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Discussion

Comment on "Source of the great tsunami of 1 April 1946: a landslide in the upper Aleutian forearc", by G.J. Fryer et al. [Mar. Geol. 203 (2004) 201–218]

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In a recent contribution, Fryer et al. (2004) have proposed to interpret many features of the 1946 Aleutian earthquake and tsunami as evidence for a major earthquake-triggered landslide. In particular, in support of their model, they claim that the T phase observed at Hawaiian Volcano Observatory (HVO) was generated at the time of the main shock, and dispute our earlier interpretation (Okal et al., 2003a; hereafter Paper I) that it is about 29 min late in this respect and was actually generated by the first major aftershock, occurring 27 min after and 86 km to the North of the main shock.

The scope of the present *Comment* is strictly limited to the issue of the interpretation of the T phase at HVO. In particular, we elect, at this point, not to comment on any aspect regarding the characteristics of the far-field tsunami, which may be addressed in a later contribution; this should not, however, be taken as an endorsement of Fryer et al.'s model (Fryer et al., 2004) for the specific properties of the Ugamak slide, or more generally for the generation of the tsunami. We simply wish to address several points raised specifically in Section 8 of Fryer et al. (2004): (i) respond to their claim that our timing of the T wave train is wrong; (ii) correct the false impression left by their work that T phase amplitudes should be in direct relation to those of P waves, and (iii) emphasize that we presently know very little about the T phases of underwater landslides.

1. Timing the *T* phase at HVO

The identification of the source of the T wave train shown on Fig. 1 (adapted from Fig. 7 of Paper I) hinges crucially on a correct timing of the trace on which the high-frequency energy is observed. We certainly regard the reading of mingled traces on historical seismograms as an occasionally difficult endeavor, which may appear as more of an art than rigorous science to the lay reader. We realize that, in the framework of Paper I, we were not able to provide all the necessary details, which might have made our argument more convincing. We are pleased to give here a full description of our identification of the timing of the T wave train, based on several redundant measurements and drawing on more than 25 years of experience using historical seismograms (Okal, 1977).

First, we emphasize that the full original smoked paper E–W Bosch–Omori seismogram was photographed on a light table at the Hawaiian Volcano Observatory using more than 20 frames of 35-mm film, with significant overlap between the frames to ensure the continuity of the copied record. A large number of prints were made from the individual negatives, both commercially and using a standard 21× analog reader-printer. These prints were subse-

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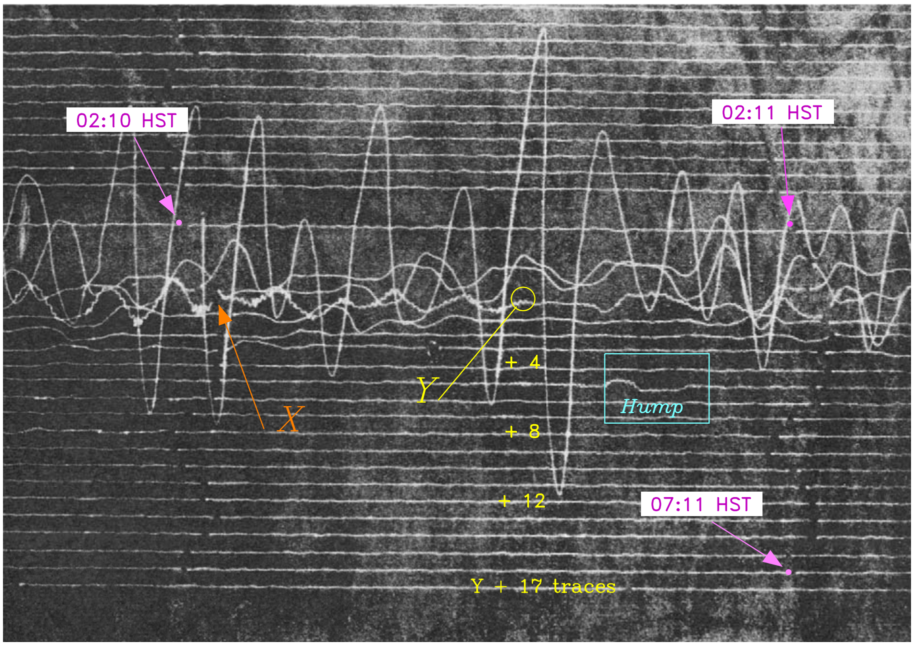

Fig. 1. Closeup of the T wave train at HVO, as recorded on the Bosch–Omori E–Wseismograph. The minute marks are confirmed by tracing continuously to a nearby hour mark and by timing the arrival of the Love wave train G_1 (see Fig. 2). See text for details of the identification of minute mark X occurring in the middle of the T wave train as 03:10 HST. The open circle (labeled Y) identifies the coda of the T phase, taking place at a time when the intermediate traces have quieted down and no longer interfere with the trace on which the T arrival is present. Also shown is the instability in the shape of a hump, occurring at 04:25:40 (adapted from Fig. 7 of Paper I).

quently enlarged using both a traditional photocopier and a digital scanner.

As mentioned in Paper I, we were careful to time the T arrival *relative* to well-documented phases, which can be recognized on the record (this is equivalent to checking for any possible error in the absolute timing of the HVO clock on that day). For the record, we recall that HVO has always, and to this day (2004), used Hawaiian Standard Time (HST), and that before 1947, the latter was GMT-10:30 (it is presently GMT-10).The principal phase used for timing was the Love wave G_1 from the main shock, whose group time at HVO is expected at 12:43:40 GMT (or 02:13:40 HST in 1946) for $U = 4.4$ km/s. As shown on Fig. 2, it is indeed observed around 02:14:00 HST; the difference is easily ascribed to the exceptional duration of the source of the main shock, the origin time (12:29:01 GMT) used to compute the G_1 group time having been derived from the relocation of the event (López and Okal, 2002), and thus being a hypocentral time rather than a centroid one. We conclude that no significant clock error exists for this record, and that we can legitimately consider absolute times (as locked, for example by the large hourly time marks, some of which are penciled for the correct hour on the original record) as accurate. We also verified through an independent check of minute marks that the drum rotated once every 15 min.

Fig. 1 (or the bottom part of Fig. 7 of Paper I) shows that the T wave train occurs four traces, i.e., 1 h, after the slightly wiggly trace corresponding to arrivals between P (expected at 02:05:55 HST) and S (02:11:26). While the latter trace is very quiet during the time window shown on these figures, its amplitude is significantly greater a few minutes earlier, around 02:08 and 02:09 HST, and this higher level of signal is perfectly compatible with the P (and PP, etc.) coda preceding the S arrival. Furthermore, we followed this trace continuously in both directions to the :00/:15/:30/:45-min marks and established beyond any doubt the timing labeled 02:10 HST (12:40 GMT) on Fig. 1.

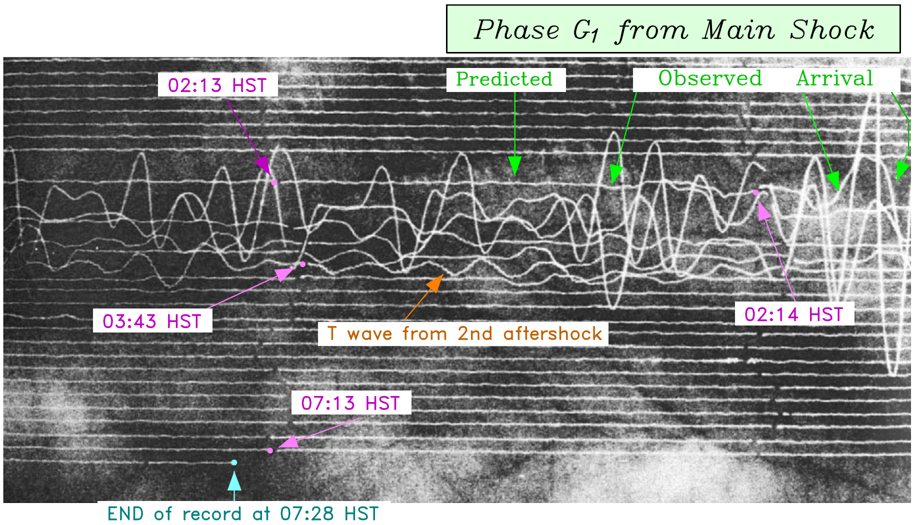

Fig. 2. Closeup of a later section of the same record as on Fig. 1. Note the G_1 arrival at right (arrows), which starts emerging from the record only 12 s after the group time predicted theoretically from the hypocentral parameters inverted by Lo'pez and Okal (2002).This figure also documents the termination of the record at 07:28 HST. Finally, note the faint high-frequency signal shortly after 03:43 HST, which is the *T* phase from the 13:29 GMT (02:59 HST) aftershock.

However, this exercise involved crossing the edge of the paper, where we documented that about 3 s of record is missing (a frequent occurrence when dealing with historical records); this makes it questionable how Walker and Okubo (1994) had managed to continuously follow the trace between the *P* and *T* arrivals, as quoted by Fryer et al. (2004).

We then take Fryer et al.'s criticism of our interpretation to mean that they dispute the figure of four intervals (1 h) separating the ''coda *P*'' trace (02:10–02:11 HST on Fig. 1) and the *T* wave train, on account of the mingling between traces. They surmise that the *T* phase arrives ''either 30 min too early or 15 min too early to have originated from the [main] aftershock''. We note that this statement by Fryer et al. (2004) is further incompatible with their claim that Walker and Okubo (1994) had followed continuously the trace on the record from the *P* arrival to the *T* phase. Had they successfully done so, there could not remain an uncertainty of one trace (''either 30 or 15 min'') in their timing.

In order to resolve this controversy, we carry out three independent checks of the timing of the *T* phase signal. Consider the minute mark occurring during the *T* phase arrival and labeled ''13:40 GMT'' on Fig. 7 of Paper I. We will now refer to it as minute mark *X* (Fig. 1), and our purpose shall be to provide an independent proof that *X* = 13:40 GMT (or 03:10 HST).

(1) We consider on Fig. 1 a part of the coda of the *T* phase (labeled *Y*), approximately 35 s later than minute mark *X*. Note that by then, the amplitude of two of the three traces separating the *T* trace (on which *T* phase energy is still distinct) from the *S* precursors between 02:10 and 02:11 has decayed to the extent that they are no longer mingled with the *T* trace. There can be no doubt that the *T* phase occurs on the fourth trace following the *S* precursors, i.e., that *X* = 02:10 + 1 h, or 03:10 HST.
(2) As a second independent proof, we time the *T* phase relative to the later, rather than sooner, part of the seismogram, given that traces are expected to be much smoother and suffer less mingling after the *T* phase, rather than before it. Consider, again on Fig. 1, that it is 17 intervals (or 4 h 15 min) to the bottom trace on the seismogram. We established that the record was interrupted at 07:28 HST (Fig. 2), which is to say that that section of the bottom trace corresponds to 07:25 HST, so that *X* = 07:25 – 04:15 = 03:10 HST.
(3) Finally, note the ''humpy'' signal (probably some instrumental instability), five traces and about 10 s

later than Y. We followed continuously this trace to the indisputable 04:30-min mark, and thus established that the hump occurs at 04:25:40. Hence, $Y = 04{:}25{:}40 - 01{:}15{:}10 = 03{:}10{:}30$ to a precision of a few seconds, and in turn, $X =$ 03:10 HST. *QED*.

We stand firmly by our earlier identification of the timing of the T wave train on the HVO record as 03:10 HST or 13:40 GMT. It cannot have been generated by the main shock.

2. How could the 12:55 GMT aftershock generate a larger *T* phase than the 12:29 main shock?

Whatever the nature of the 12:29 event (and we do not dispute that a landslide most probably took place concurrently with the earthquake; this is the only way to explain the near-field tsunami; Okal et al., 2003b), that event was anomalously slow, which, very simply, means that it was a deficient generator of high-frequency energy. This can be clearly demonstrated, for example by comparing the teleseismic P waves recorded on the Pasadena short-period Benioff seismometer from the main shock (origin time 12:29 GMT) and the main aftershock (O.T. 12:55). As shown on Fig. 3, the latter are stronger than the former, which are comparable in amplitude to those of the second aftershock (O.T. 13:29). This is supported by the report by Sanford (1946) that the main aftershock was felt stronger than the main shock at Scotch Cap, incidentally, a point underscored in Paper I but ignored by Fryer et al. (2004). On account of its very slow rupture velocity, the main shock was significantly deficient at frequencies typical of body waves or detectable by humans (typically 1 Hz). This is further documented by what amounts to the lowest ever measured energy-to-moment ratio, featuring a parameter $\Theta = \log_{10} E/M_0 \approx -7.0$ (López and Okal, 2002). At the higher frequencies characteristic of T waves, the deficiency is expected to be only exacerbated.

In summary, there is nothing anomalous in the generation of a detectable T phase by the 12:55 aftershock; and indeed, we have shown in Paper I that the smaller aftershock at 13:29 GMT also generated a T wave, barely emerging from the noise at 03:43 HST (see Fig. 2). The anomaly rests with the 12:29 main shock, which does not generate a T phase visible at HVO, because it is a slow earthquake; it has been recognized as such ever since the pioneering work of Kanamori (1972). As pointed out in Paper I, the main shock–12:55 aftershock combination constitutes a typical ''tsunami earthquake–regular earthquake'' duo; the whole point of Paper I, illustrated dramatically on its Fig. 3, was precisely to document and explain the T wave deficiency, in such sets, of the former source relative to the latter.

3. How could the 12:55 GMT aftershock generate a *T* phase comparable in amplitude to that of the much larger M_s=8.1 event to the northeast on 10 November 1938?

We are somewhat baffled by this question, which ignores the simple fact that one should not expect any high-frequency *amplitudes* to keep growing with seismic moments for very large sources. In a landmark paper, Geller (1976) has explained the saturation of *any* fixed-period magnitude (such as M_s at 20 s or m_b at 1 s) as a result of the destructive interference of time- and space-lagged elements of the source as the latter grows. As pointed out originally by Talandier and Okal (1979), because T phases are limited in frequency by the size of the SOFAR channel, their amplitude, the conceptual ingredient of a 3-Hz magnitude, is not expected to grow indefinitely with moment. This is why a better measure of earthquake size from T waves must involve their *duration* rather than their amplitude (Okal and Talandier, 1986), as implicit in the T phase energy flux (TPEF) developed in Paper I. Simply put, for very large events, it is a mistake to expect T phase *amplitudes* to scale with earthquake size. As a simple proof, let us recall that the T waves of the 1958 Fairweather, Alaska earthquake ($M_0 = 6 \times 10^{27}$ dyn-cm) were felt in Hawaii [J.P. Eaton, personal communication, 1979] and those of the great 1964 ''Good Friday'' earthquake ($M_0 = 8.1 \times 10^{29}$ dyn-cm) were not. Even the T waves of the 1960 Chilean earthquake ($M_0 > 2 \times 10^{30}$ dyn-cm) were not felt at transpacific distances, but they were recorded instrumentally at sustained amplitude for more than 5 min (Eaton et al., 1961).

In addition, we note (i) that the mechanism of generation of T phases at the location of the 1938

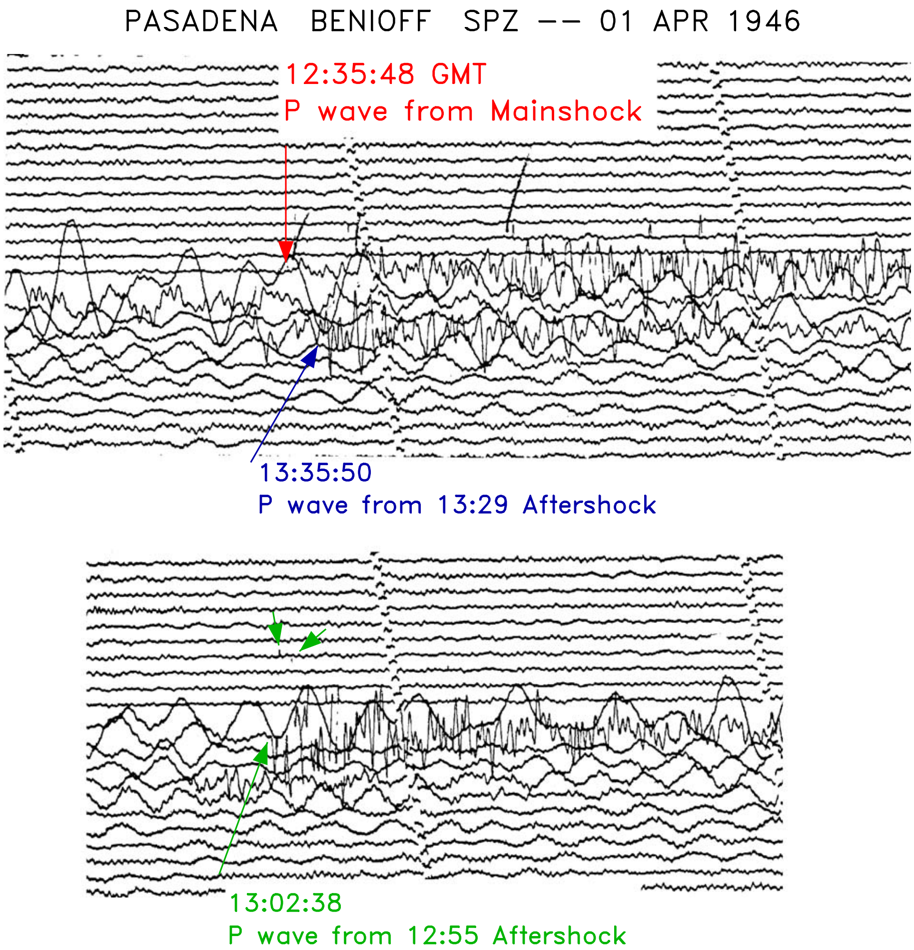

Fig. 3. *P* wave arrivals recorded at Pasadena on the short-period Benioff vertical seismometer. Top: arrivals from the main shock (red) and 13:29 GMT aftershock (blue). Note the low-amplitude, low-frequency, and long duration characteristics of the arrival. Bottom: arrival from the 12:55 GMT aftershock. The short green arrows point to the maximum excursion of the light spot on the paper, emphasizing the larger amplitude of the phase, as compared to that of the main shock. Note also the generally higher frequency content of the arrival. Time marks are at 1-mn intervals. For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.

earthquake may not be directly comparable to that at Ugamak; (ii) that the 1938 earthquake could have exhibited a slight trend towards slowness, comparable to that of the 2001 Peruvian earthquake (Okal et al., 2002), which resulted in a significantly reduced TPEF, as compared to the neighboring 1996 Nazca event; (iii) that Walker and Okubo (1994) used a N–S, rather than E–W, component for the 1938 record, which would preclude any direct comparison of amplitudes; and finally, (iv) that Walker and Okubo (1994) themselves guard against absolute comparisons of *T* wave amplitudes between events spanning several years, quoting at length a detailed description by Apple et al. (1987) of the repeated adjustments of the damping and recording characteristics of the mechanical instruments in use at HVO in those days.

4. Why then was the Ugamak slide hydroacoustically silent?

Although this question is not asked by Fryer et al. (2004), it is legitimate in the framework of our conclusion, mentioned in Paper I, that there is no signal at HVO around the time (13:11 GMT or 02:41 HST) expected for a T phase generated at the time of the main shock. We have argued in Okal et al. (2003b) that the near-field tsunami characteristics require generation by a landslide coeval with the main shock. Why then was the landslide silent?

Although we do not have a definitive answer, we propose to invoke the probable slowness, i.e., long duration, of the source, a property itself stressed by Fryer et al. (2004). A long, slow landslide would be expected to be a deficient T phase generator.

First, we wish to stress that the conditions at HVO were particularly unfavorable in 1946 for the detection of T phases; the receiver-side land path across the island of Hawaii is at least 70 km long, and the seismometer had a poor magnification given as 116 (in the limit $\omega \rightarrow \infty$) by McComb and West (1931).

Second, we should emphasize that we presently remain essentially ignorant of the properties of landslides as hydroacoustic sources. The case of the T phase identified from the 1998 Papua New Guinea (PNG) landslide (Okal, 2003) is unique and may not be directly comparable to the 1946 situation, given the much steeper slope in PNG [15° (Sweet and Silver, 2003) as opposed to 4° quoted at Ugamak by Fryer et al. (2004)] and the containment of the PNG slide within the amphitheater. It should also be kept in mind that the anomalous character of the PNG T waves stemmed from their long *duration* (Okal, 2003); their amplitude was relatively weak, and such T phases would never have been detected by seismometers under the mediocre conditions at HVO in 1946. While a large slide at Ugamak would undoubtedly have generated T waves of longer duration, we really have no clue whether their amplitude would have been larger than those of the 1998 PNG source.

As for the bench ruptures at Hawaii studied by Caplan-Auerbach et al. (2001), they involved essentially unsedimented edifices, and generated T waves of weak amplitude, a few of which were detected at transoceanic distances by hydrophones, but not by modern high-gain (let alone historical) seismic stations. Among the positively identified underwater landslides of the past decades, the events at Kitimat (1975) and Skagway (1994) (Murty and Brown, 1979; Synolakis et al., 2002) occurred in fjords with no acoustic output to the open ocean; to our knowledge, the Nice slide of 1979 (Assier-Rzadkiewicz et al., 2000) did not generate observable T phases in the Mediterranean. The 1975 Kalapana, Hawaii event did generate strong T waves (Talandier and Okal, 2001), but it most probably consisted of both an earthquake and a landslide with unknown partitioning of the T energy between the two.

Underwater slides may, after all, be hydroacoustic sources of subdued amplitude, if occasionally sustained duration. Our experience in Papua New Guinea would suggest that they feature some of the lowest values of the amplitude-duration discriminant D introduced by Talandier and Okal (2001); indeed, it is a variation of that property that revealed the anomalous nature of the 09:02 tsunamigenic event during the 1998 PNG disaster. While it is impossible to even envision applying these authors' algorithm to a record with the poor signal-to-noise ratio shown on Fig. 1, we wish to stress that the 1946 T wave train recorded at HVO is identifiable above noise level for no longer than 100 s, with its maximum amplitude sustained for only about 30 s. It certainly does not give the perception of an exceptionally long signal, as would be expected from the long landslide sources (300 s or more) modeled by Fryer et al. (2004).

Thus, given a low enough amplitude below the threshold of the old instruments at HVO in 1946, it is entirely possible that the T waves of the Ugamak slump would have gone undetected.

In conclusion, we reject Fryer et al.'s assertion that our interpretation of the 1946 HVO T wave train is erroneous. We stand firmly by our conclusion that no T waves generated simultaneously with the main shock are detectable on that record. This is perfectly well explained in the context of classical seismological source theory, once we admit the extremely slow character of the main shock and the probable inefficiency of the landslide as a T phase generator.

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References

Apple, R.A., Jaggar Jr., T.A., Hawaiian Volcano Observatory 1987. In: Decker, R.W., Wright, T.L., Stauffer, P.H. (Eds.), Volcanism in Hawaii, vol. 2. U.S. Geol. Surv. Prof. Pap. 1350, pp. 1619–1644 (Washington, DC).

Assier-Rzadkiewicz, S., Heinrich, P., Sabatier, P.C., Savoye, B., Bourillet, J.-F, 2000. Numerical modelling of landslide-generated tsunami: the 1979 Nice event. Pure Appl. Geophys. 157, 1707–1727.

Caplan-Auerbach, J., Fox, C.G., Duennebier, F.K., 2001. Hydroacoustic detection of submarine landslides on Kilauea Volcano. Geophys. Res. Lett. 29, 1811–1813.

Eaton, J.P., Richter, D.H., Ault, W.U., 1961. The tsunami of May 23, 1960 on the island of Hawaii. Bull. Seismol. Soc. Am. 51, 135–157.

Fryer, G.J., Watts, P., Pratson, L.F., 2004. Source of the great tsunami of 1 April 1946: a landslide in the upper Aleutian forearc. Mar. Geol. 203, 201–218.

Geller, R.J., 1976. Scaling relations for earthquake source parameters and magnitudes. Bull. Seismol. Soc. Am. 66, 1501–1523.

Kanamori, H., 1972. Mechanisms of tsunami earthquakes. Phys. Earth Planet. Inter. 6, 346–359.

López, A.M., Okal, E.A., 2002. Aftershock relocation, rupture area, mantle magnitude and energy estimates of the 1946 Aleutian tsunami earthquake and neighboring events. EOS Trans. Am. Geophys. Union 83 (47), F1045 (abstract).

McComb, H.E., West, C.J., 1931. List of seismological stations of the world. Bull. Natl. Res. Counc. 82 (119 pp., Washington, DC).

Murty, T.S., Brown, R.E., 1979. The submarine slide of 27 April, 1975, in Kitimat Inlet and the water waves that accompanied the slide. Pac. Mar. Sci. Rep. 79–11 (36 pp., Sidney, B.C.).

Okal, E.A., 1977. The July 9 and 23, 1905 Mongolian earthquakes: a surface wave investigation. Earth Planet. Sci. Lett. 34, 326–331.

Okal, E.A., 2003. *T* waves from the 1998 Papua New Guinea earthquake and its aftershocks: timing the tsunamigenic slump. Pure Appl. Geophys. 160, 1843–1863.

Okal, E.A., Talandier, J., 1986. *T*-wave duration, magnitudes and seismic moment of an earthquake: application to tsunami warning. J. Phys. Earth 34, 19–42.

Okal, E.A., Dengler, L., Araya, S., Borrero, J.C., Gomer, B., Koshimura, S., Laos, G., Olcese, D., Ortiz, M., Swensson, M., Titov, V.V., Vegas, F., 2002. A field survey of the Camana, Peru tsunami of June 23, 2001. Seismol. Res. Lett. 73, 904–917.

Okal, E.A., Alasset, P.-J., Hyvernaud, O., Schindelé, F., 2003a. The deficient *T* waves of tsunami earthquakes. Geophys. J. Int. 152, 416–432.

Okal, E.A., Plafker, G., Synolakis, C.E., Borrero, J.C., 2003b. Near-field survey of the 1946 Aleutian tsunami on Unimak and Sanak Islands. Bull. Seismol. Soc. Am. 93, 1226–1234.

Sanford, H.B., 1946. Log of Coast Guard unit number 368. Scotch Cap DF Station, relating to the Scotch Cap light station tragedy of 1946. U.S. Coast Guard, Washington, DC (11 pp).

Sweet, S., Silver, E.A., 2003. Tectonics and slumping in the source region of the 1998 Papua New Guinea tsunami from seismic reflection images. Pure Appl. Geophys. 160, 194–1968.

Synolakis, C.E., Yalçiner, A.C., Borrero, J.C., Plafker, G., 2002. Modeling of the November 3, 1994 Skagway, Alaska tsunami. In: Wallendorf, L., Ewing, L. (Eds.), Solutions to Coastal Disasters, Proc. Amer. Soc. Civil Eng., ASCE, Reston, VA, pp. 915–927.

Talandier, J., Okal, E.A., 1979. Human perception of *T* waves: the June 22, 1977 Tonga earthquake felt on Tahiti. Bull. Seismol. Soc. Am. 69, 1475–1486.

Talandier, J., Okal, E.A., 2001. Identification criteria for sources of *T* waves recorded in French Polynesia. Pure Appl. Geophys. 158, 567–603.

Walker, D.A., Okubo, P.G., 1994. The *T* phase of the 1 April 1946 Aleutian Islands tsunami earthquake. Sci. Tsunami Hazards 12, 39–52.