limit values after run-up.

The east coast of the U.S.A. and the Caribbean receive a tsunami wave with an upper limit less than 3 meters high. The European and African coasts receive a tsunami with an upper limit less than 10 meters high.

Computer generated animations of these calculations are available at http://t14web.lanl.gov/Staff/clm/tsunami.mve/tsunami.htm.

The use of a shallow water model to describe the propagation of waves that are really intermediate water waves dispersing to deep water waves as they travel can only furnish an upper limit estimate of the amplitude and period of the waves. To obtain more realistic wave characteristics a full Navier-Stokes model is needed.

The assumption that the Fritz wave represents the initial displacement at the source is an approximation that needs to improved as the wave was actually equivalent to the wave after it had traveled 30 kilometers of 0.0376 hours. An initial displacement was determined that would approximately reproduce the Fritz wave at 30 kilometers and the calculations repeated to obtain a more realistic model.

DISPERSION MODELING

To obtain an estimation of the effect of dispersion the procedures described in reference 5 were used.

The linear gravity wave model solves the two-dimensional linear gravity wave with a Gaussian or a square wave displacement using Fourier transforms for any time of interest. The wave description is obtained for any uniform length, density, gravity and Gaussian break width or square wave half-width. The model is two-dimensional and symmetrical about the center of the initial displacement, only half of the wave profile is calculated and thus the dispersion effects are described without any geometric spreading.

To model the dispersion effects for the case modeled previously using the SWAN code, a 650 meter high, 20 kilometer square half-width, water displacement in 5000 meter deep water was studied. The initial profile and the wave at 0.5, 2.0, 5.0 and 10.0 hours (400, 1600, 4000, 8000 km) are shown in Figure 5. The wave disperses into a train of waves with a maximum amplitude of 320, 220, 180, 150 meters or from one-half to one-fourth of the initial wave amplitude.

A 650 meter high Gaussian surface water displacement in 5000 meter deep water with Gaussian break of 10km (which is equivalent to an Airy wave with a half-wavelength of 45 kilometers) is a good approximation of the Fritz wave shown in Figure 1. The initial profile and the wave at 0.5, 2.0, 5.0 and 10.0 hours (400, 1600, 4000, 8000 km) are shown in Figure 5. The wave disperses into a train of waves with a maximum amplitude of 220, 150, 110, 90 meters or from one-third to one-eighth of the initial wave amplitude.

The ZUNI code solves the two-dimensional time dependent Navier-Stokes equations for incompressible flow and is described in detail in reference 5.

To approximate the Fritz wave shown in Figure 3, a 650 meter high Airy wave surface water displacement in 5050 meter deep water with with a half-wavelength of 40 kilometers was modeled using ZUNI for 3000 kilometers of travel. The calculations for this geometry were performed with 15 cells in the Y or depth direction and 5000 cells in the R or radial direction. The cells were 500 meters high in the Y direction and 1000 meters long in the radial direction. The time step was 0.1 seconds. These calculations are an extension of those described in references 2, 3, and 5.

The calculated wave profiles are shown in Figure 6 after various times and distance of travel. The calculated wave amplitude as a function of time and distance is shown in Table 2. The wave amplitude with geometric spreading and diffusion has been reduced to 2.6 meters after 2500 kilometers of travel. The wave amplitude after 10,000 kilometers of travel would be about half a meter.

FRITZ SOURCE AT 30 KM MODEL

The Fritz wave in Figure 1 was measured by a gauge in front of the landslide that was at an equivalent location of 30 kilometers from the slide.

In the previous study the source was described as a displacement of water with the geometry of the Fritz wave and centered at the source. The wave observed by Fritz was 2-D and would not exhibit geometric spreading like a 3-D one with distance of travel. The 2-D linear gravity wave model has 320 meters amplitude after 24 km of travel for the Gaussian wave and 460 meters for the square wave. To more closely approximate the wave Fritz observed at 30 km, the initial 40 km diameter square displacement amplitude needs to be increased to 990 meters as shown in Figure 7 and the Gaussian displacement to 1400 meters as shown in Figure 8.

Using the linear gravity model to model the dispersion effects without geometric spreading, the wave profile at 0.5, 2.0, 5.0 and 10 hours are shown in Figures 9 and 10. The amplitudes are increased by a factor of about 2.0 for the Gaussian model and a factor of about 1.4 for the square displacement over those of Figures 4 and 5.

Using the square displacement with a 990 meter initial height that gave an approximately 650 meter high wave at 30 km, the SWAN calculation gave the wave heights shown in Table 3 which are about 1.52 times higher than those of Table 1. Using the Airy displacement with a 1400 meter initial height that gave an approximately 650 high wave at 30 km, the ZUNI calculation gave the wave heights shown in Table 4 which are about 2.15 times larger than those of Table 2. The wave profiles as a function of distance are shown in Figure 11. The wave amplitude off the U.S. coast would be less than one meter. Even with shoaling the wave would not present a significant hazard.

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2. Charles L. Mader, Dennis W. Moore, George F. Carrier *Numerical Tsunami Propagation Study - III*, Science of Tsunami Hazards, **11**, 93-106 (1993).
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5. Charles L. Mader *Numerical Modeling of Water Waves,* University of California Press, Berkeley, California (1988).
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TABLE 1
UPPER LIMIT WAVE HEIGHTS
FOR FRITZ LA PALMA TSUNAMI

No	Depth Meters	Location	Maximum Amplitude Meters	Minimum Amplitude Meters
1	953	Off Lisbon	+9.0	-10.
2	4747	East of Saba	+2.2	-2.2
3	825	East of Saba	+2.0	-1.9
4	3446	North of San Juan	+1.8	-1.6
5	783	East of Miami	+0.3	-0.2
6	2922	East of Washington	+3.0	-3.0
7	178	South West of England	+2.5	-1.9
8	4574	West of Lisbon	+5.0	-5.0
9	3868	West of Lagos	+4.0	-2.0
10	3923	West of Gibraltar	+6.0	-5.0
11	4376	West of Gibraltar	+6.0	-4.0
12	1717	West of Casablanca	+7.0	-4.0
13	3314	N-W of Source	+8.0	-6.0

DISPERSION WOULD REDUCE HEIGHTS TO

LESS THAN ONE THIRD OF ABOVE VALUES

TABLE 2
ZUNI WAVE HEIGHTS
FOR FRITZ LA PALMA TSUNAMI

No	Time Hours	Radial Distance Meters	Amplitude Meters	Amp/650. Percent
1	.03	24,000	170.	26.
2	.05	40,000	120.	18.
3	.10	80,000	71.	11.
4	.125	100,000	60.	9.2
5	.250	200,000	30.	4.6
6	.625	500,000	14.	2.1
7	1.25	1,000,000	7.0	1.1
8	1.88	1,500,000	4.6	0.71
9	2.50	2,000,000	3.4	0.52
10	3.12	2,500,000	2.6	0.4

TABLE 3
WAVE HEIGHTS
FOR FRITZ LA PALMA TSUNAMI AT 30KM

No	Depth Meters	Location	Maximum Amplitude Meters	Minimum Amplitude Meters
1	953	Off Lisbon	+14.0	-15.
2	4747	East of Saba	+3.2	-3.0
3	825	East of Saba	+2.9	-2.5
4	3446	North of San Juan	+2.2	-2.0
5	783	East of Miami	+0.7	-0.5
6	2922	East of Washington	+4.1	-4.0
7	178	South West of England	+3.5	-2.0
8	4574	West of Lisbon	+7.0	-8.0
9	3868	West of Lagos	+6.0	-3.0
10	3923	West of Gibraltar	+8.0	-7.0
11	4376	West of Gibraltar	+8.0	-6.0
12	1717	West of Casablanca	+10.0	-6.0
13	3314	N-W of Source	+12.0	-9.0

DISPERSION WOULD REDUCE HEIGHTS TO

LESS THAN ONE THIRD OF ABOVE VALUES

TABLE 4
ZUNI WAVE HEIGHTS
FOR FRITZ LA PALMA TSUNAMI AT 30 KM

No	Time Hours	Radial Distance Meters	Amplitude Meters	Amp/1400. Percent
1	.03	24,000	370.	26.
2	.05	40,000	272.	19.
3	.10	80,000	160.	11.
4	.125	100,000	133.	9.5
5	.250	200,000	70.	5.0
6	.625	500,000	30.	2.1
7	1.25	1,000,000	15.0	1.1
8	2.50	2,000,000	7.3	0.52
9	4.03	3,228,000	4.2	0.30
10	5.00	4,005,000	3.2	0.21
11	5.70	4,565,700	2.6	0.18

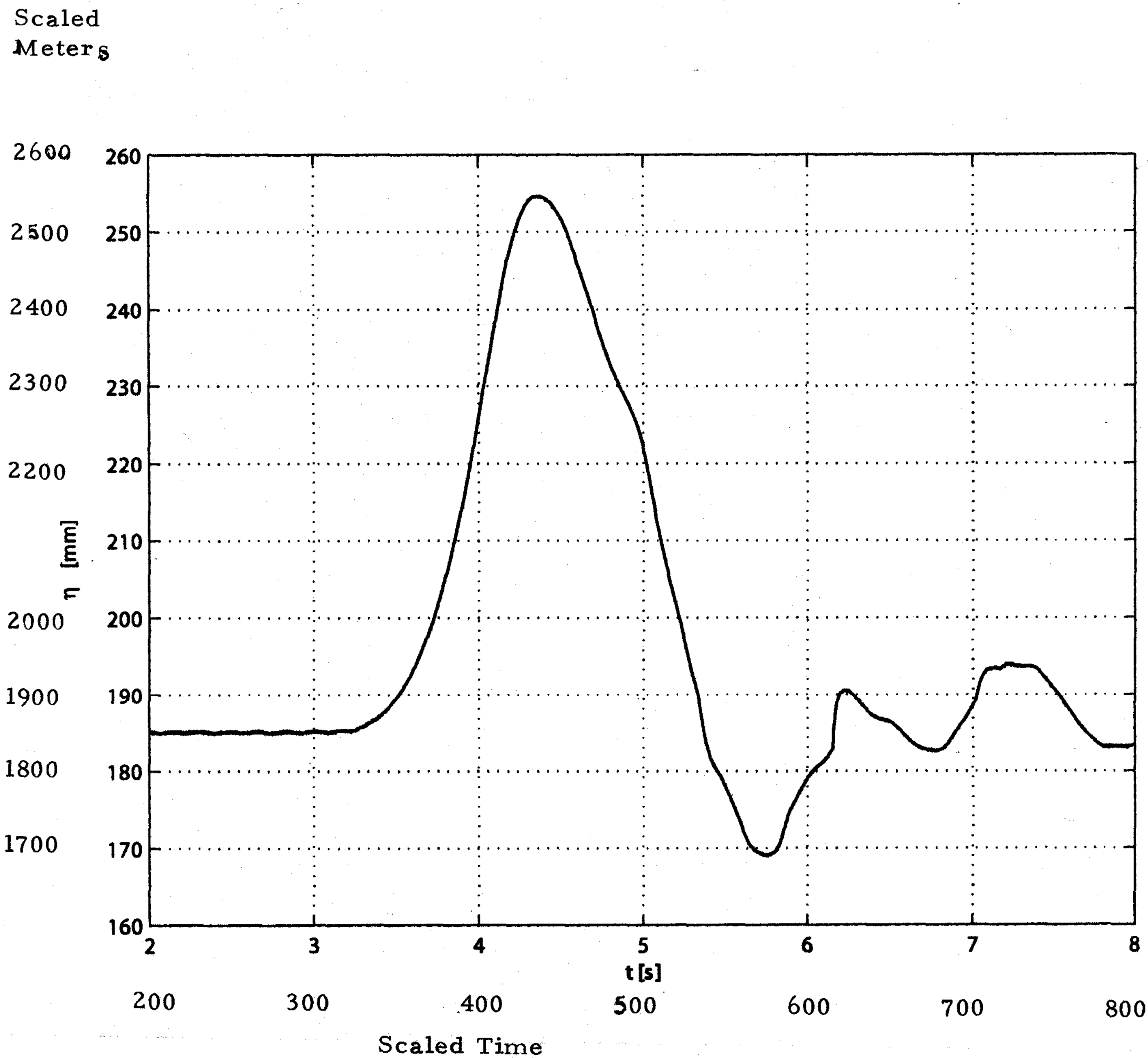


Figure 1. Fritz LaPalma Source Wave record.

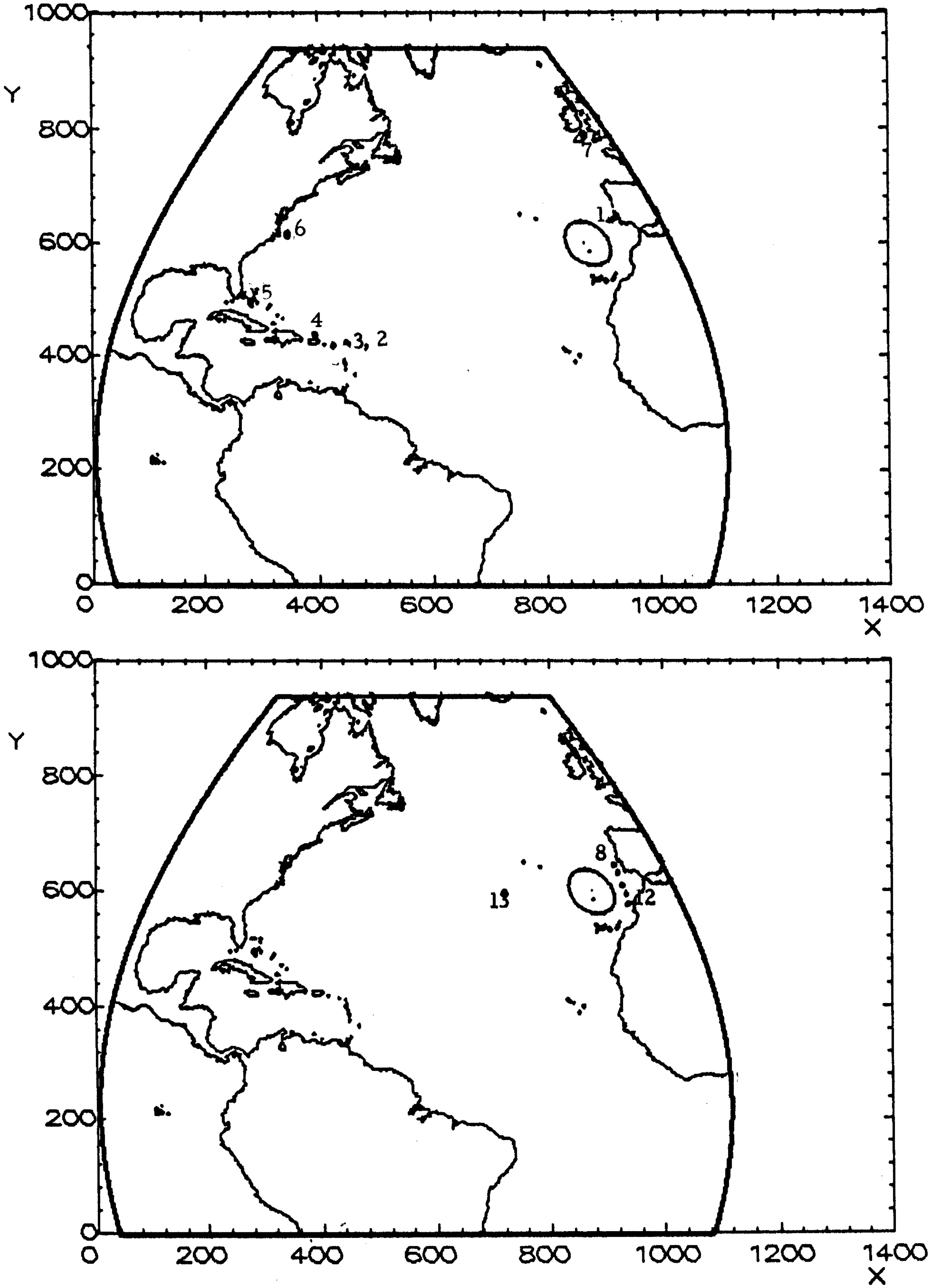


Figure 2. The source region and Table 1 locations. The X and Y axis units are 10 kilometer

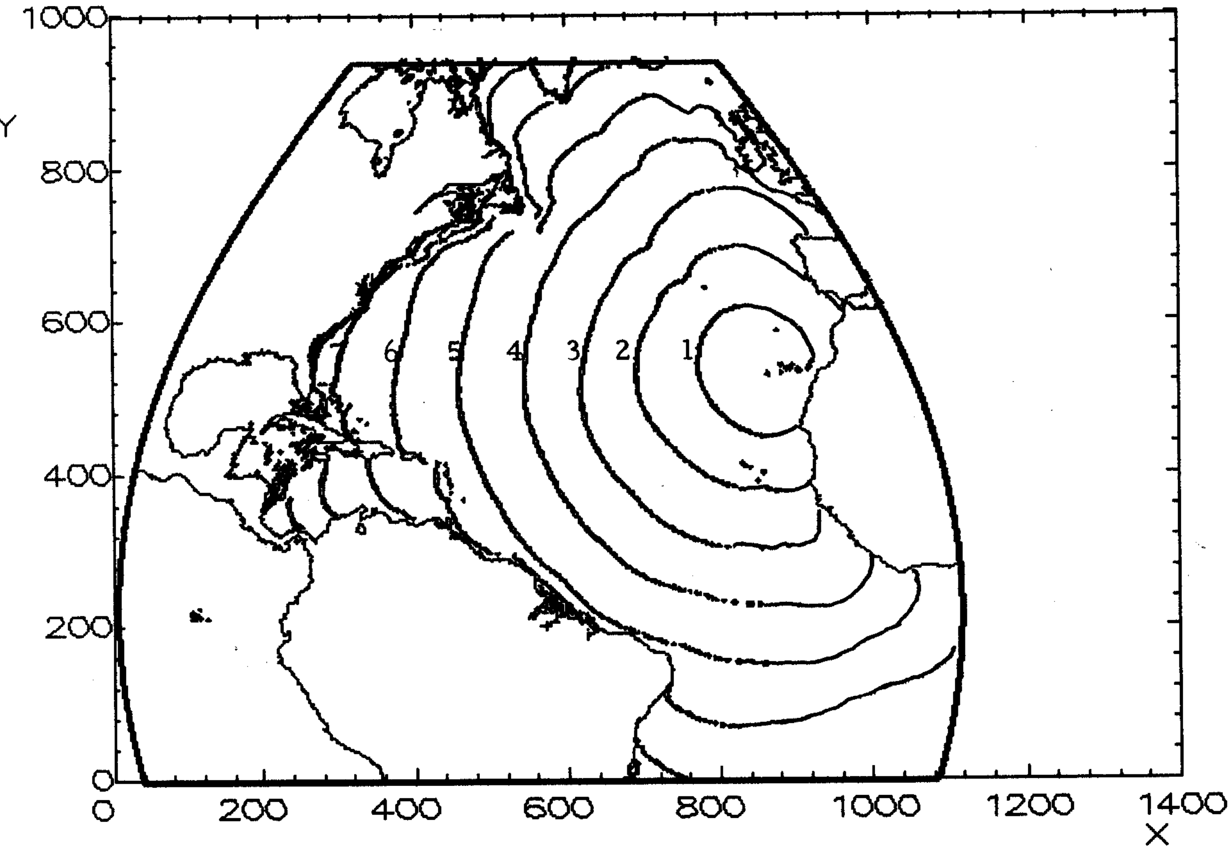
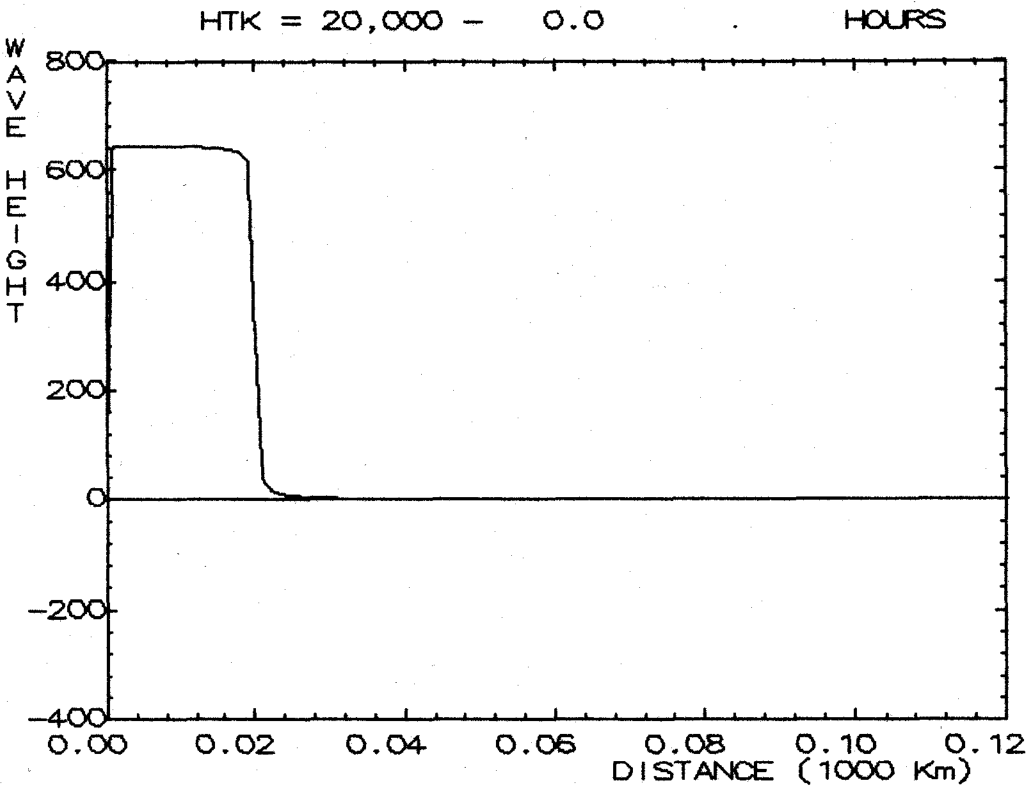


Figure 3. The travel time chart at one hour intervals.



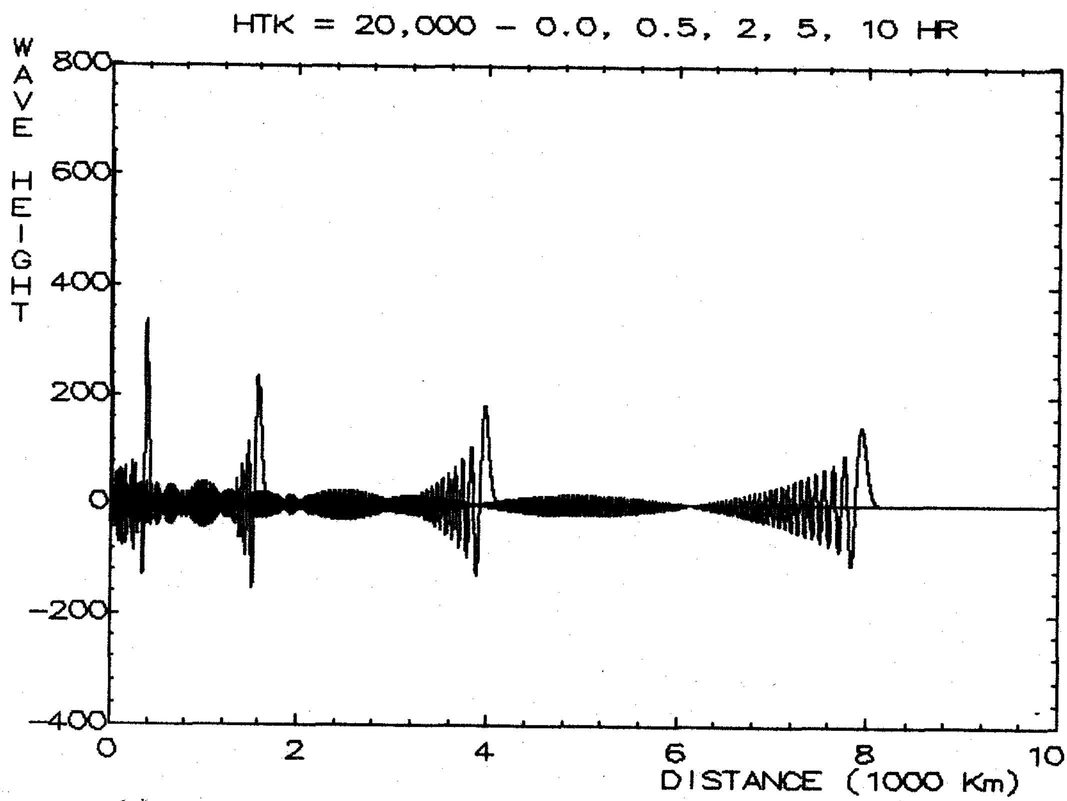
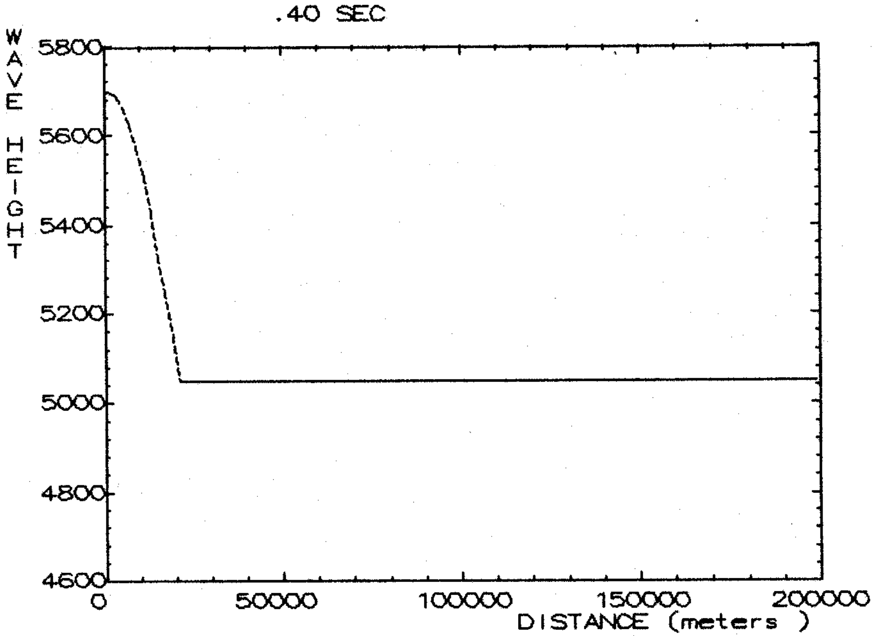


Figure 4. The wave profile in meters as a function of distance for a 650 meter high square displacement of water with a 20 kilometer half-width in 5000 meter deep water. The linear gravity model is used which models dispersion with no geometrical spreading.



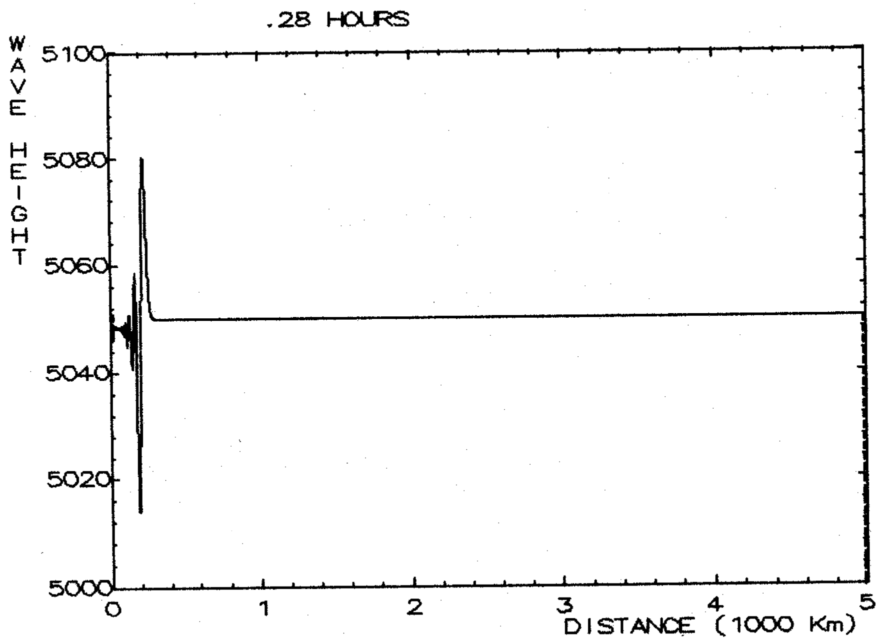
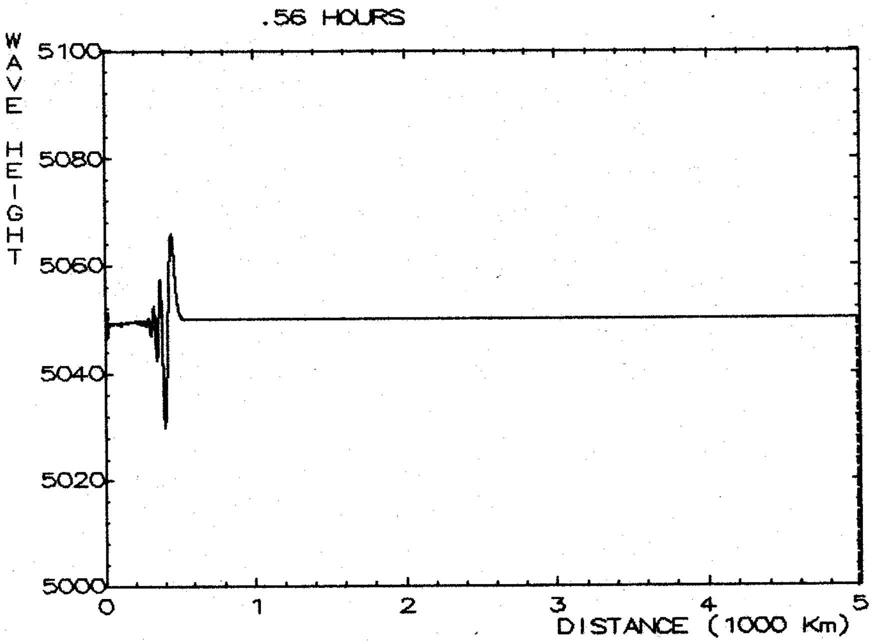
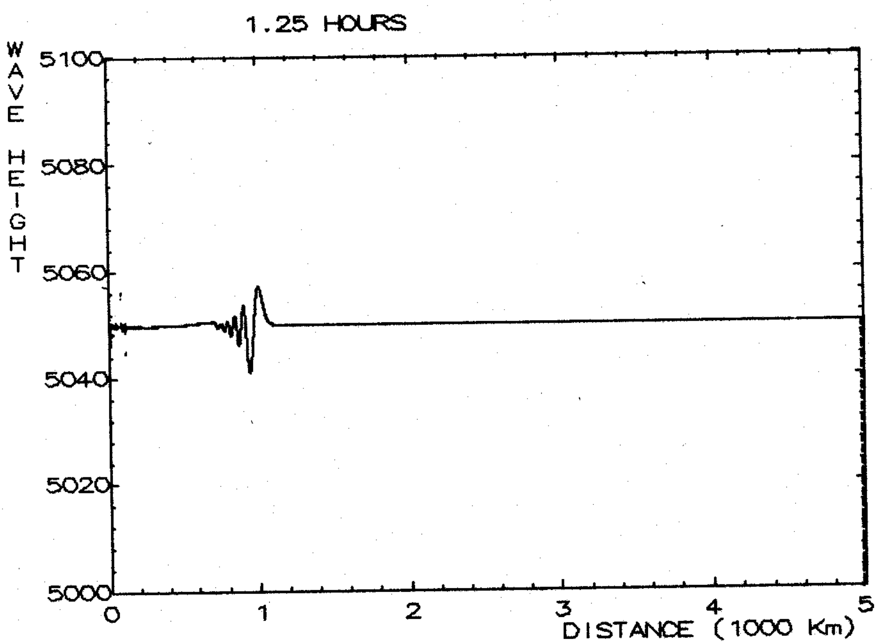
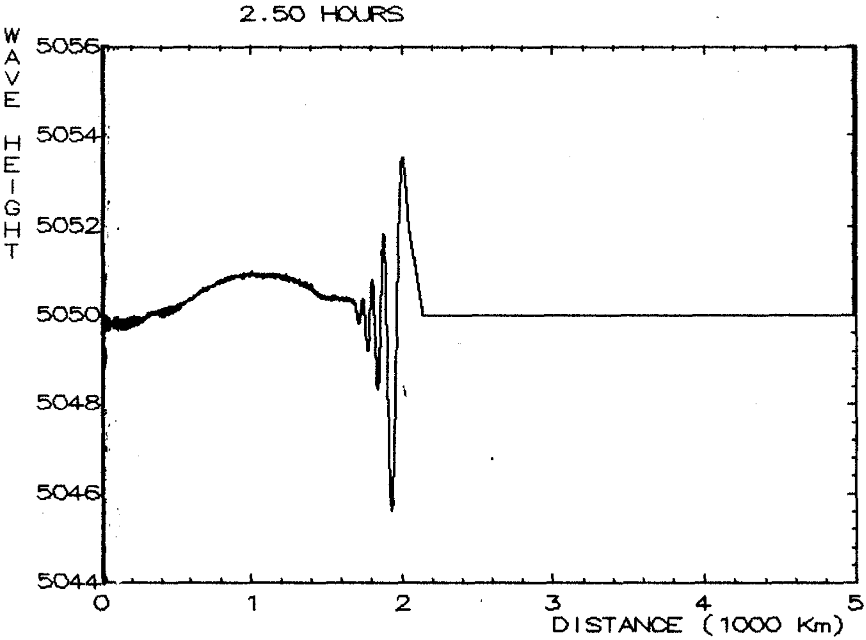
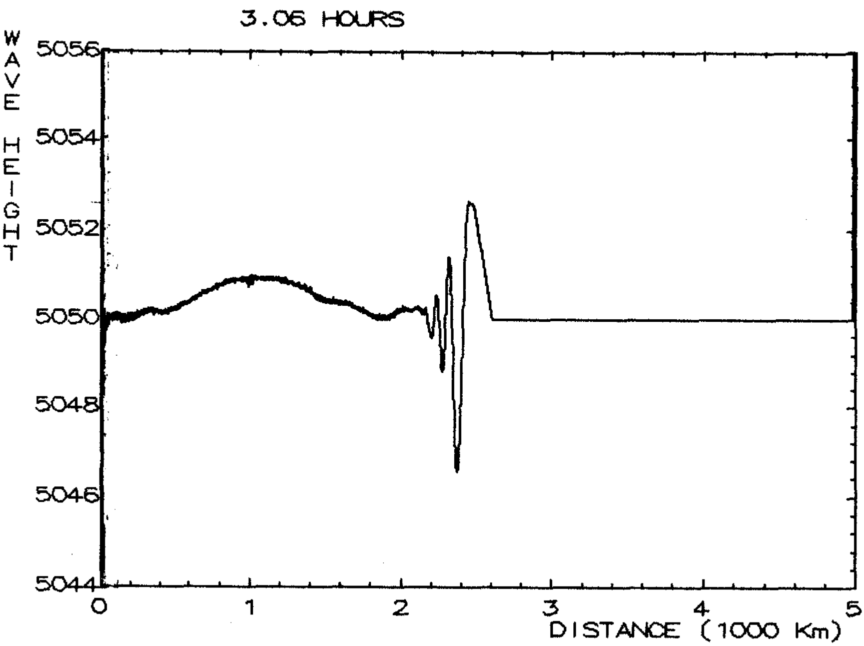


Figure 6. The wave profile in meters as a function of radius for the Fritz LaPalma displacement (first frame) in 5050 meter deep water for the full Navier Stokes Model.


.56 HOURS
WAVE HEIGHT
5100
5080
5060
5040
5020
5000
0
1
2
3
4
5
DISTANCE (1000 Km)


1.25 HOURS
WAVE HEIGHT
5100
5080
5060
5040
5020
5000
0
1
2
3
4
5
DISTANCE (1000 Km)


2.50 HOURS
WAVE HEIGHT
5056
5054
5052
5050
5048
5046
5044
0
1
2
3
4
5
DISTANCE (1000 Km)


3.06 HOURS
WAVE HEIGHT
5056
5054
5052
5050
5048
5046
5044
0
1
2
3
4
5
DISTANCE (1000 Km)

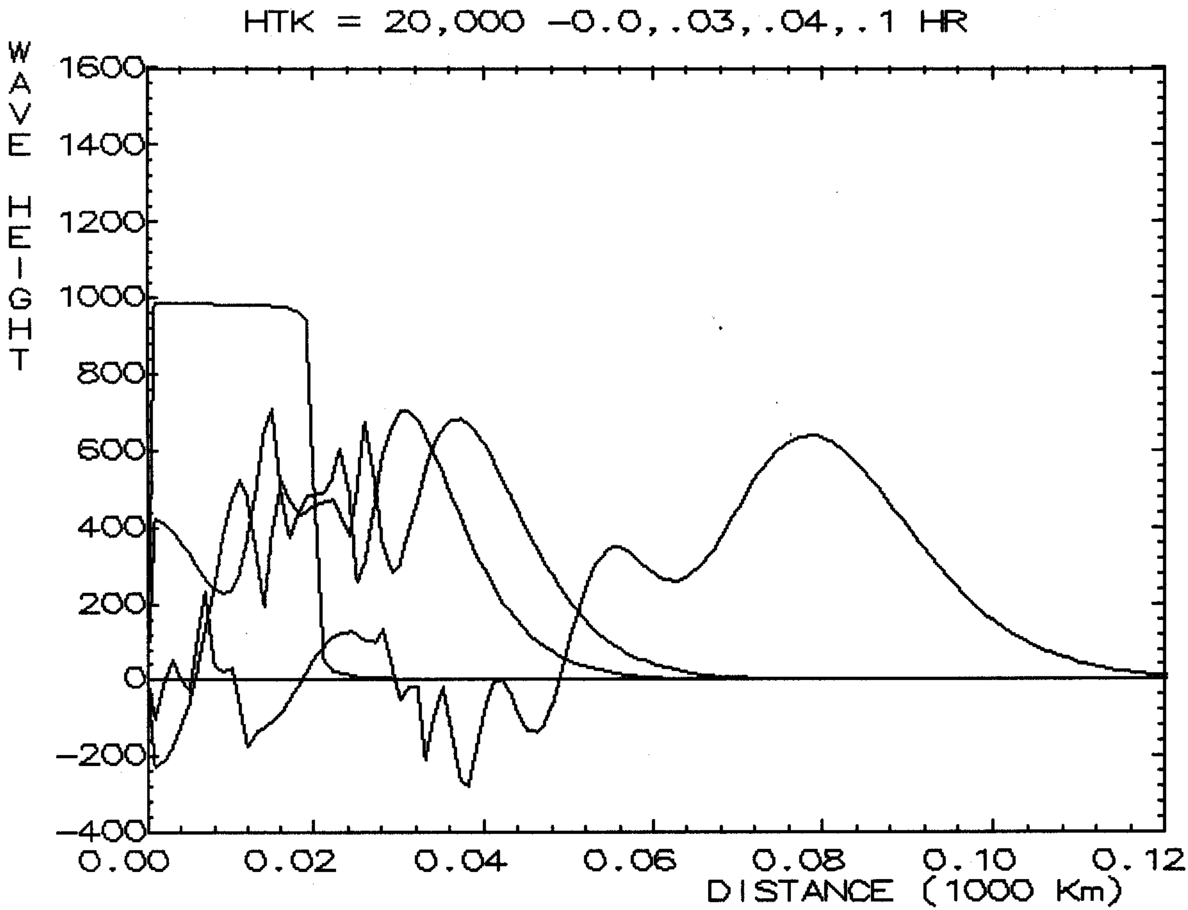


Figure 7. The wave profile in meters as a function of distance for a 990 meter high square displacement of water with a 20 kilometer half-width in 5000 meter deep water. The linear gravity model is used which models dispersion with no geometrical spreading. The wave approximates (in height and front profile) the Fritz wave shown in Figure 1 after 30 kilometers (0.038 hours) of travel.

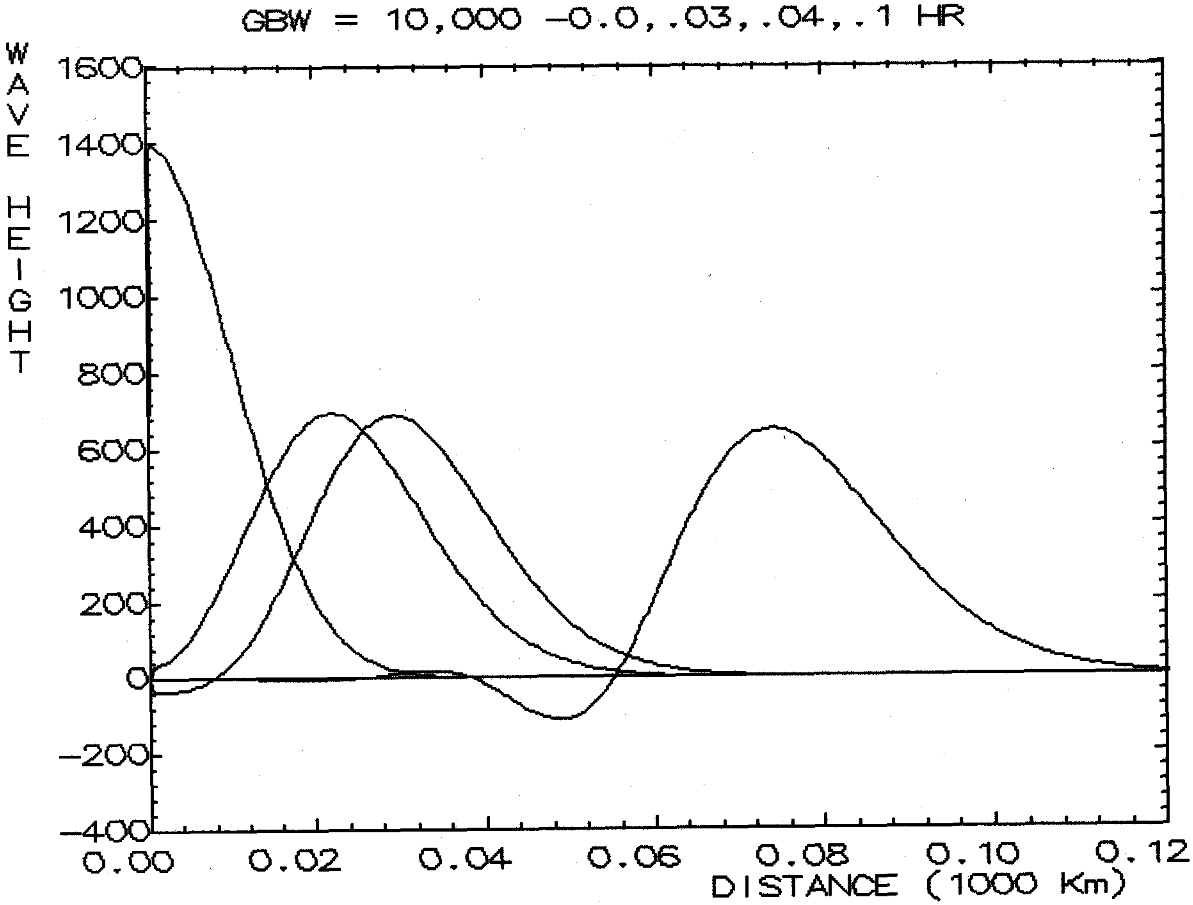


Figure 8. The wave profile in meters as a function of distance for a 1400 meter high Gaussian displacement of water with a break width of 10, in 5000 meter deep water. The linear gravity model is used which models dispersion with no geometrical spreading. The wave profile is similar to the Fritz wave shown in Figure I after 30 kilometers (0.038 hours) of travel.

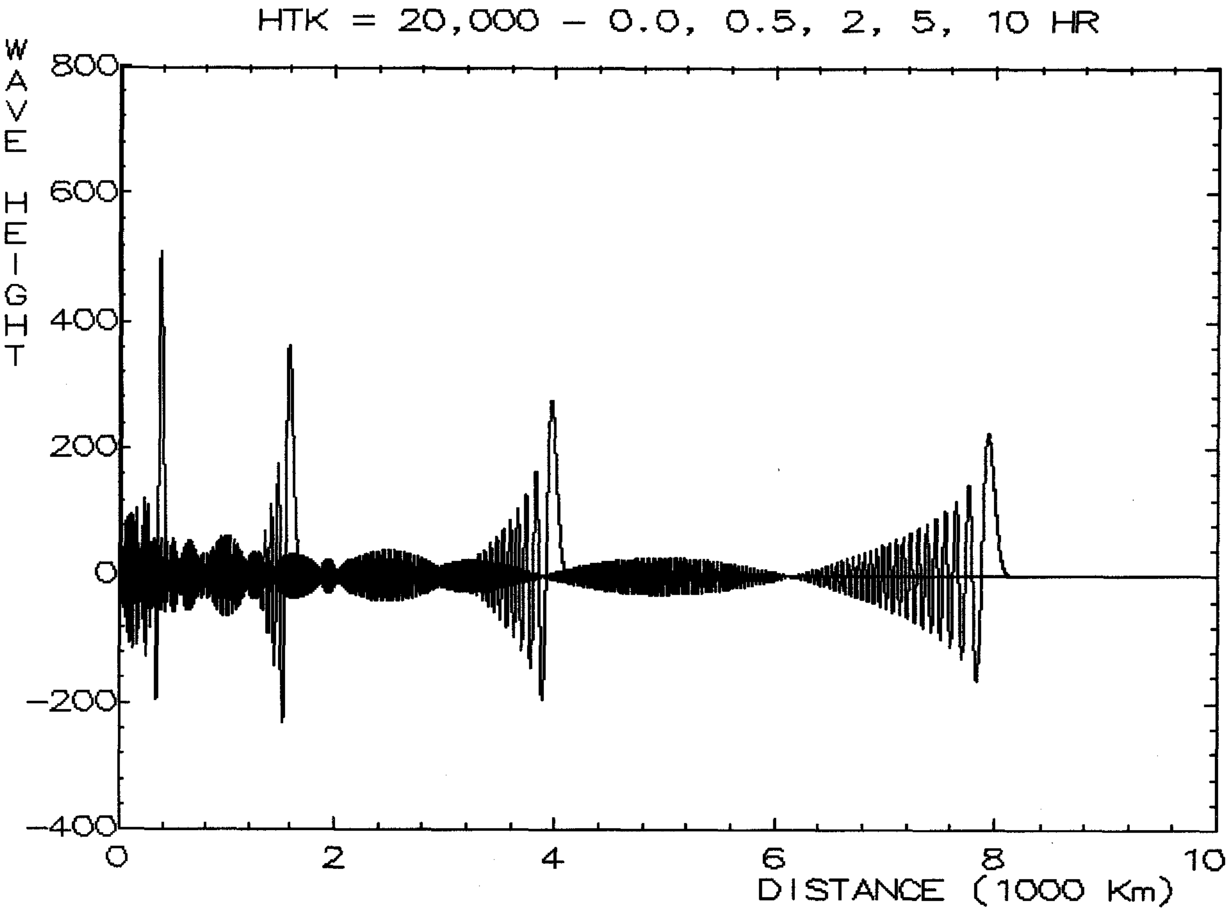


Figure 9. The wave profile in meters as a function of distance for a 990 meter high square displacement of water with a 20 kilometer half-width in 5000 meter deep water (Figure 7). The linear gravity model is used which models dispersion with no geometrical spreading. The wave amplitudes are about 1.4 times larger than those of Figure 4.

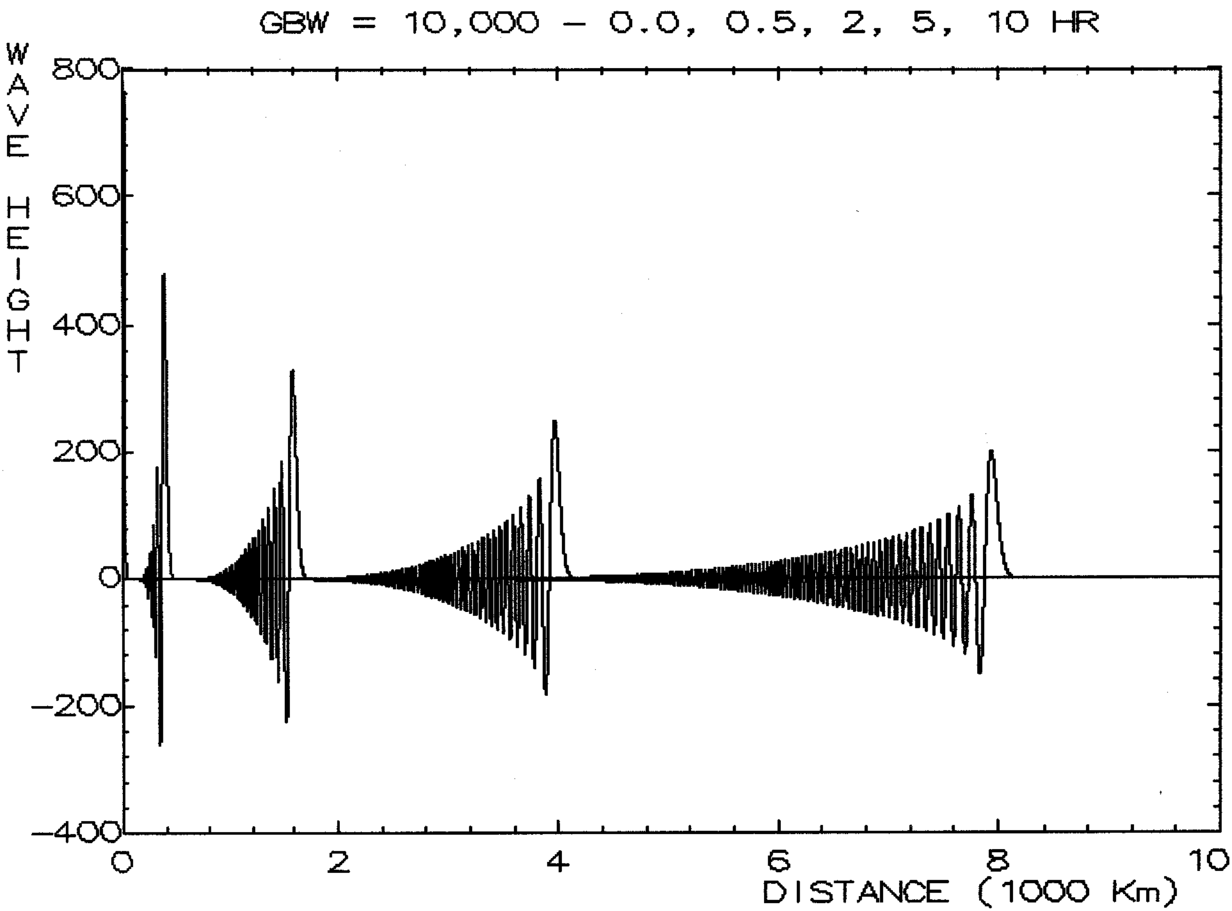


Figure 10. The wave profile in meters as a function of distance for a 1400 meter high Gaussian displacement of water with a Gaussian break length of 10 in 5000 meter deep water (Figure 8). The linear gravity model is used which models dispersion with no geometrical spreading. The wave amplitudes are about twice those of Figure 5.

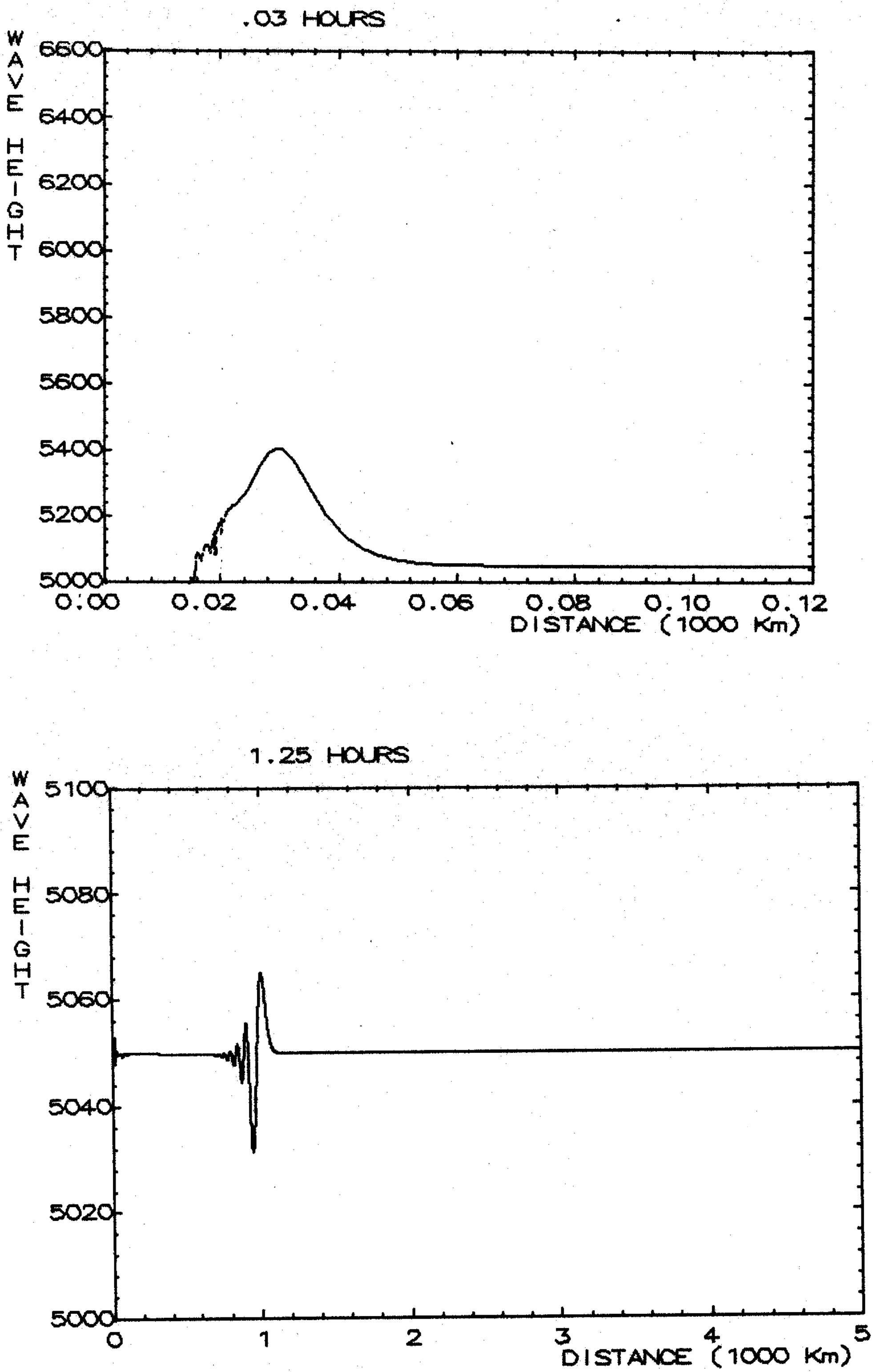
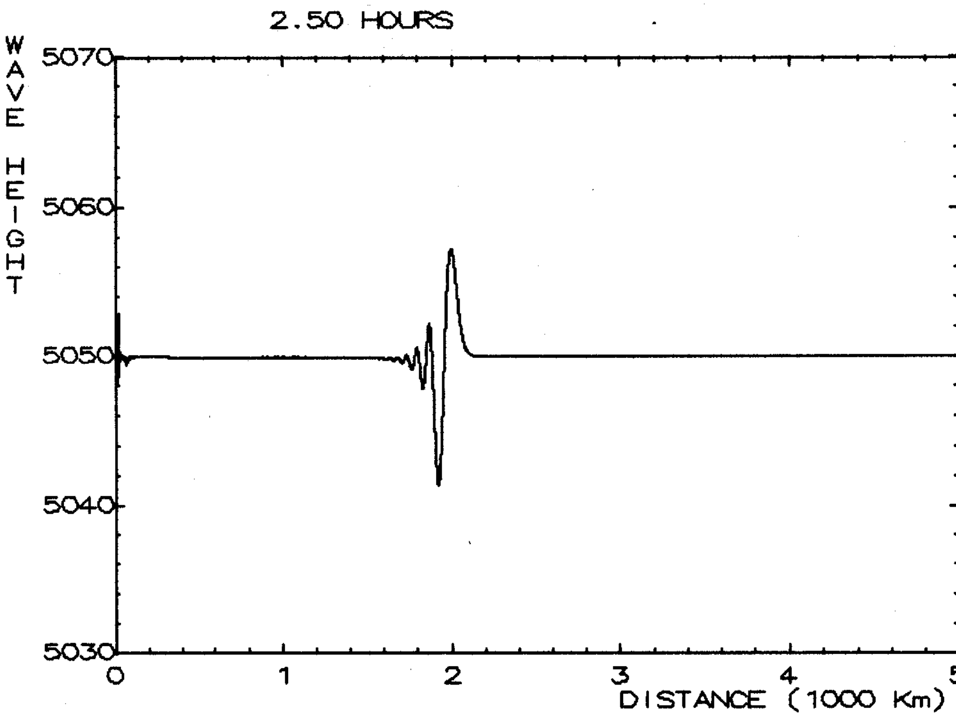
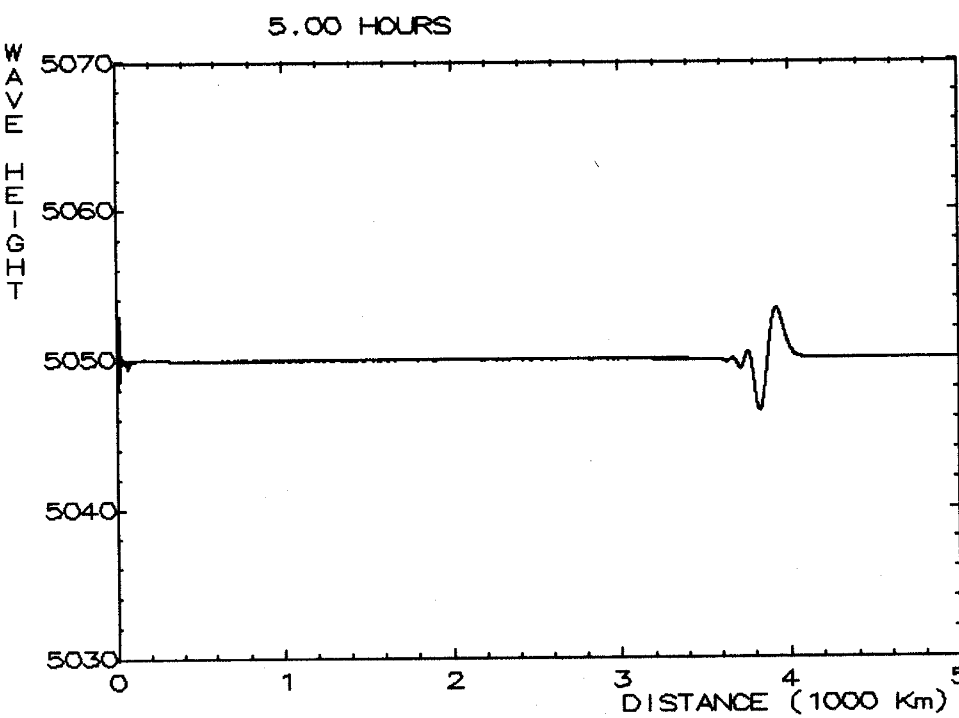


Figure 11. The wave profile in meters as a function of radius for the Fritz LaPalma wave at 30 kilometers (Figure 8) meter deep water for the full Navier Stokes Model. The wave amplitudes are about 2.15 times larger than those of Figure 6.


2.50 HOURS
WAVE HEIGHT
5070
5060
5050
5040
5030
0
1
2
3
4
5
DISTANCE (1000 Km)


5.00 HOURS
WAVE HEIGHT
5070
5060
5050
5040
5030
0
1
2
3
4
5
DISTANCE (1000 Km)

BOOK REVIEW

The Big One -
The Next California Earthquake

George Pararas-Carayannis

ISBN EB00014428

Forbes Press - - Honolulu, HI 96815

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Don't be turned off by the popularized style of the title or book. Pararas-Caryannis leads the reader - professional or layman - through a detailed background covering earthquakes, tsunamis, California geography and history, seismology, then finally summarizing factually the odds of major earthquakes in likely localities. And he does so in terms that all of us can understand and appreciate.

The author notes that he "will provide answers to reasonable questions by people concerned about their safety" and accomplishes that factually and in fact without sensationalism. Using a brief question and long answer format, this is actually a treatise on earthquakes and related natural hazards worldwide, with realistic scenarios of probable disasters in California.

There are several useful appendices and an extensive glossary (I learned a couple new terms) which are helpful and enlighting. I recommend this book.

Reviewed by George D. Curtis
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University of Hawaii, Hilo, Hawaii