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In: *Solutions to Coastal Disasters,*
Edited by L. Wallendorf and L. Ewing,
American Society of Civil Engineers, 2002.

SOLVING THE PUZZLE OF THE 1998 PAPUA NEW GUINEA TSUNAMI: THE CASE FOR A SLUMP

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Abstract: The 1998 Papua New Guinea tsunami disaster (over 2200 dead) cannot be simulated satisfactorily on the basis of the displacement field of its parent earthquake: In such simulations, runup amplitudes are too low, the extent of the devastated area too large, and the ratio of far- to near-field tsunami amplitudes also too large. Finally, the tsunami is computed to arrive at least 10 mn earlier than reconstructed from survivors' interviews. Rather, we show that evidence from underwater marine surveys, and from hydrophone records of hydroacoustic ("*T*") waves identify a slumping event as having taken place 13 minutes after the mainshock, at a site located 35 km offshore, and having involved a volume of 4 km^3 of material. Simulations based on a such a source reproduce all main characteristics of the local runup distribution.

INTRODUCTION

The tsunami of 17 July 1998 in Papua New Guinea [PNG], which killed over 2200 people (Kawata *et al.* 1999), and eradicated villages along a 30-km stretch of coastline (Figure 1), remains one of the worst disasters of its kind in the 20th century: only the 1933 Sanriku, Japan tsunami resulted in a larger loss of life. Yet, the PNG tsunami followed a relatively moderate earthquake, correctly identified in the hours following the disaster as being simply too small to be a likely generator of the tsunami. During the field survey by the International Post-Tsunami Survey Team, we proposed as early as 04 August 1998, that the tsunami may have been generated by an underwater landslide or slump (McGuiness 2001). Over the next

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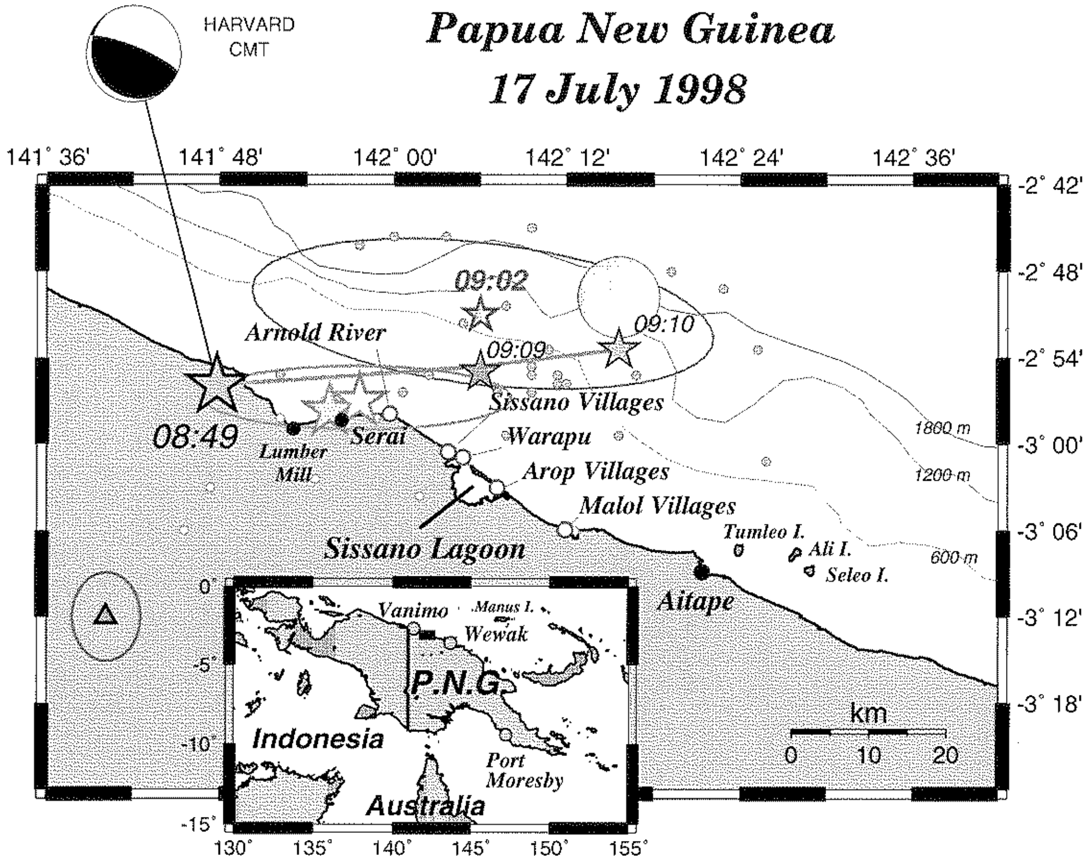

Fig. 1. Map of the coast of PNG affected by the tsunami. The big stars are various locations of the epicenter of the mainshock., the smaller stars early aftershocks, and the small dots in the background later aftershocks. The open circles are villages destroyed by the wave, the solid ones unaffected localities. The open star is the location of the 09:02 event with error ellipse. The large disk is the location of the cavity containing the slump.

few months and years, a growing body of evidence, obtained both in the field in through theoretical simulations, has brought overwhelming support to this concept, and considerably sensitized the coastal engineering community to the risk posed by the generation of local tsunamis by underwater slumps in both active and passive margins (Driscoll *et al.* 2000, Borrero *et al.* 2001). This paper reviews the saga of the PNG tsunami, from the standpoint of building the case for attributing its source to an underwater slump approximately 4 km^3 in volume, occurring 13 minutes after the main seismic event.

The idea that tsunamis may be generated by submarine landslides is not new, being found in the works of such visionaries as Milne (1898) and Montessus de Ballore (1907). Using the disparity between the geographic extent of the macroseismic and tsunami damage following the 1922 Chilean earthquake, Gutenberg (1939) went as far as to suggest that all tsunamis resulted from major underwater avalanches, taking place in the wake of large seismic events. During the following decades, the development of modern seismic source theory, concurrent with the occurrence of a few truly gigantic earthquakes (Chile, 1960; Alaska, 1964), allowed the successful modeling of their catastrophic far-field tsunamis (200 people were killed in Japan in 1960) using as a source the static dislocation field of the earthquake on the ocean floor. It was however noted that tsunami runup in the source area could occasionally present large local fluctuations, and such enhancements of its amplitude were attributed to small-scale underwater landslides (Plafker *et al.* 1969). Similar observations were made, for example at the site of the 1992 Flores, Indonesia tsunami, during the post-seismic inundation survey (Tsuji *et al.*, 1995, Plafker 1997). The latter was part of an effort, which has now become systematic, to precisely map the penetration of the waves in the immediate aftermath of major tsunamis. This program started with the 1992 Nicaraguan tsunami (Satake *et al.* 1993), and has now covered 12 tsunamis, most recently the Peruvian event of 23 June 2001 (Okal *et al.* 2001*a*). Slumps triggered by strong motion in the epicentral area were also proposed to explain the generation of highly localized systems of waves following onland earthquakes, such as the 1989 Loma Prieta event (Ma *et al.* 1991).

In the meantime, detailed seismological investigations of two intriguing earthquakes, both of them locally tsunamigenic, led H. Kanamori to propose that they must have involved significant underwater slumping. The tsunami from the 1975 Kalapana event (Eissler and Kanamori 1987) had inundated the deserted Southern coast of the Big Island of Hawaii, killing two campers, and the Grand Banks earthquake of 1929 (Hasegawa and Kanamori 1987) whose tsunami killed 29 people along the southern shores of Newfoundland (the heaviest death toll from an earthquake within the present borders of Canada), had long been known to have generated large turbidity currents, which had ruptured a series of transatlantic telegraphic cables (Doxsee 1948). These exceptions notwithstanding, the source of most recent tsunamis have been successfully modeled using the static dislocation from their parent earthquake, even in the case of the so-called "tsunami earthquakes", which feature a slower than usual velocity of propagation of the seismic rupture along the fault plane, and hence an apparent disparity between the amplitudes of their tsunamis and of their conventional seismic waves (Kanamori 1972, Polet and Kanamori 2000).

In this general context, the exceptional nature of the PNG tsunami became evident as early as 18 July 1998 when the magnitude of its disaster started to reach the outside world. Reports of 10–m high waves and the growing death toll clearly made it one of the largest tsunamis in recent decades, while amplitudes at far-field tidal gauges did not exceed 20 cm, and ongoing seismic analyses confirmed the moderate size of the parent earthquake: Magnitudes were measured at $m_b = 5.9$; $M_s = 7.0$; $M_m = 6.8$, with a preliminary moment computed at 5.2×10^{26} dyn-cm by the Harvard group (later revised downwards to 3.6×10^{26} dyn-cm), and most remarkably, the energy-to-moment ratio introduced by Newman and Okal (1998) confirmed (with $\Theta = \log_{10} E/M_0 = -5.50$) that the earthquake could not be regarded as simply one more example of an anoumously slow tsunami earthquake, all such events featuring $\Theta < -6$.

Thus, and even before the on-site tsunami survey, it was known that the earthquake dislocation source could not explain the principal features of the local runup, and therefore, an additional source of a different nature was strongly suspected to be the source of the tsunami. This was confirmed by early simulation efforts (V.V. Titov and F. Gonzalez, pers. comm., 1998), which failed to match the reported wave heights of 10 m or more from any combination of dislocation parameters compatible with the preliminary centroid moment tensor solution of the earthquake.

The on-site survey by the international team confirmed the early reports regarding the amplitude of the runup, found to be consistently in the 8–12 m range, with a maximum value of 15 m measured at one point in the vicinity of Arop. A crucial observation consisted of verifying the very limited geographical extent of the devastated area: the runup falls off quickly to the northwest and southeast of Sissano, with amplitudes as small as 2 m at Serai, a village located only 10 km from the zone of maximal destruction.

A further, key, observation was obtained from hundreds of eyewitness testimonies, compiled by Davies (1998) over the following weeks and months, namely that the tsunami washed the shoreline no sooner than 19 minutes after the mainshock. The precise timing of the arrival of the waves was made possible by the occurrence of a doublet of strong felt aftershocks taking place at 09:09 and 09:10 GMT, 19.5 to 20 minutes respectively after the mainshock (which took place at 08:49 GMT) and providing a natural *relative* clock, which allowed the precise reconstruction of the timing of the inundation, even in the absence of absolute timing, given that the witnesses could not be expected to have checked their watches (most did not wear one in the first place) while running for their lives.

THE FAILURE OF THE DISLOCATION MODEL

In simple terms, the dislocation model, which assumes that the tsunami is generated by the displacement of the water due to the static deformation of the ocean floor resulting from the earthquake, encounters formidable difficulties when attempting to model the existing datasets. Specifically, such a standard approach cannot account for the following observations:

• *The local runup amplitudes are much too large, relative to the earthquake's size.* While tsunami runup has often been observed to reach 10 m or more in the local field, the earthquakes involved had always been much larger. A typical example would be the 1993 Hokkaido-Nansei-Oki earthquake, whose tsunami devastated the island of Okushiri, Japan, but whose moment was 13 times larger than that of the PNG event. While the geometry of faulting can affect the generation and amplitude of the local tsunami, it quickly became apparent during early simulation attempts that the runup on the coast could not be modeled satisfactorily with any combination of fault parameters which would remain compatible with what has to be regarded as a moderate seismic source.

• *The large local runup amplitudes are concentrated on a very limited section of the coastline.* A remarkable aspect of our survey was the rapid fall-off of the runup as we moved away from Sissano lagoon along the coast line. The village of Serai, a mere 15 km from the center of the devastation, recorded only a 2-m wave and suffered no damage. On Figure 2, we quantify this observation by plotting the dataset of measured runup amplitudes as a function of distance along the shore, and fitting it with a function of the type

$$y = \frac{b}{\sqrt{\left[\frac{x-c}{a}\right]^2 + 1}} \tag{1}$$

where the parameters a, b, c are optimized by trial and error to minimize the misfit between the dataset and the model. While Equation (1) is purely empirical and may not be justified analytically, the parameter b clearly expresses the maximum runup at the center of the distribution, while a characterizes the width of the distribution; the dimensionless ratio b/a describes the aspect ratio of the distribution of runup along the coast. Figure 2 further compares the PNG runup distribution with a similar dataset obtained during the 1992 Nicaragua field surveys. While both datasets had been previously published (Abe *et al.* 1993, Batista *et al.* 1993, Kawata *et al.* 1999), we reproduce them here *on the same scale.* Despite the large scatter inherent in this kind of dataset, the difference in aspect ratio, reaching a factor of 20, is striking.

The larger aspect ratio of the PNG dataset cannot be reconciled with a seismic dislocation. In very simple terms, the parameter b is controled by the amplitude of seismic slip u on the fault, while the parameter a is largely controled by the dimension of the fault, L. Thus, the aspect ratio b/a is directly related to u/L, which is a measure of the strain release during the earthquake. Notwithstanding the possible influence of parameters such as fault orientation and epicentral distance on the individual values of a and b, we have found empirically from a large number of simulation attempts that an adequate match of the aspect ratio b/a observed in PNG would require increasing the strain u/L released on the fault by *ca.* one order of magnitude, over and above the maximum value (2×10^{-4}) observed in shallow earthquakes, and widely interpreted as expressing the maximum strength of the rupturing medium. In other words, the only way to account for the observed b/a using a dislocation source would be to have the earthquake take place in a material other than crustal rocks, an unrealistic assumption in the first place, and one certainly not borne out by any of the seismological evidence (Figure 3).

• *There is a strong discrepancy between the runup observed in the far- and near-fields.* The tsunami was not observed beyond Japan, where it was registered on tidal gauges with amplitudes of ≈ 10 cm, occasionally reaching 25 cm upon harbor resonance. As discussed later by Satake and Tanioka (2000), simulations using transpacific codes generally result in successful modeling of these observations based on the published models of the earthquake dislocation (Kikuchi *et al.* 1998). In order to resolve this apparent paradox in view of the modeling difficulties in the near-field, a number of authors, including Satake and Tanioka (2000) and Geist (2000), have proposed to invoke a variety of models, including the use of the conjugate fault plane, or of a splay fault. As detailed for example in Okal and Synolakis (2001), and regardless of other problems raised by such models, these

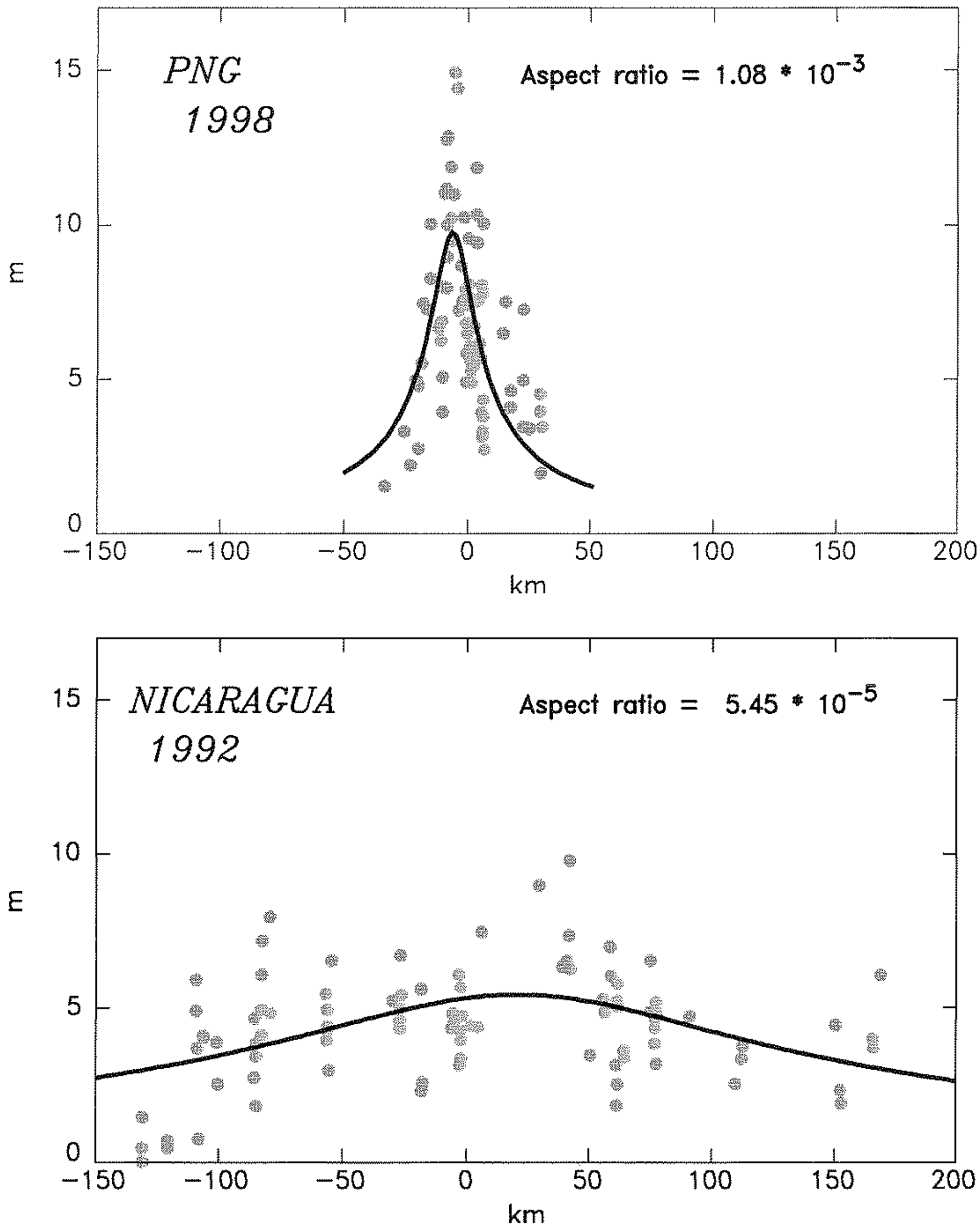

Fig. 2. Top: Distribution of runup heights (individual dots) along the PNG coast, after Kawata *et al.* (2000). The thick trace is the best-fitting curve of the form of Equation (1). Bottom: Same as top for the 1992 Nicaraguan earthquake, after Abe *et al.*. (1993) and Batista *et al.* (1993). Note the strikingly different aspect ratios of the two distributions. The two frames are plotted using the same scales.

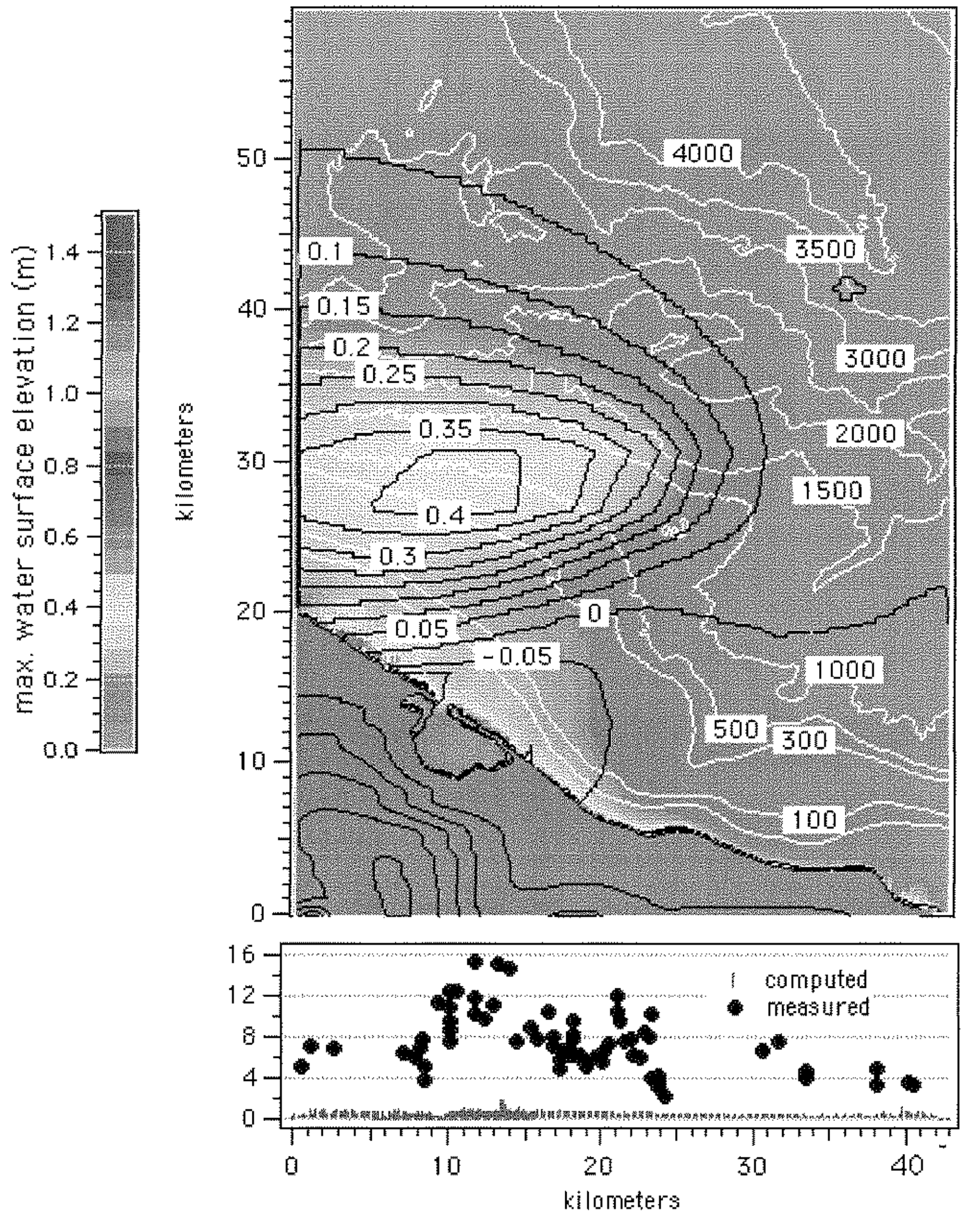

Fig. 3. Attempt at simulating the PNG runup data, using a seismic dislocation model as the source of the tsunami. Top: Contours (black) of the displacement field of the earthquake's seismic dislocation (in meters), superimposed on the bathymetry (white) contours (in meters). The shading is keyed to the maximum amplitude reached by the wave at any given point. Bottom: the dots are the measured runup values (same dataset as on Figure 2). The small bars at the bottom of the diagram are the results of the simulation, which are a factor of 5 to 10 smaller than observed.

efforts remain unsatisfactory, largely because the relative far- and near-fields of a tsunami generated by a dislocation are once again controled by the scaling laws of the seismic source, themselves deeply rooted in the material properties of the rocks involved, and in the mechanism of the dislocation process.

• *Finally, we recall here the timing discrepancy* revealed by Davies' (1998) interviews of eyewitnesses. As discussed in detail by Imamura and Hashi (2001), the tsunami is documented to have arrived ashore at least 10 minutes too late to have been generated by the main shock, under all scenarios compatible with reliable epicentral information. The only alternative would be to relocate the earthquake source far out at sea (at least 70 km), which would violate at least three epicentral locations (obtained indepedently by the U.S. Geological Survey, by Engdahl *et al.*'s (1998) method, and by Wysession *et al.*'s (1991) algorithm), as well as the distribution of aftershocks.

THE EVIDENCE FOR THE SLUMP

Evidence for a major slump off the coast of PNG was gathered starting in the Fall of 1998 using a variety of techniques. First, a number of marine surveys mapped systematically the offshore of the devastasted area, reaching several major conclusions, as described in detail by Tappin *et al.* (2001) and Sweet and Silver (2001). No evidence could be found for a fresh dislocation of an amplitude and spatial extent sufficient to be a candidate for the tsunami source. However, the most seminal discovery during the marine surveys was that of a large amphitheater located *ca.* 35 km from the shore, containing the remnants of a fresh slump, estimated by Sweet and Silver (2001) to have a volume of 4 km^3. Of course, such surveys remain incapable of providing a timing for the slumping.

In parallel with the marine surveys, and based on the timing discrepancies, a systematic effort was launched to identify, in the geophysical database, the signature of any event posterior to the mainshock which could have been the source of the tsunami. We were able to identify, on records obtained at hydrophone station WK31, located near Wake Island, the signature of T waves generated by a small event originating at 09:02 GMT, *i.e.*, 13 minutes after the mainshock. (T waves are acoustic waves propagating with extreme efficiency in a low-velocity wave guide centered at 1200 m depth in the world's oceans; they have been used for purposes as varied as the tracking of submarines, the detection of unsuspected volcanic activity on the seafloor, and the monitoring of colonies of cetaceans.) The 09:02 event was detected by its conventional seismic waves at a handful of seismic stations, described by the USGS as a small earthquake, assigned a magnitude $m_b = 4.4$, and we verified that the ellipse of uncertainty for its epicenter comprises the amphitheater (Figure 1), so that it is legitimate to assume that the 09:02 event did indeed take place in the cavity.

As we discussed in detail in Okal (2001), in addition to having a spectrum rich in high frequencies, the 09:02 event features an exceptional duration of its T wavetrain, which reaches 45 s (Figure 4), a figure more in line with the duration of the mainshock's T waves (55 s), than comparable to those (15 s) of aftershocks with similar body-wave magnitudes and located at comparable epicenters. In other words, the duration of the T wave, and hence of the source process of the 09:02

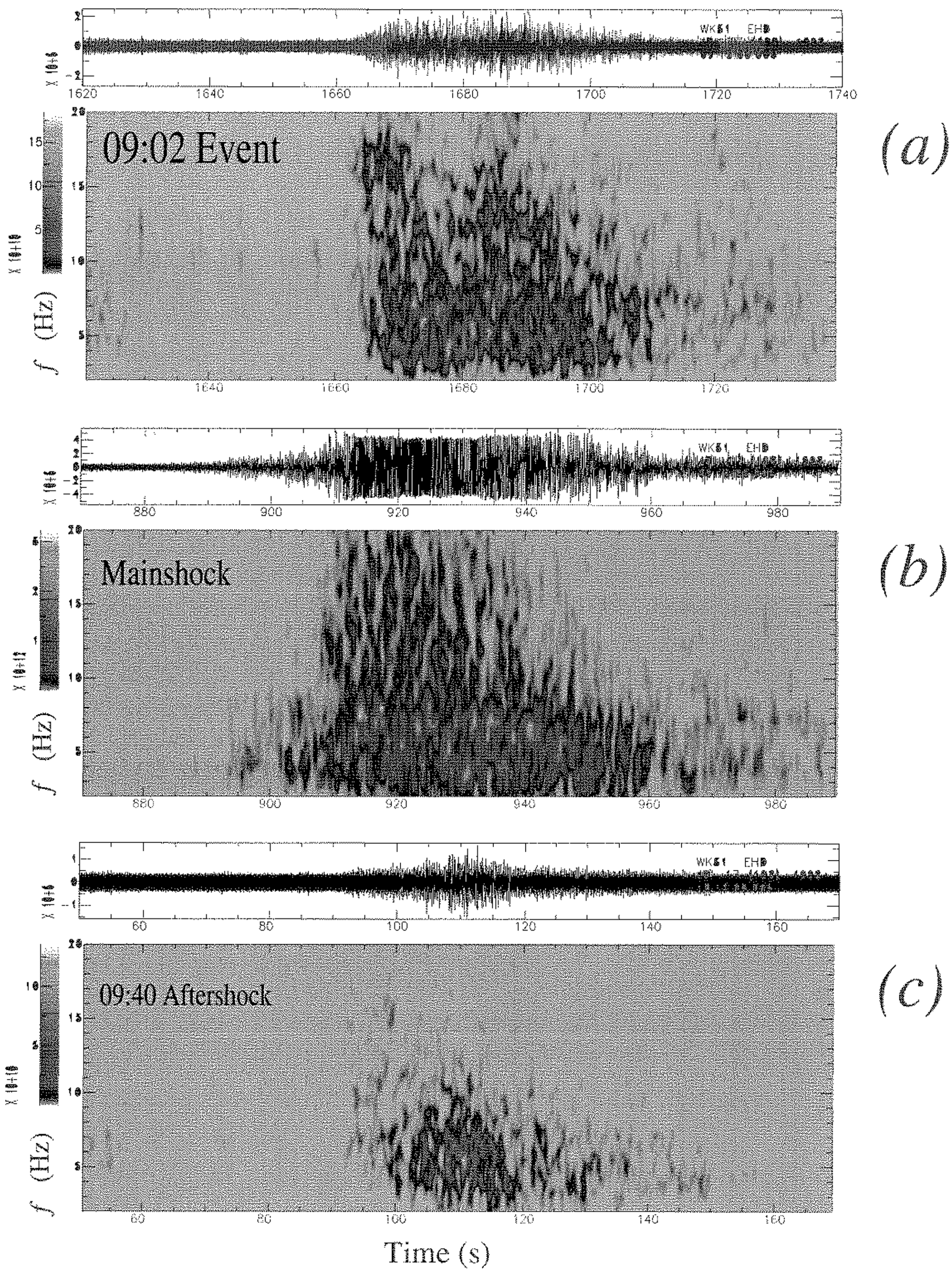

Fig. 4. Spectrograms of *T* waves recorded at hydrophone station WK31. Each frame represents a contour plot of the energy present in the Fourier spectrum of the record, as a function of time (abscissa) and frequency (ordinate). The top record is that of the 09:02 slump event; note that its duration is comparable to that of the mainshock (center), and much longer than that of a genuine aftershock of comparable magnitude (bottom).

event, is much longer than typical for an earthquake of its magnitude, *i.e.*, with comparable conventional seismic waves. This property, verified by Okal (2001) on records of the high-frequency regional phase P_o, expresses that the 09:02 event violates the *scaling laws* of seismic source theory which predict the concurrent growth of source duration and seismic wave excitation with earthquake size. Thus the 09:02 event cannot be a regular earthquake dislocation, and we interpret it as a massive slump, triggered in the walls of the amphitheater, 13 minutes after the mainshock, lasting approximately 45 s, and generating the tsunami.

THE CASE FOR THE SLUMP

We argue here that our interpretation of the 09:02 source as a slumping event generating the tsunami resolves most of the difficulties encountered when attempting to model the PNG tsunami with a dislocation source.

First, the slump is present and well documented by the seismic reflection experiment conducted on *R/V Maurice Ewing* (Sweet and Silver, 2001). This is in contrast with various models requiring a long, coherent thrust fault breaking the seafloor for which the marine surveys found no evidence (the so-called 40–km fault being fresh only in its westernmost part, and featuring a normal polarity (Tappin *et al.* 2001) rather than the reverse one required by the CMT solution). While the marine surveys cannot date the occurrence of the slumping, the T–wave evidence suggests that an intriguingly long event which cannot have the nature of a seismic dislocation took place on 18 July 1998 at 09:02 GMT, essentially at the mapped location of the slump.

Second, the 09:02 event is well timed as the source of the tsunami, since it results in inundation of the coast in the devastated area at times ranging from 09:10 to 09:12 (from slightly before to a few minutes after the occurrence of the aftershock doublet); the occasional slight discrepancies (on the order of 2–3 minutes) with the timing compiled by Davies (1998) can be ascribed to the propagation of the wave in the very shallow waters for which no accurate bathymetric data are available.

Third, and most importantly, the slump is an acceptable source for realistic simulations of the distribution of runup on the coast. As shown on Figure 5, our simulation uses a source comprising a negative trough 18-m deep in the sea surface, combined with a positive bulge 16-m high, displaced 5 km in an essentially Northern azimuth. This source is positioned over the location of the slump, as mapped by Sweet and Silver (2001), and is used as an initial condition for the MOST tsunami propagation and runup code of Titov and Synolakis (1998), which solves the nonlinear shallow water wave equations as a system of hyperbolic differential equations. The simulation of the penetration of the wave overland is handled by following the evolution of the moving shoreline at the last wet grid point seaward of the initial coastline. The performance of MOST was assessed on the case studies of the dislocation-induced 1993 Hokkaido-Nansei-Oki tsunami (Titov and Synolakis 1997), of the landslide-generated 1994 Skagway tsunami (Plafker *et al.* 2000), and on laboratory data obtained by Synolakis (1987), Briggs *et al.* (1995) and Kanoğlu and Synolakis (1998). The results on Figure 5 show that our simulation can successfully model the general amplitude of the runup along

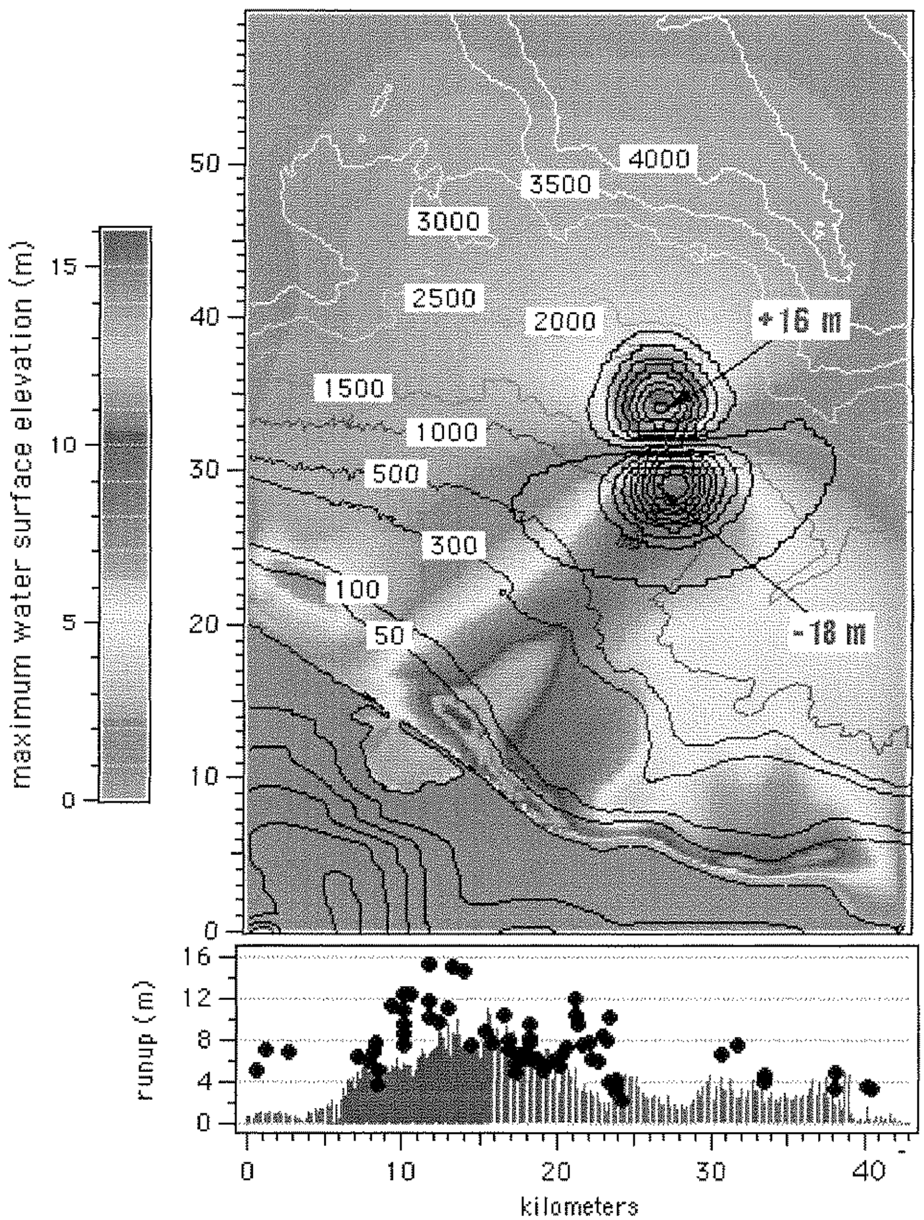

Fig. 5. Same as Fig. 3, using the slump as a source of the simulation. The black contours denote the initial deformation of the surface of the sea, ranging from a low of 18 m at the heel of the slump, to a high of 16 m at its toe. Note the wall of water approaching the coast, but constrained to a lateral extent of about 15 km. On the bottom frame, note that the simulated amplitudes provide a much better fit to the measured data, and to its lateral fall-off outside the devastated area.

the devastated coastline. The modeled runup amplitudes are consistently in the range 8–10 m; while this is slightly below the extreme values of 12 to 15 m measured at a few individual points along the coast, the difference may represent the interaction of the wave with the extreme shallowest portions of the ocean, where detailed local bathymetry is not available.

Most importantly, our results also show a rapid lateral fall-off of the simulated amplitudes, as expressed in the map view on Figure 5 which shows the onslaught of a wall of water, concentrated on a 15-km stretch of the coastline. Thus, our simulation reproduces the aspect ratio of the runup distribution.

In order to understand the success of the slumping model as a source of the PNG tsunami, we make the following observations:

• First, the slump constitutes an additional physical source, separated both in time (13 minutes) and obviously in space (it takes place on the *surface* of the solid Earth rather than inside its *body*), from the dislocation source of the mainshock. Thus, any comparison between the tsunami amplitudes in the near field and the seismic waves of the mainshock becomes irrelevant, since they were generated by different sources.

• Second, regarding the aspect ratio b/a of the distribution of runup along the coast, we note that, in the slump model, b will be controled mostly by the initial amplitude ζ of the deformation of the sea surface, and in turn by the vertical deformation ε of the seafloor created during the slumping event, while a while be related to the horizontal extent Y of the slumped body. It is clear that ε can easily reach dimensions much larger than the slip u of the mainshock (we use here $\zeta \approx 17$ m), while on the contrary, Sweet and Silver (2001) have shown that Y does not exceed 4 to 5 km, as opposed to the 35–km fault length L inverted for the main shock by Kikuchi *et al.* (1998).

• Finally, regarding the ratio of runup amplitudes in the near and far fields, we note that an underwater slide or slump consists, in very schematic terms, of sucking down a large amount of water above the heel of the slide, and creating a more or less equivalent bulge at its toe. At the large wavelengths characteristic of tsunami propagation on the high seas, this source has essentially the properties of a *dipole*; by contrast, the dislocation source can be regarded as a monopole (barring exceptional geometries such as pure dip-slip on a vertical fault). Dipoles are known in all fields of physics to be high-frequency sources and their far fields to decay faster with distance than those of monopoles. In this respect, the generation of the PNG tsunami by a slumping event also explains the apparent discrepancy between the amplitudes recorded in the near field in PNG in the far field at transpacific distances.

CONCLUSION AND PERSEPCTIVE

In addition to its occurrence at least 10 minutes too late to be directly associated with its parent earthquake, the distribution of runup of the PNG tsunami violates the scaling laws governing seismic source theory. As such, no acceptable geometry of seismic source can successfully model the near-field runup distribution along the devastasted stretch of shoreline. Rather, we show that evidence from field surveys and from acoustic records support the interpretation of a reported small aftershock, 13 minutes after the main shock, as a large underwater slump taking place 35 km from the shore, and displacing an estimated 4 km^3 of material. Simulations using the slump as their source successfully model all principal characteristics of the near-field runup distribution.

The magnitude of the PNG disaster (over 2200 dead) serves as a chilly reminder of the hazard posed by underwater slumps. It immediately reopened the question of the exact generation of tsunamis following a few large earthquakes without an appropriate description of their tsunami source, among them the great 1946 Aleutian earthquake (Okal *et al.* 2001*b*), the 1925 Mexican earthquake (Singh *et al.* 1998), or even the 1922 Chilean event (Gutenberg 1939). In addition, and perhaps more ominously, it focused the interest of the tsunami community on the risk posed by submarine slumps triggered by small to moderate earthquakes, or even occurring in the absence of a detectable seismic event. For example, Borrero *et al.* (2001) used the model of a PNG-type slump to quantify seismic risk in the Santa Barbara Channel, where Greene *et al.* (2000) have documented scarred structures which could represent incipient failure. Finally, even in passive margins, slumps could take place as a result of destabilization of underwater land masses by mechanical, thermal or chemical agents involving unstable or metastable structures such as salt domes (McAdoo *et al.* 2000) or gas hydrates (Haq 2000).

ACKNOWLEDGMENTS

This research was supported by the Federal Emergency Management Agency. Field work in PNG was supported by the National Science Foundation.

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