Tsunami warning: beating the waves to death and destruction

Emile A. Okal

Tsunamis — the massive oceanic oscillations that commonly follow earthquakes and other major disturbances in the earth's crust — can cause severe coastal damage and loss of life. This article reviews the way in which they develop and current techniques for predicting their onset within the relatively short space of time needed to warn coastal areas at risk.

Figure 1 View of the small town of Aonae, on the island of Okushiri, Japan, in the aftermath of the Japan Sea tsunami of 12 July 1993. Note the devastation wrought on the island by the tsunami wave; all housing in the left part of the photograph has been destroyed and the rubble washed out in the harbour; note also the fishing boats carried inland and the fires, still burning in this next-day photograph. (Courtesy of Y. Tsuji.)

Over a period of a little over 10 months, from 2 September 1992 to 12 July 1993, more than 2500 lives were lost to three disastrous tsunamis, in Nicaragua, Indonesia, and Japan (figure 1). Tsunamis are gravitional oscillations of seas or ocean basins, set up by major geophysical events such as earthquakes, landslides, and volcanic eruptions.*

While the cumulative death toll of tsunamis over the past few decades remains relatively minor as compared to that of earthquakes, typhoons, or volcanic eruptions, their impact is particularly striking due to the large geographic areas involved: gigantic earthquakes are capable of setting in motion the entire water body of the Pacific Ocean, bringing death and destruction to faraway shores. For example, the tsunami from the great Chilean earthquake of 22 May 1960 killed 200 people in Japan, 17,000 km away from its source, having left in its wake many island shores ravaged, including the city of Hilo, Hawaii, which suffered $20 million in damage and 61 deaths. In this respect, tsunamis come second only to major meteorological events (either cyclical such as El Niño, or triggered by volcanic eruptions) in the size of their potential area of disaster. The tsunami potential for destruction is illustrated by the fact that 62 per cent of all earthquake-related deaths in the United States over the past 50

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*In many languages, they have often been called 'tidal waves' (French: raz-de-*marée*; German: *Flut*welle), which is improper since they are not related to tides, and thus scientists the world over use the Japanese words *tsu-nami*, meaning literally 'harbour waves'.

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years were caused by tsunamis, principally the 1946 Aleutian and 1960 Chilean events.

In very simple terms, a tsunami is a long-wavelength oscillation of the mass of the ocean with typical periods in the range of 7 to 30 min. Its size and effect on a shoreline, especially at distances of more than a few hundred kilometres (the so-called teleseismic distances), are often decomposed into two factors: the amplitude of the wave on the high seas, which is primarily controlled by the size and geometry of the parent earthquake, and the response of the shoreline to the wave. The former is now relatively well understood in the framework of the linear theory of elasticity. The latter is a much more difficult problem, which must involve both the response of coastal features such as bays and harbours (which are capable of resonance, thereby amplifying selected wave frequencies), and the non-linear hydrodynamics of the breaking wave as it climbs up the shoreline (the so-called 'run-up' effect). For this reason, and since the high-seas amplitude is obviously an initial boundary condition in any computational scheme of run-up, we will discuss separately the theoretical tools available to each part.

Modelling the high-seas amplitude of a tsunami

The simplest, and historically first, model of a tsunami (or long wave) is that of an uncompressible fluid of depth H overlying a perfectly rigid substratum with perfect translational symmetry. In the long wavelength limit, the phase velocity of the wave is found to be $C = \sqrt{gH}$, where g is the acceleration of gravity [1]. Since C is independent of frequency, it is also the group velocity U at which the energy of the wave propagates. For a 5 km deep ocean, C is about 220 ms^{-1} or 800 kmh^{-1}. This, the approximate velocity of a jetliner, is slow by seismic standards: it takes a tsunami about 22 h to cross the Pacific from Chile to Japan (figure 2) and even 5 h for the much shorter run from the Aleutians to Hawaii. Sound waves in the water, on the other hand, would be about seven times faster; typical seismic surface waves, travelling around the Earth, about 18 times faster. This remark sets the stage for the possibility of distant tsunami warning by exploiting the time interval which separates seismic and tsunami waves.

A much more powerful approach nowadays consists of regarding tsunami waves as a particular family of the free oscillations of the planet, taken as a whole, and featuring of course an appropriate ocean layer at its top. This method, initially introduced by S. N. Ward [2], allows a more detailed representation of the elastic properties of the source region, including the presence of any number of crustal or sedimentary layers. Classical techniques of seismogram synthesis are easily adapted to construct 'marigrams'; that is time series of the oscillation of the height of the sea surface. The fundamental results, as summarized for example by E. A. Okal [3], are:

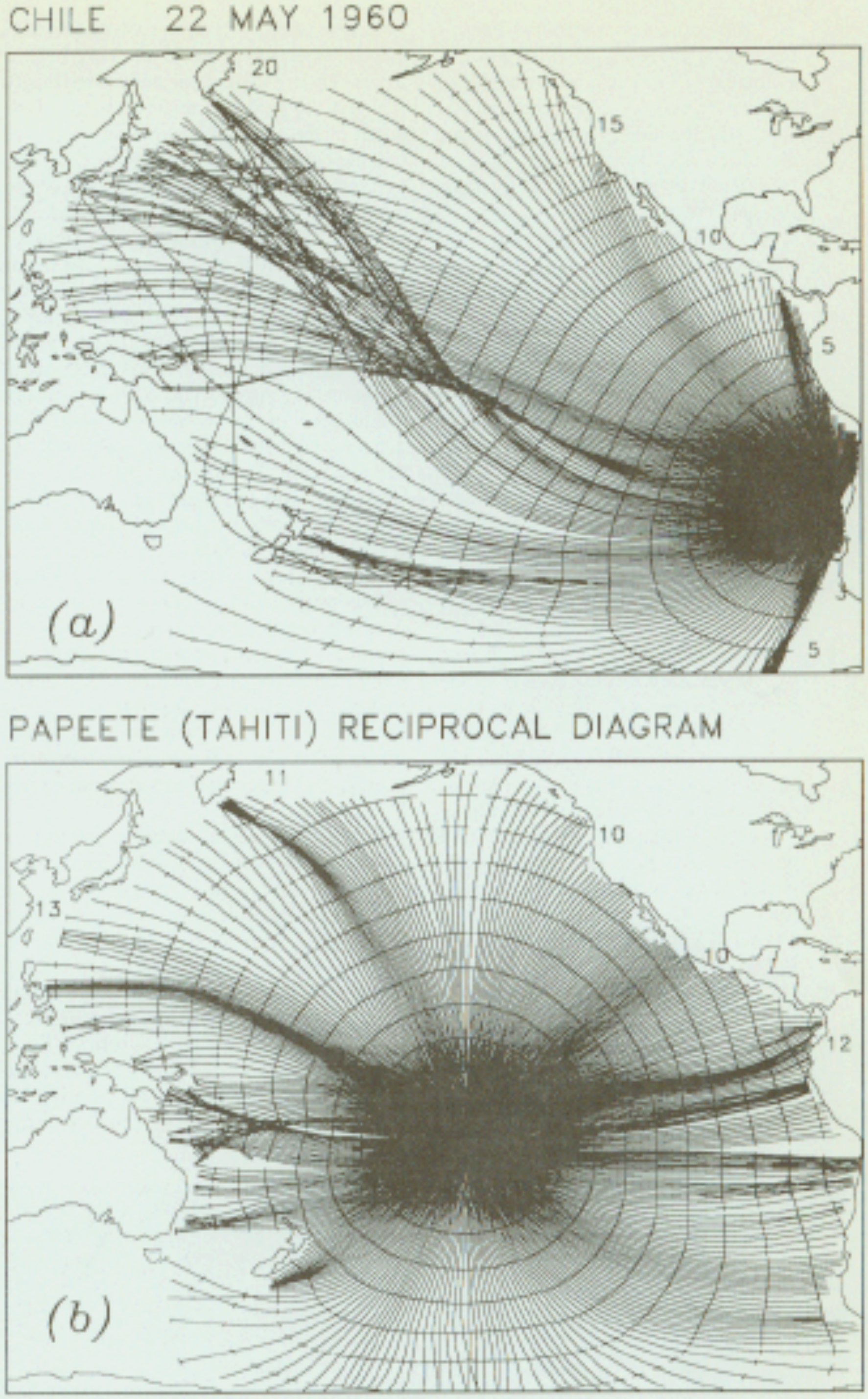

Figure 2(a) Tsunami paths traced in the Pacific Ocean from the epicentre of the 1960 Chilean earthquake. Shallow structures reduce the velocity of the wave and tend to focus its energy, in the same way as a lens focuses light, while deep, fast basins defocus the wave. The figure is obtained by shooting rays in 1° azimuth increments at the epicentre, and propagating them through the variable-velocity medium. Tick marks are group times in increments of one hour, with travel time (in hours) printed at selected coastlines.
(b) Reciprocal diagram for Papeete, Tahiti. Tick marks show lines of equal travel-time to Papeete (in increments of one hour and with selected values printed), while the areas of focusing indicate enhanced amplitudes in Tahiti for tsunamis originating in those regions.

(1) the most important parameter controlling tsunami generation is the seismic moment† of the parent earthquake;

(2) it makes a very large earthquake indeed to generate a transoceanic tsunami of any consequence (in practice, M_O =

†The seismic moment M_O of an earthquake is a measure in units of torque (dyn-cm or N-m) of the strength of a system of forces known as a double-couple, equivalent to the dislocation in the rupturing material. M_O has largely replaced the empirical concept of magnitude as the preferred descriptor of earthquake size.

10^{28} dyn-cm (or a traditional magnitude of eight) will generate a 5 cm tsunami (peak-to-peak) on the high seas at a distance of 4000 km). This amplitude, which would be greater by about an order of magnitude for a local tsunami, explains the relative rarity of the phenomenon;

(3) focal geometry and depth of source play only a minor role as long as the latter is no greater than 70 km;

(4) the presence of structures with deficient rigidity (such as sediments) between ocean and basement rock can affect their coupling and, if involved in the rupture, boost tsunami excitation by an order of magnitude;

(5) directivity (that is the direction in which rupture propagates along the fault) leads to interference patterns focusing tsunami energy perpendicular to the direction of rupture; meaning, in most geometries, towards the centre of the ocean;

(6) Propagation on the high seas is affected by the variation in the depth H of the ocean. The ensuing refraction by velocity gradients leads to focusing and defocusing [figure 2(a)], which can add or subtract about half an order of magnitude to the height of the tsunami wave. This figure is an exercise in geometrical optics, and as such lends itself to the principle of reciprocal path. One can thus prepare 'reciprocal charts' [figure 2(b) is an example at Papeete] predicting the travel times and focusing of tsunamis originating at a variety of epicenters.

Coastal response and run-up

We refer for example to [1] for a full treatment of this difficult problem, and will simply attempt to describe it in very general terms. There are fundamentally three length scales in the problem: the local value of the depth of the ocean H, the wavelength L of the tsunami, and the amplitude z of the vertical displacement at the surface of the ocean; the Ursell parameter, $U = zL^2/H^3$ controls the validity of various approximations. When U is very small, the linear long-wavelength approximation (described above), is valid; when U becomes of order 1, the effect of finite amplitude becomes significant and the so-called Boussinesq equations are appropriate. This is typically the regime of the continental shelf; the problem remains linear and has a number of solutions known as 'solitary' and 'conoidal' waves, which can be derived analytically with the help of elliptic functions. When U becomes very large, that is, in very shallow waters such as inside bays and harbours and during bore formation, the problem becomes non-linear because the surface boundary condition must be expressed at the true location of the ocean surface, which itself depends on z. The problem can be solved only by finite difference techniques, with the particular geometry of the coastline handled on a case-by-case basis.

In practice, coastal response and run-up can typically add up to one, and exceptionally two, orders of magnitude to the high-seas amplitude of the wave. As a rule of thumb, a tsunami will be destructive if its amplitude, including local and run-up effects, reaches a few metres. Thus, a distant tsunami will have a potential for disaster if its amplitude on the high seas reaches a few tens of centimetres, which according to our earlier remark, requires — in principle — a moment greater than 10^{28} dyn-cm.

Hazard mitigation

A significant component of hazard reduction can be achieved through adequate preparation. The latter can take several forms:

Relocation is probably the most drastic aspect of tsunami hazard mitigation. After being destroyed twice in 14 years, the waterfront business district of the city of Hilo, Hawaii was simply rebuilt further inland and the flood-prone area converted into a park — including a tsunami memorial. Similarly, critical areas of the city of Petropavlovsk, Kamchatka were displaced to higher elevations. Such procedures are only available when real estate is plentiful, and the political system has the strength to overcome the leniency of the population towards a relatively rarely recurring form of natural hazard. Building codes and proper zoning are a milder form of mitigation, and over a period of years, can change significantly the impact of destruction in a coastal area.

Walls and levees and similar waterfront structures are used extensively in areas such as Japan where land is at a premium and the hinterland inhospitable: it is obviously impractical to move several kilometres inland entire villages whose living is derived primarily from fishing. Such structures are elaborate in their design, which can be tested, both with computer models and in scaled-down laboratories, in order to minimize run-up, and to provide maximum obstruction to the debris (boats, trees, etc.) carried by the second and subsequent waves in the tsunami. A very large fraction of the Japanese coastline is presently lined with tsunami walls (figure 3).

Education remains a fundamental aspect of tsunami hazard mitigation. Thousands of

Figure 3 Example of tsunami wall in a typical fishing village (Ryoishi) along the Sanriku coast of Japan. Note the two groups of gates which allow traffic (vehicular — open at left; boat — closed at right), and which can be closed in case of a tsunami warning. The car near the left gate gives the scale: the wall is approximately 6 m tall.

lives were lost in the past when coastal fishermen, bewildered by a fast recess of the sea, rushed to collect shells and crabs from the exposed beaches, only to be swept away 15 min. later by the positive half of the oscillation ... In this respect, landmark programmes have been introduced in grade schools, notably in Japan and Chile [4], complemented by evacuation plans, of which a remarkable example is the yearly drill conducted in the port city of Callao, Peru, a suburb of Lima, in which more than 100,000 persons regularly participate [5].

In the United States, while the awareness of tsunami danger is rightly perceived at Government level, it is clear that further education of the general public and of the media is necessary, as demonstrated for example by the episode of the May 1986 Aleutian earthquake. A tsunami warning was issued for the Hawaiian Islands, with the result that half of the population left the low-lying beaches of Waikiki for the heights of the Pali hills (a mere 200 m away from the beach would in all cases have been enough), while the other half drove down to the beach to watch the show, with traffic totally gridlocked for the entire afternoon. At the same time, journalists in Oregon chartered helicopters to fly out at sea to 'meet the wave', disregarding repeated statements by scientists that the swell on the surface of the sea would be at most 25 cm, spread over a wavelength of 300 km! In the end, there was no perceptible tsunami, in either Hawaii or Oregon.

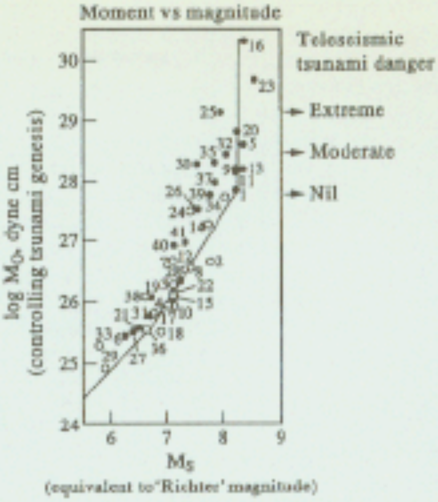

Figure 4 This figure illustrates the saturation of the conventional ('Richter') magnitude, M_S for very large earthquakes, after R. J. Geller [18]. The seismic moment M_0, which controls the excitation of tsunamis keeps increasing even for the largest seismic events, while M_S does not grow beyond 8.2. The scale at the right sketches qualitatively the level of tsunami danger at large distances from the earthquake. Clearly, the conventional magnitude cannot be used to assess tsunami danger.

Teleseismic tsunami warning

This episode brings us to the question of distant tsunami warning. Its fundamental goal is to use the significant time delay between the arrival of seismic and tsunami waves to evaluate the parent earthquake and assess its tsunamigenic potential. The main challenge of tsunami warning stems from the fact that measuring the seismic moment M_O of an earthquake (which controls the size of its tsunami) is a relatively difficult task. The moment is the amplitude of the deviatoric stress tensor, a mathematical object of dimension 5, whose full resolution necessitates the inversion of a large amount of seismic data. Thus individual observatories still rely largely on magnitude readings ('quick and dirty') measurements taken at single stations) to quantify earthquakes in the immediate aftermath of the event.

The problem is that magnitude scales, tied to a particular reference period of seismic waves (20 s for the surface-wave magnitude, M_S) saturate for very large events, when the duration of the source becomes comparable to the reference period (figure 4). In practice, M_S saturates around 8.2, which amounts to saying that a magnitude $M_S \geq 8.0$ is not a reliable descriptor of the true size of the event: the seismic moment M_O could vary from 10^{28} to more than 10^{30} dyn-cm. The former would be benign in terms of tsunami danger, the latter catastrophic; the exclusive use of traditional magnitudes for tsunami warning is thus dangerous, if not outright criminal.

When considering the case of distant islands isolated in the middle of a large ocean, self-reliance is important and even in these days of supposedly instant worldwide telecommunications, it is necessary to develop a method allowing for the real-time one-station estimate of the seismic moment of a distant earthquake. Following the development of broad-band seismometers, capable of a reliable rendition of the Earth's motion throughout the full spectrum of seismic waves from 1–300 s, and in collaboration with Dr J. Talandier (former Director of the Geophysical Laboratory in Papeete, Tahiti), we have introduced the concept of a mantle magnitude M_m [6]. It retains the philosophy of a magnitude (a one station, simple measurement) while at the same time avoiding the effects of saturation by considering all frequencies in the surface wave trains, including the longer waves, sampling deep into the mantle of the Earth (hence the name 'mantle' magnitude), and providing an estimate of the very-low frequency behaviour of the source. Specifically, M_m is computed from the spectral amplitude of the Earth's ground motion, $X(\omega)$, measured in μm-s through:

$$M_m = \log_{10}X(\omega) + C_S + C_D - 0.90, \quad (1)$$

where C_S and C_D are appropriate frequency and distance corrections, and the constant 0.90 ensures that M_m will be an estimate of the quantity $\log_{10} M_0 - 20$. Having verified on extensive datasets that the mantle magnitude does not saturate, even for the largest events ever recorded, it then became feasible to derive a self-contained, automated algorithm for the evaluation of distant tsunami risk [7].

This system, known as TREMORS, is built around a 3-component broad-band seismic station and is composed of a teleseismic event detector, an event locator, and a software algorithm providing in real-time an assessment of the moment M_O. The event detector and locator use standard seismological practice, based on the recognition and interpretation of the body wave phases P and S of the seismogram. The distance information is used to obtain the window of arrival times of the mantle Rayleigh waves, which are then dumped into the computer and the calculation (1) performed; the combination of M_O and distance can in turn be used to estimate the high seas amplitude (A, in cm, peak-to-peak) of the tsunami wave upon reaching Polynesia through

$$\log_{10} A = \log_{10} M_O - 0.5\log_{10} (\Delta \sin\Delta) - 26.4, \quad (2)$$

a formula derived from the theoretical investigation of tsunami excitation (Δ is expressed in degrees along the great circle route: $1° = 111.2$ km). the algorithm outputs automatically the estimates of location, seismic moment, and tsunami amplitude for Papeete, a location where harbour response and run-up are usually negligible. Additional corrections can be included for other Polynesian sites. A geophysicist, on call 24 hours a day, is automatically alerted, and has the ultimate responsibility of recommending to civilian authorities either a tsunami watch, or a tsunami warning, or no action. Values of M_m are also transmitted in real time to international organizations, such as the Pacific Tsunami Warning Center in Ewa Beach, Hawaii, and the National Earthquake Information Center in Golden, Colorado.

A flow chart of the algorithm, and an example of the processing of an earthquake are given on figure 5. Ever since the system was implemented in 1987, more than 500 earthquakes have been processed; its performance can be measured by the average value ($\bar{r} = 0.07$ units of magnitude) and standard deviation ($\sigma = 0.22$) of the residual between our real-time measurements at Papeete, and values subsequently published after a full investigation. While no tsunami warning has been issued for Polynesia during this period, a number of false alarms have been avoided — including for the 1986 Aleutian event mentioned above [8, 9].

Warning at short distances

A number of tests have shown that TREMORS can be applied to epicentral distances as short as 150 km, before it becomes unreliable due to the breakdown of a number of theoretical approximations. In addition, at those short distances, it becomes

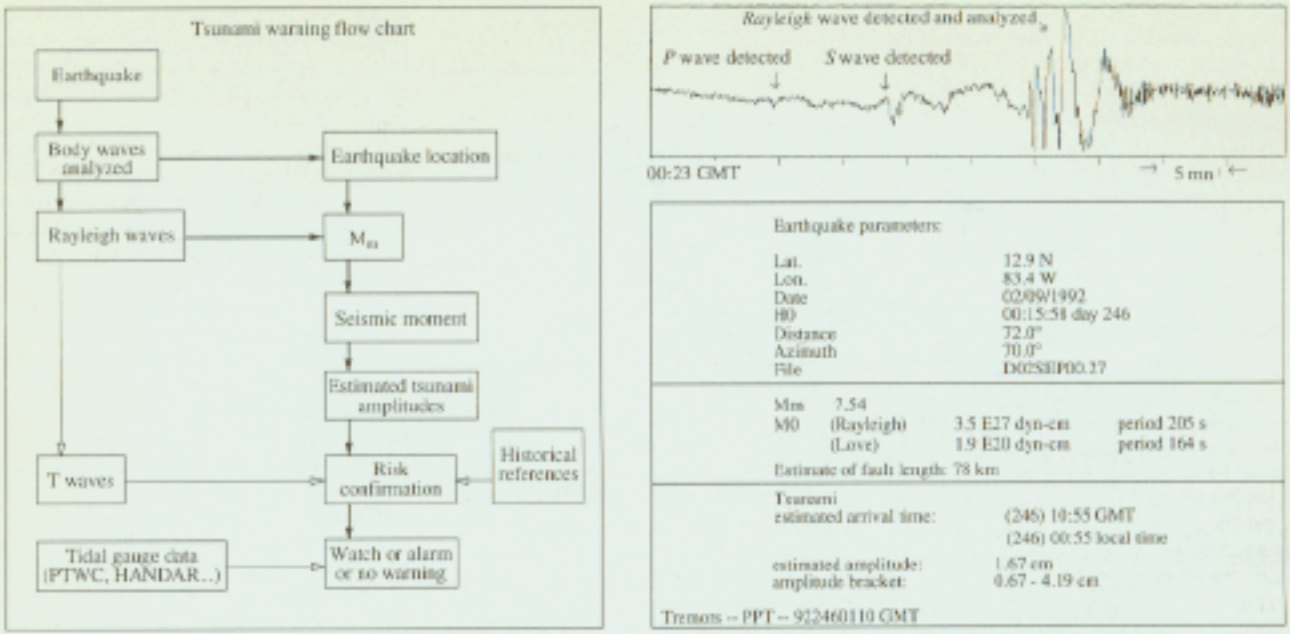

Figure 5 *Left*: flow chart of the TREMORS tsunami warning algorithm. The thinner traces (with open arrows) are complementary ingredients occasionally available to the geophysicist to help him confirm the level of tsunami risk. The thicker traces represent the various steps in the automated algorithm. *Right*: application to the 1992 Nicaraguan earthquake as recorded at Papeete. The bottom panel is a copy of the output from the computer, as printed at 0110 GMT, 55 mn after the origin time of the event. It gives estimates of the epicentre, of the mantle magnitude and of the expected parameters (arrival time and amplitude) of the potential tsunami.

Figure 6 View of the southern tip of Okushiri island at Aonae, showing total destruction of the residential area to the South of figure 1. Note that the 4.5 m tall tsunami wall was overrun, and some of its sections even completely destroyed. (Courtesy of Y. Tsuji.)

inappropriate to separate the generation of the tsunami wave on the high seas from the response of the coast line. Finally, the question of tsunami warning meets the fundamental difficulty of the reduced time delay between seismic and tsunami waves. For an epicentral distance of 100 km, and assuming an average tsunami velocity of 200 ms^{-1}, only 8 mn will separate the first seismic wave (either felt or recorded) from the tsunami. Is there enough time to properly record the event, assess the hazard, order and above all, carry out an evacuation? While this may seem like an extremely short time, it is clear that some level of mitigation can be achieved, for example by closing the harbour doors on figure 3. It is also obvious that in such cases, education of the population becomes critical, with the only form of practical tsunami warning at very short distances being 'If it shakes really strongly, waste no time and run for the hills'.

This, however, is not always possible. In the case of the tsunami of 12 July 1993 in the Sea of Japan, more than 100 people died in the town of Aonae, on Okushiri Island; the high death toll was a combination of the size of earthquake (6×10^{27} dyn-cm), the proximity of the epicenter (80 km), the geometry of the island which focused the wave on to Aonae, and finally the time of the day (10:30 p.m. local time), which caught the population resting at home and in the dark. The breakwater walls did reduce the run-up, but unfortunately they were themselves overrun (figure 6).

'Tsunami Earthquakes'

Enter the ominous 'tsunami earthquakes'. This term, coined by H. Kanamori [10], describes seismic events whose tsunami is significantly larger than would be expected from its seismic waves. Examples include the great 1896 Sanriku event in Japan, the 1946 Aleutian earthquake, and more recently the 2 September 1992 event in Nicaragua, in which 160 fatalities were reported locally. The latter is an example of a 'slow' earthquake, whose energy took a long time (180 s) to build up, with the result that its body-wave magnitude m_b (measured at a 1 s period and characteristic of the social effects of the event) was only 5.3, while its seismic moment reached 3×10^{27} dyn-cm. Consequently, in certain sections of the Nicaraguan coast, the event was not even felt by part of the population, who were then swept away by the tsunami a few minutes later. The seismic moment was correctly assessed at Papeete, but by the time the Rayleigh waves had reached Polynesia, the tsunami had already devastated the Nicaraguan coast.

While local tsunami warning of slow events remains a challenge, the adoption of modern seismological techniques based on the processing of the ultra-low frequency part of the seismic spectrum [6, 11] should solve the problem of recognizing in real time their true potential both at regional and teleseismic distances.

More complex are those events for which no level of investigation of their seismic records has managed to explain the exceptional size of their tsunamis. Examples include the 20 October 1963 Kuriles event, and the 1 April 1946 Aleutian earthquake, the latter is generally considered to have unleashed the most powerful tsunami of the century (173 deaths in Hawaii) despite a relatively low magnitude of 7.2. A number of investigators has proposed to invoke the role of sediments in order to explain them. This can take the form of seismic rupturing through the sedimentary structure (Kuriles [12]) or more importantly, of the massive slumping of an unstable sedimentary structure on the ocean floor, caused by the earthquake. The latter can enhance the tsunami excitation, since it exchanges large masses of water and sediments; however the modelling becomes very difficult since the cohesiveness of the material at the bottom of the sea is destroyed during the source process. Several techniques are available to recognize landslide components in the seismic signature of earthquakes, but their application remains difficult [13, 14] and in the case of the 1946 event, neither the quantity nor the quality of the seismic data are sufficient to allow an appropriate modelling.

Local underwater slumping has also been invoked in the case of the 12 December 1992 tsunami in the Sea of Flores, Indonesia, in which more than 2000 people were killed. It would explain excessive run-up (up to 27 m) at Kroko, at the eastern end of Flores Island, while the remainder of the coastline posted run-up heights of only 5–8 m [15]. Underwater slumping has also been recognized as responsible for local tsunamis, which even if spatially constrained, can still be hazardous, such as the event on 16 October 1979 near Nice, France (three fatalities).

Such observations give rise to shivering thoughts when considering what must have been the grand-daddy of all underwater slides: the gigantic slumps recently documented along the flanks of the Hawaiian Ridge, which took place during the past few million years, and involved as much as 5000 km^3 of displaced material [16]. While no precise modelling of such events is available, run-up heights of up to 325 m have been suggested for some Hawaii slumps, based on gravel deposits on Lanai Island [17]. It is clear that similar landslides may have caused the eradication of civilizations on oceanic islands, even at teleseismic distances; their tsunamis must have been catastrophic on a scale beyond anything documented in the historical record. Despite a very low rate of recurrence, they will eventually occur in the geological future, as a part of the slow process of birth, life, and death of oceanic hotspot islands, and some level of hazard evaluation and mitigation for these mega-tsunamis is clearly warranted.

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