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Comment on "Origin of the 17 July 1998 Papua New Guinea Tsunami: Earthquake or Landslide?" by E. L. Geist

Emile A. Okal
Northwestern University

Costas E. Synolakis
University of Southern California

In a recent contribution, Geist (2000) examines several characteristics of the 1998 Papua New Guinea (PNG) earthquake and tsunami and suggests that they can be explained by a standard dislocation in an elastic Earth without the need to invoke a landslide or slump as the mechanism of generation of the locally catastrophic tsunami.

We wish to emphasize that we agree with overwhelming evidence indicating that tsunamigenic events are never purely "tectonic" (*i.e.*, in the nature of an elastic dislocation) nor due to pure slumps. Most earthquakes trigger landslides; indeed aerial rockslides are well documented in the immediate vicinity of the USGS epicenter of the PNG earthquake west of the Serai lumber mill (Kawata *et al.*, 1999). Conversely, most landslides are presumably triggered by seismic events, however small and difficult to detect.

In several instances of recent tsunamis, small-scale underwater landslides or slumps have been documented or suggested to explain occurrences of very localized enhanced run-up amplitudes. This was the case at Riang-Kroko for the 1992 Flores event (Tsuji *et al.*, 1995), and on the western coast of Okushiri in 1993 (Shuto and Matsutomi, 1995). What truly sets apart the PNG tsunami (and challenges the scientific community) is the large extent of the coastal segment with very high run-up (25 km as opposed to a scale of hundreds to thousands of meters for the above examples). On the other hand, the aspect ratio of the run-up (sustained at 10 m, with peak values of 16 m) to the length of devastated coastline (25 km) is much too large to be explained in the framework of the scaling laws known to control elastic dislocations in the appropriate range of seismic moments. These simple, albeit qualitative, observations, all made before and during the posttsunami survey, rule out both a purely (or mostly) elastic dislocation and localized, small-scale underwater sliding or slumping as origins of the locally devastating waves. Rather, they suggest generation by a very large underwater slump.

In this general framework, we wish to discuss and occasionally rectify a number of assertions in Geist's paper, focusing first on the seismological aspects of the issue.

SEISMOLOGICAL ASPECTS

Seismic Moment

It is worth stressing that the final Harvard solution for the main shock has a moment of 3.7×10^{26} dyne-cm (3.7×10^{19} N-m), only 70% of the "QUICK" solution obtained in quasireal time. This makes it, if anything, even more difficult to argue for a dislocation source.

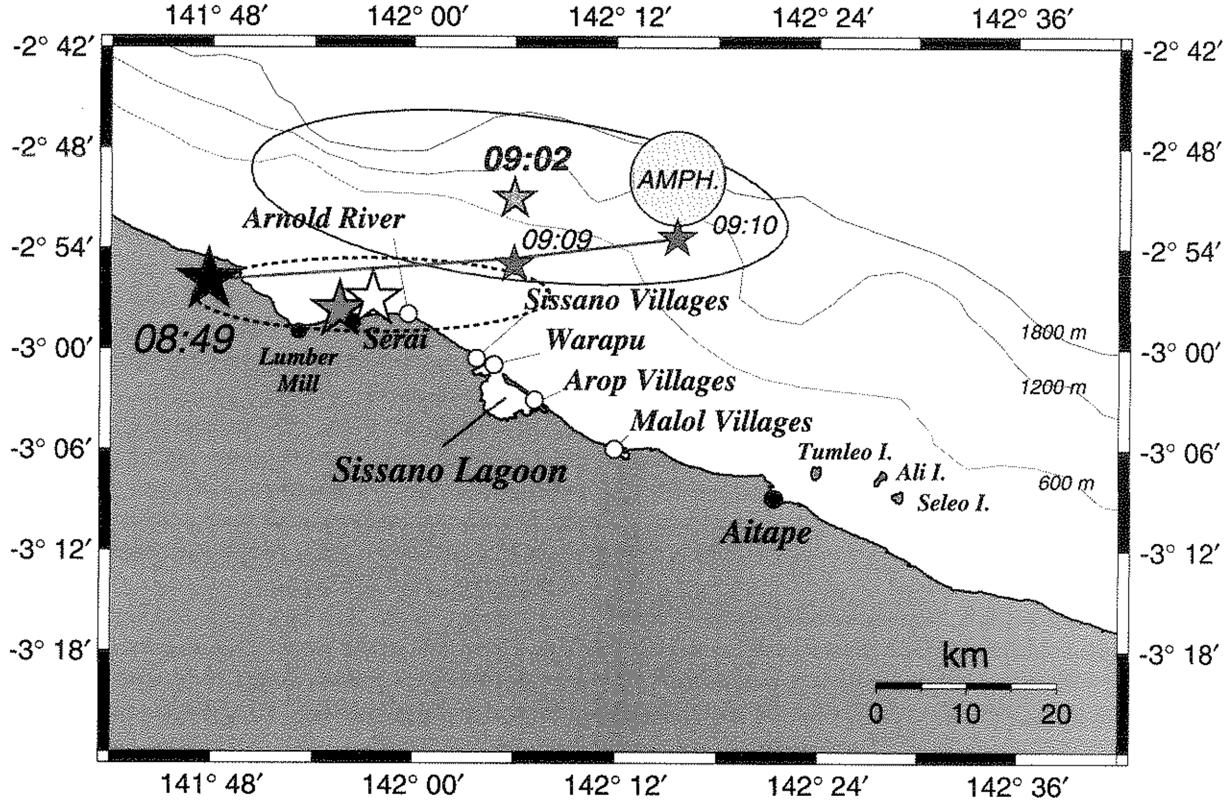
Epicentral Location of the Main Shock

Geist's (2000) Figure 3 is certainly incomplete in that it uses (and implicitly compares) a very preliminary USGS location, achieved on the basis of a partial data set of *P* arrivals, a preliminary epicenter computed by the regional ERI network, and a Harvard centroid. As with all solutions emanating from regional centers, the ERI epicenter may be strongly biased by station geometry, and we cannot regard it as significantly different from the final USGS solution. As for the Harvard centroid, we note from the full catalog entry (Dziewonski *et al.*, 1999) that it was obtained exclusively from mantle waves using 500 km wavelengths, thus casting doubt on the significance of any difference between epicenter and centroid.

The most comprehensive epicentral location available to date is the final USGS one, at 2.961°S 141.926°E. It is important to note that this epicenter is significantly south of the 40 km fault (see below). Our own relocation of the event, based on USGS earthquake phase data (Synolakis *et al.*, 2001), yields essentially the same epicenter, at 2.95°S 141.96°E. As shown in Figure 1, the error ellipse computed using Wysession *et al.*'s (1991) Monte Carlo algorithm (injecting gaussian noise with $\sigma_G = 1$ s) has a strong east-west trend and does not intersect the 40 km fault (see below). Furthermore, all our relocation efforts confirmed that the available data set has no depth resolution.

Aftershock Distribution

We believe that the aftershock studies, aiming at determining the fault plane of the main shock, lack depth resolution



▲ **Figure 1.** Map of the Sandaun coast of northwestern Papua New Guinea. The open circles on the coastline identify villages devastated by the tsunami; the solid circles are other, mostly spared, communities. The large stars are epicenters of the main shock (black: initial USGS; gray: final USGS; white: as relocated in this study, with error ellipse (dotted line)). The smaller, gray stars are epicenters of the relocated doublet at 09:09 and 09:10 GMT. The line joining them is the extent of the seismic rupture, as inferred from the seismological modeling of Kikuchi *et al.* (1998). The open star is the relocated epicenter of the 09:02 seismic event (with solid-line error ellipse). The superimposed disk labeled "AMPH." schematizes the location of the amphitheater where the slump was identified by the surveys (after Okal, 2000a).

and are therefore inconclusive. The study by Hurukawa (1999) uses CTBT data, which are overwhelmingly teleseismic in nature and offer no depth resolution (as documented by the very large error bars in his Figure 4).

Another study, by Tsuji (2000, and personal communication, 1998), used data from a temporary deployment of three seismometers along the Sandaun coast in the weeks following the event. Because of the inaccessibility of the hinterland, the deployed triangle had a very flat aspect ratio, offering no resolution along the polar angle of a cylindrical coordinate system whose axis is the longest dimension of the triangle; in the present geometry, this means that hypocentral depths trade off with distance across the shoreline, and the data set cannot help distinguish between the two fault planes.

Our own relocations of the forty-three larger aftershocks ($m_b > 4.4$) were based on USGS phase data, but all had to be carried at constrained depth. The resulting epicenters, broadly scattered over a 70 km by 40 km area, would favor the shallow-dipping fault plane (Synolakis *et al.*, 2001).

In conclusion, we regard the aftershock data set as having no depth resolution whatsoever and its generally very scattered pattern as agreeing better indeed with the shallow-dipping plane.

Choice of Fault Plane

Geist (2000) reports Kikuchi *et al.*'s (1999) source modeling of the main shock as giving "slightly better [variance reduction] for the steeply dipping plane, in comparison to the shallowly dipping plane". This statement is taken out of context, since Kikuchi *et al.* (1999) state that "the differences in variance reduction are so close that the teleseismic data alone cannot well judge which of the *P*-wave nodal planes was the actual fault plane" and elect to favor the steep plane on the basis of Hurukawa's (1999) relocations, which we believe cannot resolve the indeterminacy (see above).

Association with the 40 km Fault

There are at least four significant problems in asserting that the main shock rupture took place along the so-called 40 km fault mapped by Tappin *et al.* (1999). First, only the westward half of the fault has evidence of fresh motion, making the fault an unlikely candidate for the minimum 35 km of rupture determined by most directivity studies (Kikuchi *et al.*, 1998; 1999) and supported by dislocation scaling laws.

Second, fault motion along those fresh sections is mostly normal faulting (Tappin *et al.*, 1999), rather than the reverse thrust required by the focal mechanism of the event.

Third, the epicentral location, as determined by the final USGS solution (and our own relocations) is to the southwest of the 40 km fault and thus incompatible with this fault expressing the surface trace of the steeply (and north-northeasterly) dipping fault plane; under that geometry, the hypocenter, and hence the epicenter, of the earthquake would have to be north-northeast of the fault, not southwest of it.

Finally, the azimuth of the 40 km fault (approximately N90°E) agrees well with the general direction of rupture required by the directivity studies but is about 20° off that of the steeply dipping fault plane. Propagating a rupture in that azimuth while at the same time keeping the rupture on the steeply dipping fault requires deepening of the source during the rupture by as much as 10 to 20 km, which would then considerably reduce the local tsunamigenic potential of the earthquake.

Use of Far-field Tsunami Records

Geist (2000) mentions that the modeling of tidal gauge records of the tsunami at teleseismic distances favors a seismic dislocation and marginally so, the steep plane over the shallow one (Tanioka, 1999). This is hardly surprising, but also inconclusive. First, one should keep in mind that an underwater slide or slumping source consists essentially of sucking down a large mass of water to push it back up a short distance away and thus is fundamentally *dipolar* in nature, as opposed to a seismic dislocation which in gross approximation can be described as a monopole (Okal and Synolakis, 2001). Classical theories in all fields of physics show that dipoles are fundamentally high-frequency, short-wavelength sources with negligible far-field potentials. Therefore, it should not be surprising that the modeling of teleseismic tsunami amplitudes (Tanioka, 1999; Satake and Tanioka, 2000) should not require a slump source, but that should not be taken as evidence against a slump, because the far field is simply the wrong place to look for such evidence.

Further, as reported by Synolakis *et al.* (1997), the problem of hydrodynamic inversion of tidal gauge records is ill-posed, and no criteria exit for regularizing it to ensure uniqueness of the solution. Hence, the interpretation of distant tidal gauge records cannot be used to argue either side of the issue, particularly when their signatures are weak and noise-prone.

In addition, basic seismological principles state that no seismic wave can distinguish between the two fault-plane solutions of a point-source double-couple. This applies to tsunamis, which are normal modes of an Earth covered with an oceanic layer (Ward, 1980; Okal, 1982). This property will hold as long as both the epicentral distance and the wavelength used are large with respect to the dimensions of the source. In the present case, those numbers are at least 1,500 km (at Guam; more realistically 3,500 km in Japan), at least 150 km, and at most 40 km, respectively, and the data set offers very limited if any resolution between the two fault planes. Once again, the far field is the wrong place to resolve the fault plane indeterminacy.

Timing of the Tsunami

Regarding this critical issue, Geist (2000) elects to dismiss as "inconclusive" the data set assembled by Davies (1998) over the course of several months of field interviews. While we agree in principle that absolute timing of the arrival of the tsunami is bound to be imprecise under the circumstances, the careful analysis of Davies' data set, notably by Imamura and Ashi (2000), indicates a significant geographic pattern in the timing of the tsunami relative to the main aftershock at 09:09–09:10 GMT. We see no reason to dismiss this crucial observation more readily than, say, aftershock patterns obtained in a geometry possessing no depth resolution (see above).

The question of the radiation pattern of radial S waves then becomes moot in view of the time delay (dismissed by Geist, 2000) between the main shock and the slump. The S waves observed and analyzed by Kikuchi *et al.* (1999) are those of the 08:49 source and are unaffected by whatever took place 13 minutes later.

The above discussion shows that the model proposed by Geist (2000), generation of the local tsunami by a purely seismic dislocation on the steeply dipping plane, encounters formidable difficulties when confronted with existing data sets. Possibly because some of the work is presently in press or under review, Geist (2000) does not refer to a significant set of observations regarding the slump and its potential role in the generation of the tsunami.

First and foremost, the slump is present. It is well documented in the morphology of the seafloor in the amphitheater (Tappin *et al.*, 1999; 2000), and its presence, structure, and geometry were precisely mapped using seismic reflection by Sweet *et al.* (1999). While these techniques cannot put a precise date stamp on its occurrence, the mere presence of the fresh slump cannot be simply ignored.

Over the past two years, we have documented in several instances (Okal, 1998, 1999, 2000a, b) the existence of a seismic event at 09:02 GMT whose confidence ellipse includes the amphitheater and hence the locus of the slump. This event generated acoustic ("T") waves featuring exceptional characteristics of duration, amplitude, frequency content, and directivity of source blockage which are irreconcilable with a seismic dislocation of body-wave magnitude $m_b = 4.4$, but are generally consistent with the slump model derived in Synolakis *et al.* (2001) on the basis of the marine surveys.

The location of the slump and its proposed timing at 09:02 GMT combine to predict accurately the pattern of tsunami arrival times at the coast relative to the 09:09–09:10 aftershock, something no tsunami source coeval with the main shock can achieve (Heinrich *et al.*, 2000; Synolakis *et al.*, 2001).

HYDRODYNAMIC ASPECTS

We now address the issue of the hydrodynamic modeling of the tsunami. While appreciating Geist's (2000) discussion on inadequacies of present numerical methods, we wish to emphasize that hydrodynamical modeling may be used to speculate on the generation mechanism only if the same complete numerical code is used with both a "tectonic" initial condition and a "landslide" one.

Performance of Numerical Codes

For the record, and to our knowledge, most numerical simulations of the PNG event suggestive of landslide sources have been carried out using inundation codes such as Tohoku University's TUNAMI-N2 and USC/NOAA's MOST, which account for frequency dispersion and dissipation adequately, as evidenced by the solution of the benchmark problems presented in NSF's Friday Harbor Workshop (Yeh *et al.*, 1997). In particular, MOST uses an appropriate and constant number of grid points per wavelength, *i.e.*, the grid size is decreased as the wave propagates to shore, which achieves a consistent resolution regardless of the steepness of the waves. What was a pioneering idea when suggested by Goto and Shuto (1983) has now become a routine technique.

The Question of Breaking

First, we cannot help but note that Geist (2000) builds his case around witness reports of wave breaking while at the same time he dismisses as unreliable the timing of the tsunami relative to the main aftershock by what must have been the same group of witnesses.

Turbulence during wave breaking is not believed to affect the global evolution of the wave, as breaking—particularly for subduction zone tsunamis—is believed to be a highly local effect taking place very close to shore, often treated as physical "noise." In this respect, the reference to Sato (1996) is not appropriate, as no one- or two-dimensional numerical procedure can correctly predict the evolution of waveform through breaking: It is well established that turbulent dissipation is entirely different in two and three dimensions, not to mention that the scaling of this physical phenomenon remains essentially controversial, with the result that breaking is often handled in an ad hoc fashion in numerical simulations.

Numerical procedures cannot predict the details of the wave evolution through breaking, except perhaps the overall profile. Thus, and on account of the inherent dissipative nature of their differencing algorithms, both TUNAMI-N2 and MOST can still produce adequate results when mild breaking takes place in one region of the computation, as demonstrated by the satisfactory prediction of run-up during the extreme inundation flows of the Hokkaido-Nansei-Oki tsunami of 1993 (Titov and Synolakis, 1997).

Based on these two paragraphs, the predictions of these inundation codes cannot be discarded outright in favor of analytical results.

Use of an Analytical Approach

There are two fundamental shortcomings to the analytical approach of Geist (2000): First, it uses a single slope, which fails to model the continental shelf properly; second, it remains one-dimensional. On both accounts, it is inconclusive to compare his results with those of a three-dimensional numerical simulation using real bathymetry. We will see that the two factors actually have opposite effects on the wave amplitude at the shore.

To explore the first issue, we used Kanoğlu and Synolakis' (1998) analytical results for wave run-up on composite slopes to obtain correct amplification factors to be used as kernels in the run-up integrals. Specifically, we used a slope profile composed of two linear segments: a 3,500-m-long ramp at a 15° angle, from the source depth of 1,600 m to a transition depth of 676 m, and a 19-km-long shelf sloping gently at 2° from 676 m to the shoreline. Over the range of relevant normalized wave heights (from 0.0001 to 0.01), we found that the run-up over the composite beach was about twice as much as over an equivalent uniformly sloping beach. This was entirely unexpected and we can only attribute it to a resonant combination of wave and shelf parameters. This would lead one to believe that the 8 m height calculated by Geist (2000) could become 16 m over a composite beach, which would match exactly the extremum of the field observations.

However, this ignores the second issue, the three-dimensional character of the problem. To explore its effect, we generated a one-dimensional initial tsunami field by essentially cutting a one-dimensional slice through the three-dimensional landslide profile used in Synolakis *et al.* (2001). We then propagated this wave analytically using a one-dimensional algorithm and calculated a run-up of 30 m using a single slope and 60 m using a composite beach, as compared with the 16 m found using MOST and TUNAMI-2. This suggests that the dispersion of the energy in three dimensions decreases the amplitude of the wave at the shore by a factor of at least 3.5, as compared to values predicted using a hypothetical one-dimensional model. The implication of this experiment is that, even if one uses a composite beach slope and thus increases the 8 m predicted from Tadepalli and Synolakis (1996) by a factor of up to 2, the effects of three-dimensionality and directionality reduce the amplitude down to at most 4.5 m, thereby producing larger differences with the observations than implied by Geist (2000).

CONCLUSIONS

In conclusion, the argument in Geist (2000) is to cast doubt on the performance of numerical modeling methods on account of his perception of computational inadequacies and to prefer an analytical solution, which under an optimal, albeit not justified, choice of his constant γ can result in an 8 m run-up. We show that the shortcomings of this approach actually lead to overestimating the final amplitude to the extent that the maximum run-up predicted for the dislocation model (4.5 m) provides only a mediocre fit to the observed values.

Based on the above discussion of both seismological and hydrodynamic issues, we believe that the local devastating tsunami was the result of the slump documented inside the amphitheater by the surveys and which took place at 09:02 GMT, *i.e.*, 13 minutes after the main shock, as evidenced by the hydroacoustic records. Several independent modeling efforts using various modern codes (Imamura and Ashi, 2000; Heinrich *et al.*, 2000; Synolakis *et al.*, 2001) do provide acceptable, if arguably not perfect, matches to the observed run-up heights.

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Department of Geological Sciences
Northwestern University
Evanston, IL 60201, USA
emile@earth.nwu.edu
(E.A.O.)

Department of Civil Engineering
University of Southern California
Los Angeles, CA 90089, USA
(C.E.S.)