T-phase Stations for the International Monitoring System of the Comprehensive Nuclear-Test Ban Treaty: A Global Perspective

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INTRODUCTION

The enforcement of the Comprehensive Nuclear-Test Ban Treaty (CTBT) requires the monitoring of acoustic disturbances in the world's oceans in order to guard against the possibility of a nuclear test being hidden either in the water column itself or on a remote island. The traditional means of acoustic monitoring in the water column builds on the experience of the SOSUS project (*e.g.*, Mikhalevsky, 1999), and consists of deploying hydrophones within the SOFAR channel, the layer of low acoustic velocity extending between depths of approximately 600 to 1800 m in the world's oceans, and functioning as a waveguide for acoustic waves with frequencies greater than 2.5 Hz.

While such stations offer superb recording capabilities, notably at high frequencies, they must be tethered to a nearby island or continental shore for the purpose of providing power and retrieving data, which makes for technologically complex, and hence expensive, installation and maintenance procedures (*e.g.*, Schrodt, 1999).

On the other hand, the acoustic energy in *T* waves is readily converted into seismic energy upon hitting the shore of an island or continent (I will show that *T* waves were indeed first identified on seismic records); in addition, under exceptionally favorable circumstances, *T* phases have actually been felt by island populations at distances of several thousand kilometers from both natural sources (Talandier and Okal, 1979) and man-made ones (the underwater nuclear test WIGWAM on 14 May 1955 was felt as far as the Bonin Islands [Wadati, 1960]). The concept of a so-called "*T*-phase station" is simply that of recording seismically the elastic energy resulting from the conversion of the acoustic wave at the continental or island shore; issues such as powering, data storage, and retrieval then become relatively straightforward, resulting in considerable savings during installation, operation, and maintenance.

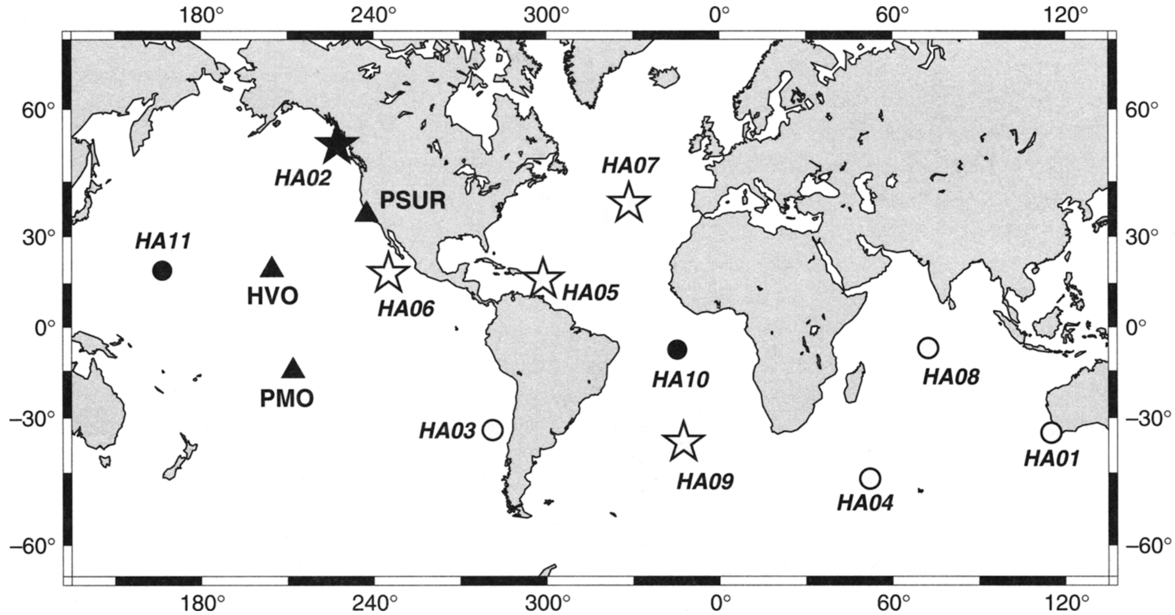
Thus, the CTBT has mandated, as part of its International Monitoring system (IMS), the deployment of five *T*-phase stations, complementing the network of six hydrophone stations independently planned as part of the IMS (Figure 1). The only essential difference between a *T*-phase station and a regular seismic station is a higher sampling rate (mandated 50 Hz; recommended 100 Hz), although the GEOFON seismic network already operates routinely at 80 Hz (Hanka *et al.*, 1995).

In this general framework, this paper first has a historical review of the development of the observation of *T* phases, and then discusses their contribution to several sectors of marine geophysics, interpreted in the context of the monitoring of the CTBT. Next I use the example of the Polynesian Seismic Network to describe the evolution of the concept of a *T*-phase station; and lastly I discuss present research directions of relevance to the operational use of *T*-phase stations in the context of the IMS.

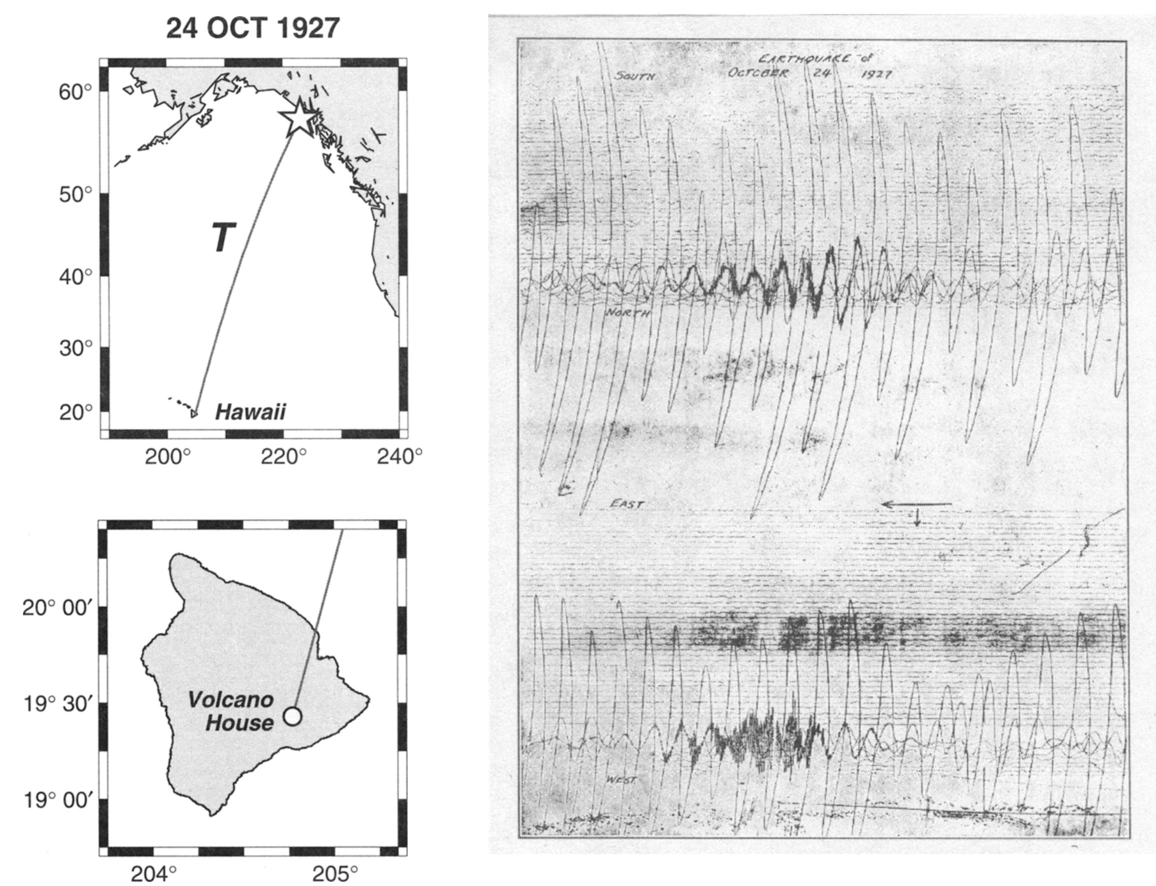
EARLY OBSERVATIONS OF *T* PHASES: A HISTORICAL PERSPECTIVE

The first published description of a *T* phase is, to our knowledge, the seismogram reproduced in Figure 2, recorded in Hawaii from the large earthquake of 24 October 1927 in the Alaskan panhandle (M_{PAS} = 7.1) and published by Jaggar (1930) in *The Volcano Letter*. Jaggar mistakenly, but understandably, described the *T* waves as "made by a local Hawaiian earthquake, which was possibly 'touched off' by the big earth waves." However, a modern interpretation of the wavetrain (arrival time, duration, frequency, and spindlelike shape) positively identifies it as the *T* phase from the Alaskan earthquake. What makes this record remarkable in the present context is that the Bosch-Omori seismometer, a mechanical instrument with a period of 7 seconds, deployed at the Volcano House, 40 km from the estimated conversion slope, was indeed functioning like a *T*-phase station.

Ten years later, *T* phases were recognized as teleseismic in nature in two independent studies. Linehan (1940) reported high-frequency wavetrains at the Weston, Massachusetts observatory following earthquakes in the Caribbean, and Ravet (1940) described a high-frequency (3–10 Hz) arrival in Tahiti following Aleutian earthquakes. Apparently, the label *T* (for tertiary) had been coined as early as 1935 (Leet *et al.*, 1951), but Linehan's report describes the nature of the waves as "unidentified." On the other hand, Ravet correctly interpreted his so-called "very short period waves" as surficial in nature and measured their group veloc-



▲ **Figure 1.** Map of the hydroacoustic network of the IMS, including hydrophone stations (circles) and *T*-phase stations (stars). Solid symbols indicate operational sites, open ones are planned. Stations are identified by their IMS codes. The triangles indicate other sites, not part of the IMS, described in the text.



▲ **Figure 2.** *T* phase observed at Volcano House, Hawaii, following the Alaskan earthquake of 14 October 1927. The maps at left show the geometry of the great circle from epicenter to receiver. The exact conversion of the *T* phase at the Hawaiian shore and its propagation inside the island are not modeled. The record at right is reproduced from Jaggar (1930) and is believed to be the first published seismogram of a *T* phase (top: north-south component; bottom: east-west component); the record is approximately three minutes long.

ity at 1.5 km/s, but, surprisingly, stopped short of interpreting this result as the speed of sound in water.

Considerable progress was made during WWII toward understanding the propagation of sound in the ocean, in particular with the discovery of the existence of the SOFAR low-velocity channel. These results were first summarized after the war by Ewing *et al.* (1946) and expanded from both a theoretical and an experimental standpoint by Ewing and Worzel (1948), Pekeris (1948), and Worzel and Ewing (1948). A model for the reverberation of sound in the channel was proposed independently by Brekhovskikh (1949), following the experiments of Rozenberg (1949).

The first comprehensive description of T phases generated by distant earthquakes is found in the work of Tolstoy and Ewing (1950), who studied mainly records of Caribbean earthquakes at Fordham, Weston, and Ottawa. The paper not only establishes the basic framework of the acoustic-to-seismic conversion, including the possibility of $T \rightarrow P$ and $T \rightarrow S$ conversions at multiple points along the shoreline, but also reports detectable T phases as far inland as Shawinigan Falls, Québec, a distance of at least 700 km from the conversion line.

The interpretation of T waves as propagating in the water column was, however, not without dissent, with a strong rebuke to Tolstoy and Ewing's model presented by Leet *et al.* (1951), who favored interpreting T as shear modes of a sedimentary layer at the bottom of the ocean, an idea already expressed by Coulomb and Molard (1949). Unfortunately, and due to both the large width of the continental shelf and its irregular geometry, the coast of New England and the Maritime Provinces does not lend itself to a simple interpretation of T phases, even at stations located close to shore such as Fordham, and to a much lesser extent Harvard and Weston. Of course, from the large-scale program of hydrophone monitoring in the 1960's (*e.g.*, Johnson *et al.*, 1963) we now know that T waves are indeed sound waves in the water column, but Leet *et al.*'s suggestions remain puzzling. In this framework and because sedimentary layering can give rise to Scholte modes contributing to a wide range of seismic frequencies (Okal and Talandier, 1981; Nolet and Dorman, 1996), possibly within the bandwidth of T-phase seismic records, a critical reinterpretation of Leet *et al.*'s (1951) observations may be warranted in the context of the CTBT.

CONTRIBUTION OF *T* PHASES TO ADVANCES IN GEOPHYSICS

Over the past 50 years, T-wave records have been used in a large number of investigations in the general field of marine geophysics. Among the results most relevant to the CTBT one can cite:

Detection of Volcanism

The possibility of using T waves to detect submarine volcanism is mentioned by Ewing *et al.* (1946), albeit without examples, most probably because these authors' wartime investigations focused on the Atlantic Ocean. The first reported detection of underwater volcanism by hydroacoustic waves is found in Dietz and Sheehy (1954), who describe the eruption of Myojin, approximately 400 km south of Japan, on 18 September 1952, as recorded at Point Sur. This very shallow volcanism led to the repeated emergence of a small island, which eventually subsided. Dietz and Sheehy provided a detailed compilation of the various eruptive phases of the volcano, including the probable timing of the sinking of the Japanese Coast Guard vessel *Kaiyo Maru No. 5*.

As reviewed for example by Talandier and Okal (1987), several decades of systematic monitoring of acoustic waves, notably in the Pacific, have led to the identification and investigation of numerous volcanic sources. Among those, the most spectacular and seminal observation was undoubtedly the discovery of Macdonald Volcano, the then-missing end member of the Austral Island chain, following its eruption on 29 May 1967 (Norris and Johnson, 1969; Johnson, 1970). Macdonald was continuously monitored by the French Polynesian network during its eruptive crises of the 1970's and 1980's, allowing in particular for the identification of the characteristics of volcanic explosions during eruptive sequences (Talandier and Okal, 1982, 1984; Talandier *et al.*, 1988). Such studies may find renewed importance in the context of the CTBT in view of the explosive nature of the associated volcanism.

In a more recent study (Talandier and Okal, 1996), T phases of exceptionally monochromatic character were used to identify a powerful volcanic source in the vicinity of the Eltanin fracture zone at 54°S 140°W. Subsequent shipboard exploration of the area (Géli *et al.*, 1997; Vlastelic *et al.*, 1998) mapped the Hollister Ridge as a massive extrusion of basaltic volcanics, 450 km long by 20 km in width at its base and reaching only 135 m below sea level, and determined that its morphology and its isotopic and trace element signatures have no known equivalent in the Pacific. Talandier and Okal (1996) suggested that the monochromatic spectrum of the Hollister Ridge T waves could express resonance of a water column populated with gaseous bubbles, resulting from geyserlike activity during the volcanic episode, made possible by the exceptionally shallow depth of the eruption. In this context, the analysis of teleseismic T waves resulted in the eventual discovery of both a volcanic ridge of unsuspected size and characteristics and a new mechanism of generation of monochromatic sound in the ocean.

Source Tomography of Major Earthquakes in Abyssal Plains

T waves received at teleseismic distance from a large oceanic earthquake can be used to locate the various components of its rupture precisely, and, because of their ability to propagate high-frequency seismic energy efficiently, to identify the frequency spectrum and hence the stress drop of the various subevents and/or aftershocks. This procedure was used by Reymond *et al.* (1998) in the case of the 1998 Balleny earthquake and may prove important in the context of the CTBT by furthering our understanding of the conditions under which various parts of the hydroacoustic spectrum can be excited by underwater earthquake sources, in particular at high frequencies.

Structure of Subducting Slabs

In a recent study, Okal and Talandier (1997, 1998) noted that the arrival times of the *T* waves generated by the great 1994 deep Bolivian earthquake and propagated throughout the Pacific Ocean could be properly modeled only as involving an $S \rightarrow T$ conversion at the South American shoreline. In turn, the ability of the *S* wave to deliver high-frequency energy ($f > 4$ Hz) to the oceanic column requires propagation through a medium with low anelastic attenuation or in practice through a thermally and mechanically continuous slab. Okal (2000) later extended this approach to deep earthquakes reported as "detached" worldwide. This work emphasizes the locally important role played by *S* waves as vectors of converted *T* energy when an adequate low-attenuation structure is present.

T Waves in the Context of Tsunami Generation

Having noticed that several tsunamigenic earthquakes had generated strong *T* wavetrains, Ewing *et al.* (1950) proposed to use *T* phases for tsunami warning, their rationale being that both tsunamis and *T* waves could be generated only by very shallow earthquakes. Unfortunately, the correlation is far from perfect, as noted early on by Leet (1951) and Wadati and Inouye (1953). In addition, Ewing *et al.*'s remark on *T*-wave excitation is inaccurate (*e.g.*, Northrop, 1974), and indeed some of the very first observations of *T* phases involved earthquakes of intermediate depth (Linehan, 1940; Shurbet, 1955). We now understand that the ability of an earthquake to send high-frequency energy into the SOFAR channel can be a complex function of the actual bathymetry of the ocean floor (Talandier and Okal, 1998; de Groot-Hedlin and Orcutt, 1999). In addition, the excitation of *T* waves and tsunami modes by an earthquake source involves opposite ends of its spectrum; slow events deficient in high frequencies (*e.g.*, the noted "tsunami earthquakes" of 20 October 1963 in the Kuriles [Talandier, 1966] and 2 September 1992 in Nicaragua) had weak *T* waves despite enhanced tsunamigenesis. In Okal and Talandier (1986) we have proposed that the duration of the *T* phase might be better correlated to seismic moment, and hence to tsunami excitation, than the amplitude. Later, Walker *et al.* (1992) and Walker and Bernard (1993) have suggested some level of correlation between *T*-wave strength and tsunami excitation based on a qualitative data set which unfortunately did not include the principal tsunamigenic events of the past decades.

A major development regarding *T* waves in the context of tsunamigenesis came with the devastating Papua New Guinea tsunami of 17 July 1998 (Synolakis *et al.*, 1998), whose parent earthquake was neither large nor slow. A growing consensus now favors interpreting this tsunami as resulting from the failure of a 4 km^3 underwater slump, taking place thirteen minutes after the seismic main shock (Tappin *et al.*, 1999). Critical to this model is the observation, on the Wake Island hydrophone, of a *T* wavetrain of prolonged duration and complex spectrum, which could not have been generated by a bona fide seismic dislocation at the magnitude ($m_b = 4.4$) suggested from conventional seismic waves (Okal, 1999) but rather emanated from a slump that also generated the tsunami. In addition to sensitizing the community to tsunami hazard from slumps, this event underscores slumps as generators of *T*-wave energy and emphasizes the need to investigate in depth the characteristics of their *T* phases, an issue of obvious relevance to the monitoring of the CTBT.

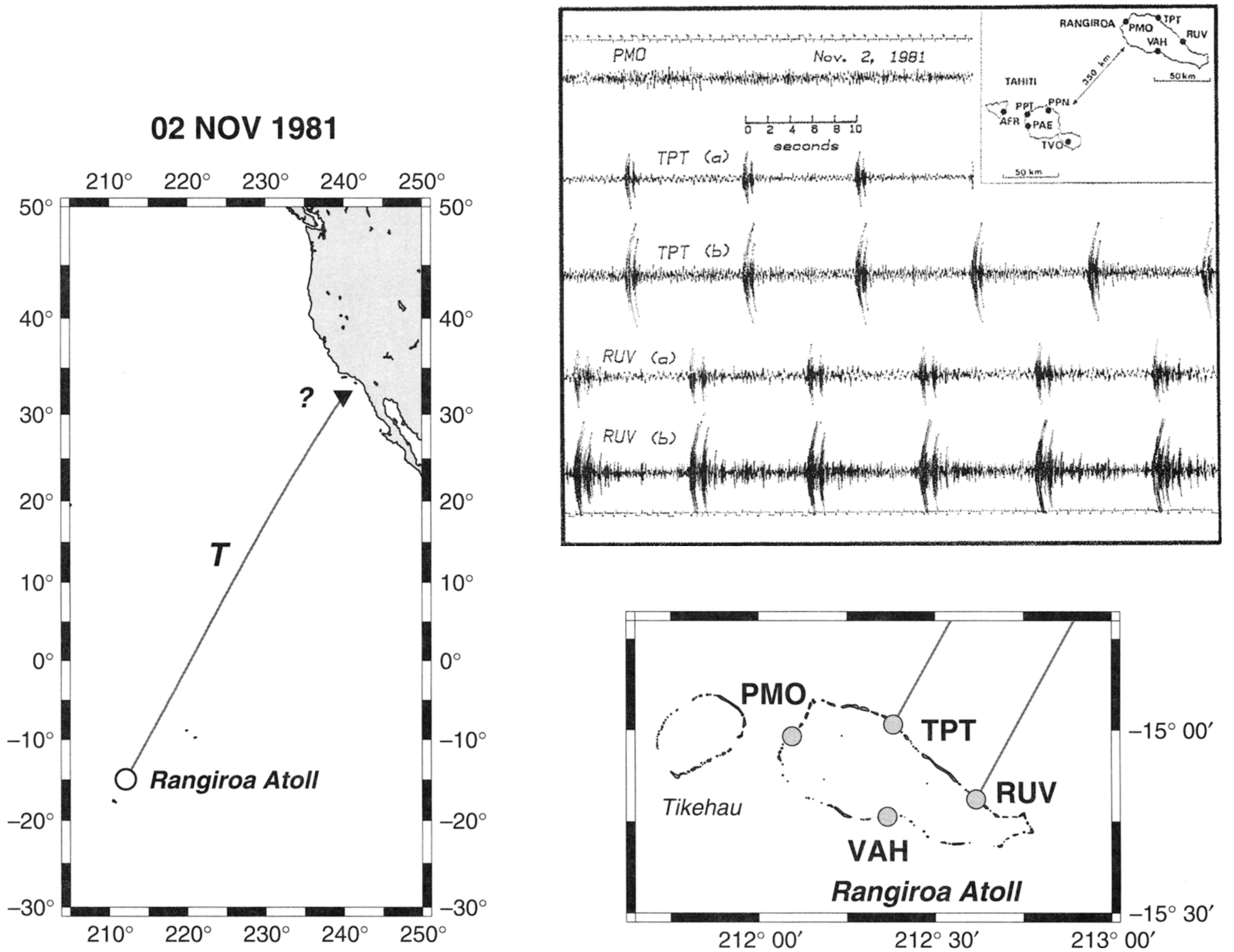
Global Climate Change

Finally, we mention for reference the projected use of *T* waves for the Acoustic Thermometry of Ocean Climate (ATOC), with the goal of directly investigating any possible temporal evolution of the thermocline of the world's oceans, in the context of global climate change. The Heard Island Feasibility Test, carried out in 1991, has demonstrated the possibility of recording essentially worldwide a source emitting an acoustic energy of 3 kW at a frequency of 57 Hz (Munk *et al.*, 1994). The ATOC program could be of great importance to the long-term monitoring of the CTBT if it were to reveal significant temporal changes in sound velocities, which could affect the precise location of acoustic sources in remote areas of the Earth's oceans.

The common denominator of many of the above examples is the ability for *T* waves to detect extremely small sources at extremely large distances. An additional example is given in Figure 3, which illustrates the routine recording of a seismic reflection campaign off the coast of California using seismic stations TPT and RUV on the northern coast of Rangiroa Atoll in French Polynesia (Okal and Talandier, 1986), at a distance close to 6,000 km. The stage is then set for the systematic use of seismic stations recording *T* phases in order to improve detection capabilities greatly in the marine environment: This is the concept of the *T*-phase station.

T-WAVE CHANNELS AT POLYNESIAN SHORT-PERIOD STATIONS: THE PROTOTYPE *T*-PHASE STATIONS

Both the early work of Tolstoy and Ewing (1950), and more recent studies such as Cansi and Béthoux (1985), Pasyanos and Romanowicz (1997), and Stevens *et al.* (1999), have suggested that once converted into seismic energy, *T* phases can, under favorable conditions, propagate several hundred kilometers inside a continental structure. In practice, however, this propagation suffers anelastic attenuation, and for that reason *T*-phase stations are best deployed in the vicinity of shorelines. We refer to Figure 1 for a display of the locations mandated for *T*-phase stations, in complement to their

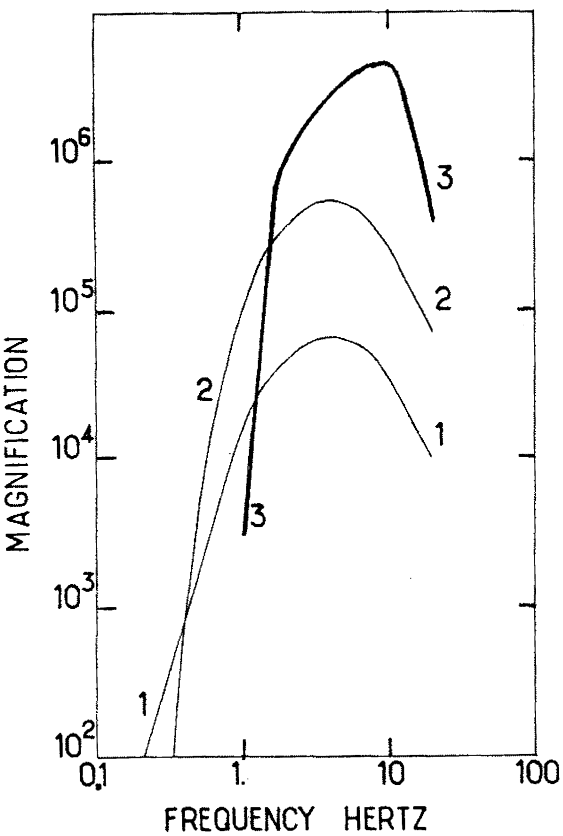


▲ **Figure 3.** *T* waves recorded at Rangiroa Atoll, French Polynesia during a seismic reflection campaign off the coast of southern California on 2 November 1981. An estimate of the source is given based on records throughout the Polynesian network, but its exact location is unknown. Shots were fired every 10 seconds. The box at top right shows seismograms at PMO, TPT, and RUV. Note that PMO is masked by the structure of the atoll itself. For the other two stations, traces (A) are high-gain seismic channels (see Figure 4), and traces (B) are the so-called *T*-wave channels. The map at left gives the general source-receiver layout, and the close-up at bottom right is a detailed map of the Rangiroa subarray of the RSP.

hydroacoustic counterparts, in a configuration which allows transmission along unblocked paths to a minimum of two receivers from 90% of non-Arctic oceanic areas in the world (Mikhalevsky, 1999).

The eventual concept of a *T*-phase station evolved over the past decades from advances in observational seismology at oceanic island sites, in particular in Polynesia. Starting in the 1960's, and as part of the development of the Polynesian seismic network (Réseau Sismique Polynésien, hereafter RSP), special efforts were undertaken to improve the detection and recording of *T* phases on seismic stations of the RSP in essentially two directions: the optimization of the siting of the station and the developments of appropriate filters allowing greater magnification of the *T*-phase displacement field. The latter were essentially narrow band-pass units eliminating the signal at frequencies lower than 1.5 Hz and allowing routine magnifications of 2×10^6 at 3 Hz and 5×10^6 at 10 Hz on the so-called "*T*-wave channels" (Figure 4). The rapid fall-off at low frequencies is made necessary by the presence in the marine environment of harmonics of the sea swell at 0.17 and 0.33 Hz, which are well known to plague the detection capabilities of standard short-period seismic stations to the extent that many of the short-period channels at analog WWSSN oceanic stations had to operate routinely at such mediocre magnifications as 6,250 or even 3,125 at 1 Hz. While nowadays modern broadband digital recording alleviates the need for hardwired filters upstream of recording, the study and interpretation of *T*-phase signals will have to take place in appropriate frequency ranges where *T* energy is abundant, which implies the use of essentially the same filters in the processing software, so that the experience gained from the Polynesian *T*-wave channels remains very precious.

Regarding the siting of the stations, it became rapidly apparent that the best detection capabilities were afforded by stations located on coral atolls (*e.g.*, Rangiroa in the Tuamotu Islands) as opposed to volcanic "high" islands (*e.g.*,



▲ **Figure 4.** Magnification curves for analog channels of the RSP. 1: Standard, low-gain short-period channel; 2: High-gain short-period channel with high-pass filter ($f \geq 0.3$ Hz) rejecting sea swell; 3 (thick curve): "*T*-wave channel" band-passed between 1 and 10 Hz, with typical magnification of 2×10^6 at 3 Hz. After Talandier and Kuster (1976).

Tahiti in the Society group). There are several reasons behind this observation. First, a major advantage of deploying a station on an atoll is the excellent basement coupling provided by the mechanically strong coral platter in the immediate vicinity of the shores on which the station will, per force, be installed (the coral ring on a typical atoll is only about 100 m wide). This is in contrast to the case of high islands, for which erosion of the volcanic structure results in occasionally thick layers of detritic material, responsible for significant high-frequency background noise. This more than offsets the disadvantage of the forced proximity to the ocean shore, which could result in enhanced noise from sea swell; we note in particular that, in the context of *T* waves, sea swell is a low-frequency noise with wavelengths greater than 10 km, which does not attenuate significantly over the size of an oceanic island, whatever its nature, and which can, at any rate, be eliminated through the use of high-pass filters (see above).

Additionally, and as discussed more in detail in Section 5, the underwater morphology of atoll structures generally features much steeper slopes than those of high islands, favoring the conversion of acoustic energy into seismic waves, thus enhancing the performance of an atoll site as a *T*-phase station.

Present-day *T*-phase Stations

Of the five *T*-phase stations mandated by the CTBT, only one is presently operational: Van Inlet, British Columbia, Canada (HA02; see Figure 1), which is also the only one sited at a continental margin (on the Queen Charlotte Islands) rather than on a small oceanic island. A comprehensive study by McCormack and Woodgold (1999) compared the performance of VIB with that of other seismic stations in the Queen Charlotte Islands along a 150 km section of coast and up to 30 km inland. It concluded that detection characteristics (expressed by signal-to-noise ratios obtained during recording of actual teleseismic events) did not differ significantly at the various sites, except for very high frequencies (>20 Hz), in the immediate vicinity of the shoreline. This result may express a generally homogeneous geological structure along the continental margin where the conversion is expected to take place.

Studies carried out at the other sites mandated as *T*-phase stations have included preliminary experimental setups in order to optimize the exact location of deployment. For example, in the case of station HA05b at Martinique, Piserchia and Rodrigues (1999) have shown that the amplitude of the signal can be strongly affected by moving the receiver as little as 400 m inland. As discussed more in detail in Section 5, this is most probably due to conversion of acoustic energy into surface waves, which can be strongly attenuated depending on the structure of the island.

SCIENTIFIC RESEARCH IN THE CONTEXT OF *T*-PHASE STATIONS

The operational use of *T*-phase stations for the purpose of monitoring the CTBT requires a thorough understanding of the processes of generation of acoustic energy in the water and of its conversion into one or more seismic waves propagating in the solid Earth to the receiver. We review here advances in two areas of research critical to this understanding.

The Mechanism of Conversion at the Shore

The history of investigation of the mechanism of conversion of acoustic energy to (and from) seismic energy goes back to the remark by Linehan (1940) that the *T* phases at Weston featured "a large transverse component." Later, Tolstoy and Ewing (1950) confirmed complex ground motion, observed on all three components, for the *T* phases they studied and suggested that the recorded *T* phases had to be combinations of $T \rightarrow P$ and $T \rightarrow S$ conversions, unresolvable in time on account of multiple conversion points at the continental shelf. Båth and Shahidi (1971) noticed that *T* phases from the Arctic midoceanic ridge could be observed as far south as Uppsala after having crossed the Scandinavian shield as P_g, S_g, or R_g wavetrains, taking advantage of a high crustal Q, estimated at 700. These authors also studied theoretically the reflection and conversion coefficients of *T* waves into *P* and *S* at the shelf and in particular explained the partitioning of shear waves into *SV* and *SH* components.

Later, Cansi and Béthoux (1985) used synthetic seismograms to study the conversion of acoustic energy at several points along the French Atlantic continental shelf and successfully identified the $T \rightarrow P$ and $T \rightarrow S$ wave packets. They concluded that $T \rightarrow S$ conversions could become trapped inside the crust as L_g wavetrains, whose subsequent propagation can benefit from regionally high Q values.

Talandier and Okal (1998) used simple ray theory to investigate the role of the morphology of the conversion slope on seismic-to-acoustic conversions on the source side. Their very general conclusion is that steep slopes (with dips of ~45° or more) favor efficient conversion to or from acoustic energy because they allow a direct penetration of the SOFAR channel by the elastic ray with few if any reverberations required. In contrast, a shallow-dipping slope (typically 20° or less) requires repeated reflection of the acoustic ray between the surface and bottom before it can be trapped within the SOFAR channel (Grinda, 1960). This rather inefficient mode of conversion, known as "down-slope conversion" (Johnson *et al.*, 1963), results in weak amplitudes and spindle-shaped waveforms for most earthquake-generated T phases. The importance of steep slopes in converting seismic energy into acoustic T waves was first suggested by Wadati and Inouye (1953).

By virtue of the principle of reverse path of light, these results are easily transferred to the case of the acoustic-to-seismic conversion relevant to the response of T-phase stations. With their steep submarine reefs, atolls will generally provide a very favorable geometry for the conversion of acoustic energy into seismic waves. Furthermore, because of the extremely short distance traveled on land by the seismic wave (often less than even one acoustic wavelength in the 3–10 Hz range), its exact nature (P, S) becomes largely irrelevant, and for all practical purposes a T-phase station built on an atoll can be thought of as recording the acoustic water wave. This is the geometry used for the best performing T-phase stations built in Polynesia, *e.g.*, Pomariorio on Rangiroa Atoll (see Figure 3).

In contrast, high volcanic islands have slopes characteristic of shield volcanoes, dipping at a gentle angle of about 15°. In the case of Tahiti, Talandier and Okal (1998) showed that direct refractions (not involving multiple reverberations which would lead to diminished, emergent wavetrains) are impossible under this geometry. Using the hybrid numerical modeling developed by Piserchia *et al.* (1998) (which consists of combining Maslov summation for propagation on the high seas with finite differences calculations for wavefields in the vicinity of the shoreline and beyond), Piserchia and Rodrigues (1999) have shown that under such conditions the acoustic energy is converted into a surface wave, a conclusion also reached theoretically by Stevens *et al.* (1999) and observed in northern California by Pasyanos and Romanowicz (1997).

However, the situation at many high islands is made more complex by the frequent presence, at depths shallower than the axis of the SOFAR channel, of a coral reef which can provide a very limited range of localized steep slopes, allowing the conversion of T waves into P energy after only one reflection at the surface. This geometry is common in the Society Islands, and a similar one is found on the big island of Hawaii; although the latter has no coral reef, extended small-scale bathymetry (Smith, 1994) has revealed the presence of steep fronts of underwater basaltic flows, reminiscent of the aerial "palis", and whose slope can approach 50°. In such geometries, the $T \rightarrow P$ conversion results in a dipping P ray, which can emerge at the surface only after being refracted by a faster, deeper structure, thus creating a shadow zone for $T \rightarrow P$ conversions at the surface. In the geometry of Tahiti, Talandier and Okal (1998) established a width of 9 km for the shadow zone. Okal (2001) has verified these results on Hawaii by using T phases recorded across the Hawaii Volcano Observatory network from French nuclear tests at Mururoa. The exact extent of the shadow zone for $T \rightarrow P$ conversions, where the bulk of the T phase then corresponds to a $T \rightarrow S$ wavetrain, was found to depend critically on the morphology of the coastal area.

In general, once converted into a seismic wave, the fate of a T phase will be that of a regional seismic phase, hinging fundamentally on the local characteristics of anelastic attenuation. Exceptionally long paths, such as observed in Australia (Stevens *et al.*, 1999), Scandinavia (Båth and Shahidi, 1971), or New England and Canada (Tolstoy and Ewing, 1950), result from the combination of low attenuation inside shield structures with the efficient channeling of crustal phases such as R_g or L_g. On the other hand, in the case of the island of Hawaii, the geometry of the volcanic system totally controls the further propagation of the T phase inside the island; on the south flank of Kilauea, the local $Q_\mu \approx 20$ (Koyanagi *et al.*, 1995; Talandier and Okal, 1998) strongly hampers propagation, but those rays able to dive under the East Rift magmatic system can propagate inside Crustal Layer 4 and have been observed across the entire island structure (Okal, 2001).

Once the exact geometry of the conversion is known, arrival times at T-phase stations can be corrected for on-land propagation, according to the various theoretical models described above, and then merged into the data set of arrival times at hydrophone stations to obtain a precise location of the acoustic source, the reliability of such station corrections being of course greater for stations located close to the conversion point.

Source Discrimination Using *T*-phase Records

The deployment of T-phase and hydrophone stations as part of the IMS seeks to enhance detection capabilities in the marine environment, assuming the scenario of an event detected exclusively by such receivers. In this framework, we address here the question of the identification of the source, *i.e.*, of the discrimination between a natural and a man-made event, based entirely on T-phase data.

In the oceanic environment, a considerable amount of experimental and theoretical work has led to a detailed understanding of many factors controlling the source spectrum of an underwater explosion (Cole, 1948; Chapman, 1985). Recent developments in this respect include, for example, the works of Blackman (1999) and Fisk and Jepsen (1999). In very general terms, an explosion would be expected to have a shorter duration than an earthquake, leading to a higher-frequency spectrum. In land-based seismology, this remark forms the backbone of the time-honored methods used for identification of explosions from conventional seismic waves, such as the m_b:M_S discriminant (Marshall and Basham, 1972).

However, the situation in the context of the CTBT is made very complex because of the many scenarios of possible sources, both natural and man-made. During five decades of nuclear testing, diverse source configurations have been used in the oceanic environment: inside the water body of the high seas (WIGWAM, Sheehy and Halley, 1957), inside the closed lagoon of an atoll (HARDTACK, Milne, 1959), underground inside the structure of an atoll (Mururoa, Adams, 1979; McLaughlin, 1997; Okal, 2001), or in the atmosphere, in the immediate vicinity of an atoll structure (DOMINIC, Talandier and Okal, 2001), each of them leading to significantly different characteristics for the resulting T waves. In addition, the case of natural sources, *i.e.*, earthquakes, is itself complex; because such sources are located inside the solid Earth, the characteristics of their T phases will be affected by the seismic portion of their path on the source side to an extent depending on the particular geometry of the conversion. For example, de Groot-Hedlin and Orcutt (1999) have investigated the variation in shape of the wavetrain of T phases when the earthquake source is moved in the vicinity of the Aleutian Trench. Similarly, Talandier and Okal (2001) have shown that intraplate earthquakes located in the vicinity of oceanic islands may benefit from very favorable conversion conditions, leading both to the retention in their spectrum of significant energy at high frequencies and, for small magnitudes, to a relatively short duration of the T phase, with the result that these events could fail any discriminant based on such criteria.

In this context, Talandier and Okal (2001) used a data set of 150 records obtained at Polynesian T-phase stations, mostly on atolls, from thirty man-made and twenty-nine natural sources and proposed a new discriminant

$$D = \log_{10} e_{\text{Max}} - 4.9 \log_{10} \tau_{1/3} + 4.1 \qquad (1)$$

comparing the maximum amplitude of the recorded ground velocity envelope of the T phase, e_{Max} (in μm/s), with its duration, measured at 1/3 of that maximum, $\tau_{1/3}$ (in seconds). Figure 5B demonstrates the computation of e_{Max} and $\tau_{1/3}$ for the example of a chemical explosion with a yield of 1,000 kg detonated in Polynesia as part of the MIDPLATE campaign (Weigel, 1990) and recorded at station PPL on the south coast of the island of Hawaii (estimated distance to conversion slope: 6 km) at a distance of 3,929 km; the original analog record is shown on Figure 5A. Figure 5C examines the performance of the discriminant (1). The dashed line $D = 0$ effectively separates explosions (with positive values of D) from earthquakes (with negative values of D), except for two small intraplate earthquakes with slightly positive values of D (bull's eye symbols). While this discriminant was derived in an empirical way, its foundation remains the comparison of an overall maximum ground velocity reached by the T phase, essentially a high-frequency parameter, with the duration of the wavetrain, essentially a low-frequency one. In this respect, it builds on the concept of duration, long used for the purpose of computing local magnitudes (Bisztricsány, 1958; Lee *et al.*, 1972; Lee and Stewart, 1981).

There remains, however, a class of signals apparently failing Talandier and Okal's (2001) proposed discriminant, namely the explosive events taking place during swarms of activity at underwater volcanoes. Because these phenomena are, essentially, explosions, they are expected to share at least some of the properties of man-made explosions, and their automatic identification remains a challenge. However, because of their occurrence as part of volcanic sequences, they do not take place as isolated events, and this additional criterion can be used for their recognition.

CONCLUSION

T-phase stations, and in particular those deployed as part of the IMS of the CTBT, can provide spectacular detection capabilities in the marine environment at a fraction of the cost of a comparable network of hydrophones. Sustained research on oceanic T phases over the past five decades, both on the experimental and theoretical fronts, has resulted in a thorough understanding of many aspects of the conversion of acoustic energy to and from seismic waves. The principal challenges still lying ahead concern a refinement of the relevant quantitative models, which will be necessary for the ultimate goal of quantifying the yield of an explosive source based on T-phase data, as well as a continuous investigation of the origin of acoustic energy in the deep marine environment, notably regarding abyssal, intraplate earthquakes at low magnitude levels.

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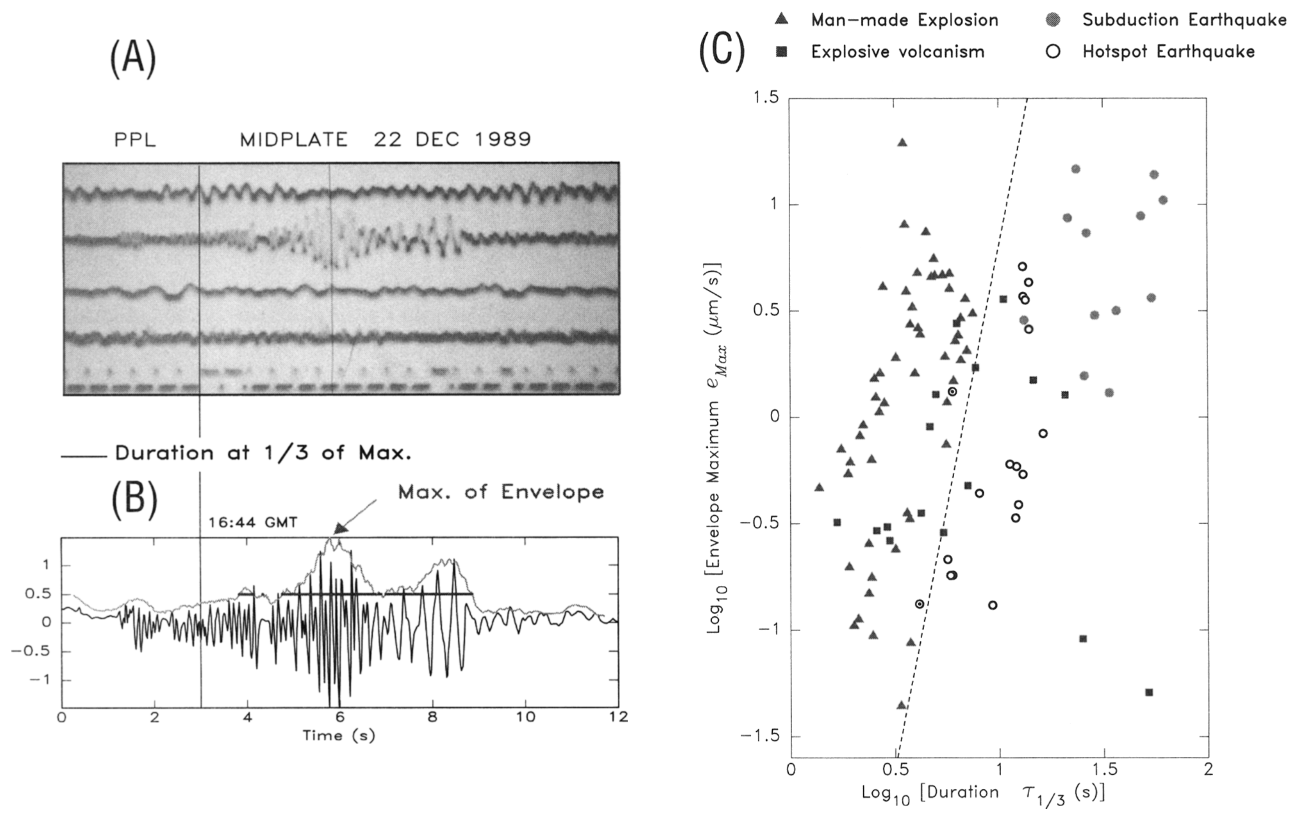

▲ **Figure 5.** Discriminant proposed by Talandier and Okal (2001). (A): Original signal and (B): Method demonstrating the computation of the elements of the discriminant (1); (C): Performance of the discriminant examined on 99 records, from the data set of Talandier and Okal (2001). Various symbols are used for the various sources considered. Note that the dashed line ($D = 0$ in [1]) effectively separates man-made explosions (to the left) from earthquakes (to the right), the latter being either from subduction zones or from intraplate, hotspot regions. Episodes of explosive volcanism (squares) cannot be separated.

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