Hydroacoustic Signals from Presumed CHASE Explosions off Vancouver Island in 1969–1970: A Modern Perspective

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INTRODUCTION

In the period 1966–1970, a number of chemical explosions were conducted off the Pacific coast of the United States, several of them large enough to be located teleseismically by the United States Geological Survey (USGS) and the International Seismological Centre (ISC). At least one (CHASE V) was described as the scuttling of a naval vessel through the disposal of surplus ammunition, with a yield of 1 kt (Northrop, 1968). A second explosion, part of the IITRI program with a yield of 340 tons, is also described in the literature (Northrop and Morrison, 1971). By association with those published events, the other sources are presumed to be similar in nature (Pulli *et al.*, 2000). To our knowledge, experiments on this scale have not been repeated since then.

In the context of the Comprehensive Nuclear Test Ban Treaty (CTBT), these events provide a unique opportunity to study hydroacoustic signals from explosions of a caliber, if not a nature, precisely comparable to the potential sources that the International Monitoring System is being set up to detect. The purpose of the present paper is to analyze seismograms and hydrophone *T* phases recorded routinely at Pacific observatories during the experiments in question, using modern techniques of location and identification of the source. With the exception of the CHASE V and IITRI III shots, this work is performed "blindly", *i.e.*, without the benefit of ground truth. Should the latter become available in the future, it could provide an evaluation of the accuracy of our results, notably with regard to location.

DATA SET

Events

Table 1 lists the events investigated in the present study. The CHASE V explosion (1 kt) took place on 24 May 1966 off the coast of central California and the IITRI III shot (340 t) on 6 September 1968 off Amchitka in the Aleutian Islands. In addition, we considered five explosions detonated off the coasts of Washington State and Vancouver Island (hereafter "the Vancouver shots"). Of them, only three (Events 3, 5, and 6) were located by ISC, on 13 August 1969, 1 October 1969, and 28 May 1970. Figure 1 gives their reported epicentral coordinates. Additional events were detected on 9 September 1969 and 4 September 1970 by the Polynesian Seismic Network (Réseau Sismique Polynésien; hereafter RSP). Of the two announced explosions, full ground truth (epicenter and origin time) is available only for CHASE V (Event 1); Northrop and Morrison (1971) quote a location, but no origin time, for IITRI-III (Event 2). Events 3–7 are reported by Pulli *et al.* (2000) as part of the CHASE series of explosions. The source parameters proposed by these authors are severely incompatible with the USGS and ISC locations for Events 3, 5, and 6, however.

Records

Our primary source of records is the archives of the RSP. This network, previously described in many publications (*e.g.*, Talandier and Kuster, 1976), features special high-frequency channels using band-rejection filters to eliminate noise generated by sea swell, thus allowing magnification in the range of 400,000 at 3 Hz, later upgraded to as much as 2×10^6. When coupled with station locations optimizing acoustic-to-seismic conversion (Talandier and Okal, 1998), this property gives the short-period stations of the RSP exceptional performance as detectors of hydroacoustic signals. They can be regarded as the prototypes of the *T*-phase stations mandated under the CTBT (Okal, 2001). At the time of the main Vancouver shots, recording at the RSP was exclusively on paper charts, thus largely precluding modern data processing of the records. Figure 2 shows an example of RSP records.

In addition, we searched systematically for *T*-phase records at island sites of the WWSSN. Unfortunately, those stations suffered from mediocre operational gains (occasionally as low as 6,250 at 1 Hz) due to the high level of swell-

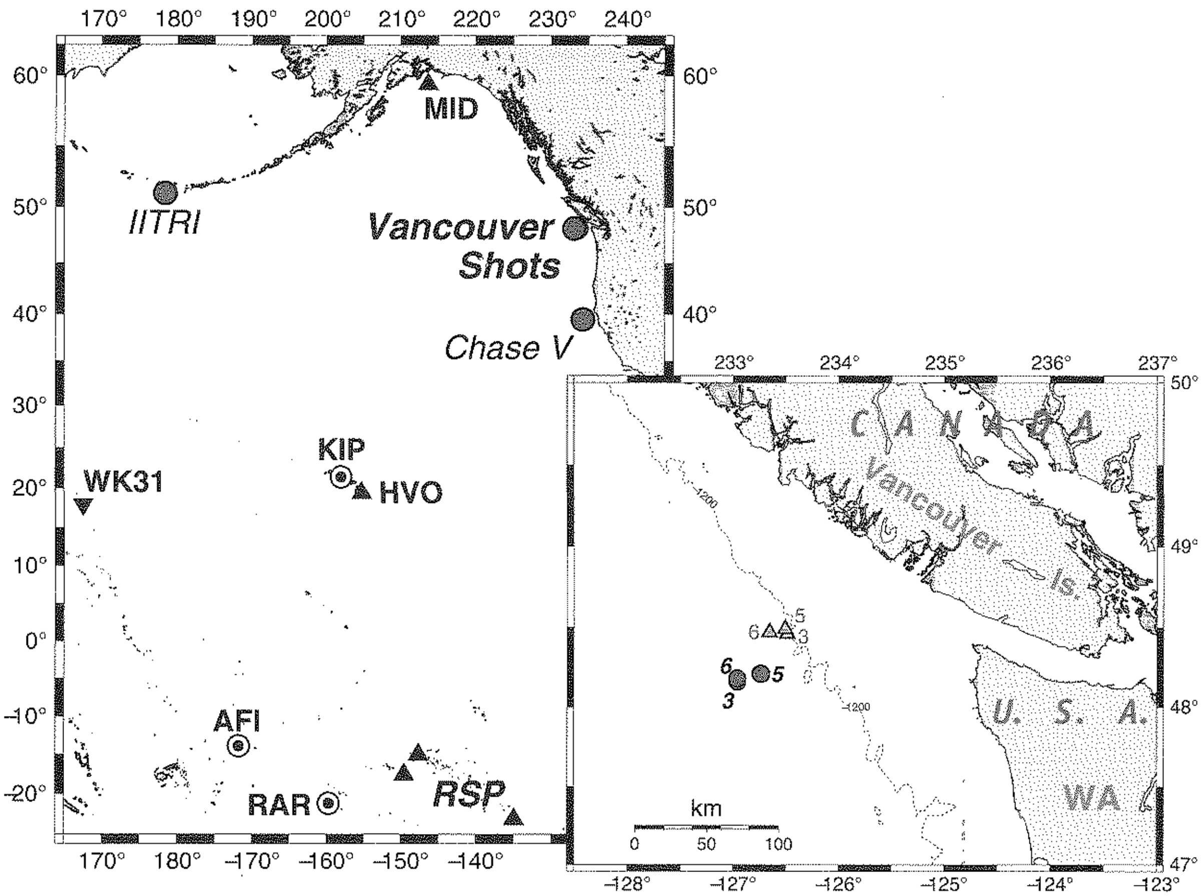
TABLE 1
Source Parameters of Events Investigated in this Study

Event	Name	Date D M (J) Y	Lat. °N	Lon. °E	Origin Time GMT	m_b	e_{max} (μm/s)	$\tau_{1/3}$ (s)	D	Yield (t)	Reference
			Source Parameters								
1	CHASE V	24 MAY (144) 1966	39.45	−125.56	05:49:06	4.7					a
			39.467	−125.80	05:49:06.85					1000	b*
			39.5	−125.5	05:49:06.5	4.9					c
			39.44	**−125.60**	**05:49:06.1**		**159**	**8.8**	**1.67**		**d**
2	IITRI III	06 SEP (250) 1968	51.27	178.37	02:07:13	4.4					a
			51.2	178.4						340	e*
			51.150	178.407	02:07:09.5	4.5					c
			51.212	178.449	02:07:15	4.4					f
			51.060	**178.313**	**02:07:08.7**		**94**	**8**	**1.65**		**d**
3	CHASE XVI	13 AUG (225) 1969	48.46	−126.49	16:12:17.5	4.6					a
			48.483	−126.474	16:12:16.9	4.6					c
			40	−130							g
			48.43	−126.55	16:12:13.4						h
			48.296	**−127.143**	**16:12:10.7**		**170**	**11**	**1.23**	**346**	**d**
4	CHASE XVII	09 SEP (252) 1969	42	−128	21:00:00						g
							40	11	0.60	53	i
5	CHASE XVIII	01 OCT (274) 1969	48.49	−126.51	17:11:11.4	4.9					a
			48.506	−126.485	17:11:11.3	4.7					c
			48.520	−126.492	17:11:12.0	4.9					f
			40	−130	18:00:00						g
			48.49	−126.54	17:11:08.6						h
			48.346	**−126.961**	**17:11:06.4**		**85**	**11**	**0.93**	**141**	**d**
6	CHASE XIX	28 MAY (148) 1970	48.47	−126.66	17:38:32.3	4.9					a
			48.450	−126.659	17:38:32.1	4.9					c
			43	−126	18:00:00						g
			48.44	−126.64	17:38:32.4						h
			48.296	**−127.121**	**17:38:30.4**		**63**	**11**	**0.80**	**95**	**d**
7	CHASE XX	04 SEP (247) 1970	40	−130	21:08:00						g
			48.208	**−127.280**	**21:23:06.2**		**72**	**11**	**0.85**	**113**	**i**

References: a: International Seismological Centre (ISC); b: Northrop (1968); c: USGS Preliminary Determination of Epicenters; d: this study (using both T and P times); e: Northrop and Morrison (1971); f: Engdahl *et al.* (1998); g: Pulli *et al.* (2000); h: this study (using only P times); i: this study (using only T times).

*These references report ground truth location (b, e) and timing (b only).

For each event, the final solution obtained in this study is given in bold.



▲ **Figure 1.** *Left:* Reference map of the positions of sources and receivers considered in this study. Solid dots are locations of the CHASE V, IITRI, and Vancouver shots. Bull's-eye symbols are WWSSN stations where T phases of the Vancouver shots were detected. Middleton Island, RSP, and HVO stations are shown as upward-pointing triangles (to reduce clutter, not all stations used within a network are shown). Downward-pointing triangle is the location of Wake hydrophone WK-31. *Right:* Close-up of the epicentral area of the Vancouver shots. Dashed line is the 1200-m contour taken as representative of the axis of the SOFAR channel. Small triangles are ISC locations of Events 3, 5, 6; larger solid dots are our final locations, obtained by joint relocation of teleseismic P and T arrivals.

generated noise on small islands. Nevertheless, we were able to obtain a few records at the island stations of KIP (Kipapa, Hawaii), RAR (Rarotonga), and AFI (Afiamalu, Western Samoa). Additional records were obtained from a systematic search of the archives of the Hawaiian Volcano Observatory (HVO); a lone but important record from a temporary station at Middleton Island (MID) in the Gulf of Alaska was obtained from the Geophysical Observatory of the University of Alaska. Finally, a number of hydrophone records were obtained from the Hawaii Institute of Geophysics. A sample of records is shown in Figure 3. On the other hand, and while we detected wavetrains interpretable as T phases at nearby continental stations such as Longmire and Corvallis, we were unable to model accurately their conversion to seismic waves at the ocean margin, and thus decided against their use for the purpose of epicentral location.

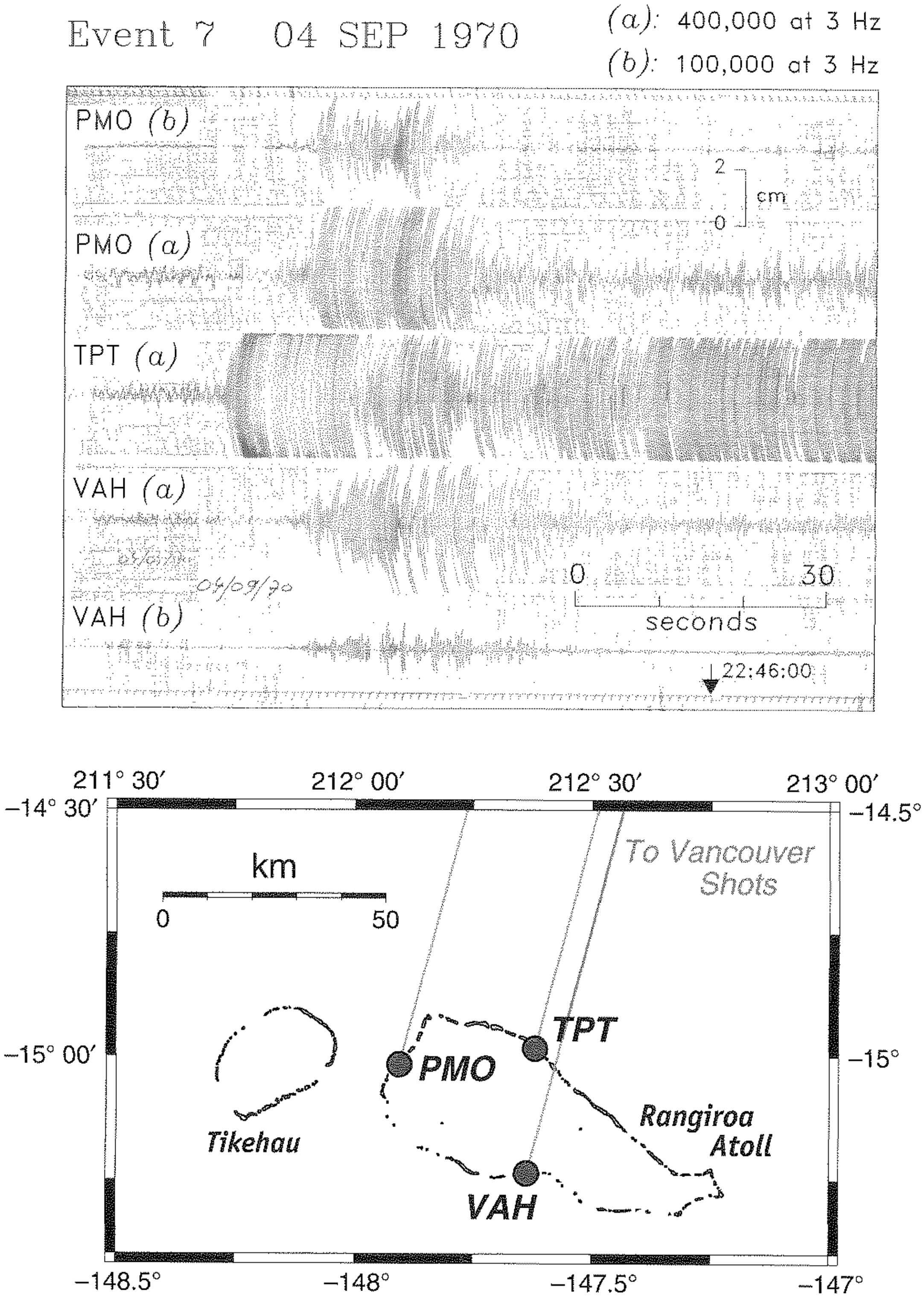
Photographic and smoked paper records were simply photocopied on site. In the absence of printing capability, develocorder records had to be photographed directly from the screen of the few surviving develocorder reader machines.

RELOCATION

Measuring *T*-phase Arrival Times

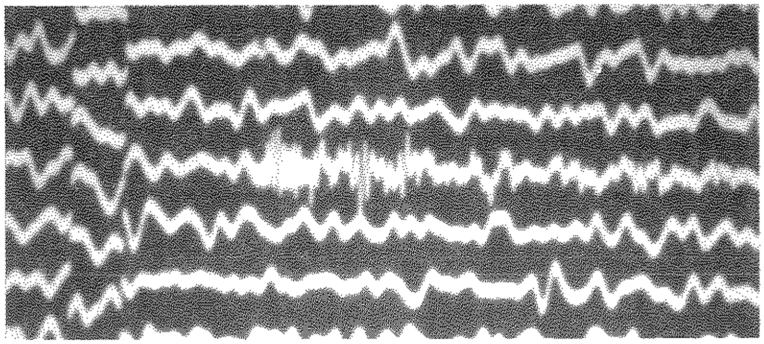
As discussed more in detail in Reymond *et al.* (2003), the measurement of arrival times at T-phase stations involves several steps. On paper records, the arrival time is taken as the onset of the first sequence of large amplitude, characterized by a sharp increase in frequency, relative to background noise. This determination is usually made with a precision on the order of 1 s. Clock corrections were obtained by checking the P-wave arrival times of well located teleseismic sources bracketing the occurrence of the events under study. For example, for Event 5, we used the large Peruvian earthquake of 1 October 1969 (05:05 GMT; m_b 5.9) and the nuclear explosion MILROW (2 October 1969, 22:06 GMT; m_b 6.5); at HLK, the correction reached a staggering 72 seconds.

Further systematic corrections were applied to records obtained on high islands, to offset the effect of postconversion propagation as a seismic wave. This technique was described in detail by Talandier and Okal (1998). The correc-

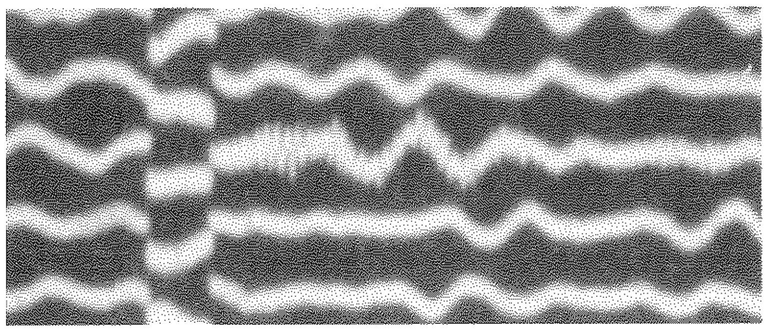


▲ **Figure 2.** Example of *T* phases recorded at the Rangiroa subarray of the RSP (Event 7, 4 September 1970). Top frame shows the original paper records at the three stations then making up the subarray (shown on a map on the bottom frame), as recorded with original gains of 400,000 (channels "(a)") and 100,000 "(b)" at 3 Hz. Note that the record is heavily saturated at station TPT but can be studied meaningfully on both channels at VAH and on the low-gain channel at PMO.

AFI 13-AUG-1969 17:44 (12500)

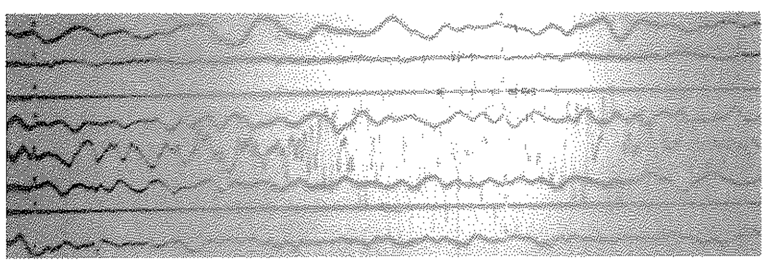


RAR 04-SEP-1970 22:57 (6250)

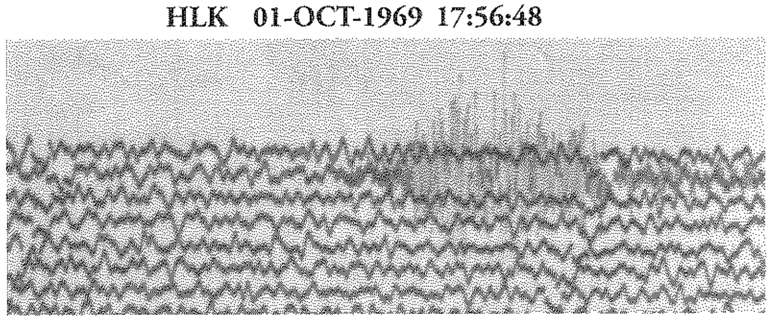


▲ **Figure 3A.** Examples of *T* phases from the Vancouver shots recorded on short-period channels of WWSSN stations. Times refer to the minute marks preceding the *T* phases. The duration of the minute mark is the standard 2 seconds. Note that despite the low gains (original magnifications are given in parentheses), the *T* wavetrains are still detectable.

MID 04-SEP-1970 21:43



HLK 01-OCT-1969 17:56:48



▲ **Figure 3B.** Further examples of *T* phases recorded from Vancouver shots at Middleton Island (*top*) and Haleakala, Maui (*bottom*). Time marks at MID are every 10 seconds, with the full minute (21:43) omitted; the record is 32 seconds long. Time marks at HLK are every minute, with the left one at 17:56:48 after effecting the clock correction derived from the record of the nuclear shot MILROW (see text for details); the record is 83 seconds long.

tion, to be added to the observed arrival time, is always positive. It is unnecessary for stations located on atolls in the immediate vicinity of the conversion point and, obviously, for hydrophones.

A final correction is applied to compensate for the dispersion in group velocity between the frequencies prevailing in our records (typically 5 Hz) and those used to compile the available models of *T*-wave speed in the Pacific Ocean (in the kHz range). This correction to the observed time, which is always negative, was developed empirically from the records of chemical explosions with published ground truth (Taber and Lewis, 1986; Nava *et al.*, 1988; Weigel, 1990) and is detailed in Reymond *et al.* (2003).

Residuals

We compute predicted *T*-wave arrival times based on available ISC epicentral parameters, using the regionalized, seasonally adjusted tables of Levitus *et al.* (1994), and list in Table 2 residuals r defined as observed minus computed arrival times. It is immediately apparent that the residuals for Events 3, 5, and 6 are unacceptably large and negative (from −24 to −34 s on average). This is in contrast to the case of the CHASE V explosion (Event 1), for which we note (see Table 2) that the average value of residuals (−1.6 s) is smaller, in absolute value, than the standard deviation (2.6 s) and thus not statistically significant. In the case of Event 2 (the IITRI-III explosion at Amchitka), the residuals are strongly negative (on the average −18 s).

These results are in sharp contrast with the case of other, documented explosions in essentially the same region, such as the Grays Harbor series (Taber and Lewis, 1986; B. T. R. Lewis, personal communication, 2000), whose ground-truth parameters were used to define the dispersion corrections to the velocity models used in the present study. As such, our observations are intriguing and deserve some comment.

We first consider the robustness of the various seismic epicenters as computed by different agencies, *i.e.*, the USGS (PDE) and ISC Bulletins, Engdahl *et al.*'s (1998) catalog (only Events 2 and 5), as well as our own relocations, based on *P*-wave arrivals published by the ISC, using the method of Wysession *et al.* (1991). We note different trends between the CHASE V event in the south, the Vancouver shots, and the IITRI-III event off Amchitka Island. The seismic epicenters of the Vancouver shots (Events 3, 5, and 6) appear quite robust when computed by different agencies, with the USGS and ISC estimates differing by less than 4 km, and the origin times by at most 0.6 s. Event 5 also appears in Engdahl *et al.*'s (1998) catalog, only 3.6 km away from the ISC epicenter, but its depth estimate, 14 km, is obviously erroneous, the source being an explosion in the water column (see below). As a result, the large *T*-phase residuals are consistent, regardless of the particular seismic epicenter chosen. Similarly, for CHASE V (Event 1), the two available epicenters differ by less than 8 km and 1 s, although the ground truth location is 20 km west of the ISC epicenter. In contrast, in the case of IITRI-III (Event 2), the ISC and USGS estimates differ by 13 km and 5 seconds, the latter probably resulting from a trade-off with hypocentral depth, which is left floating by ISC. The reported ground-truth location is closest to Engdahl *et al.*'s (1998) estimate, about halfway between the ISC and USGS ones. *T*-phase residuals are large and negative for the ISC and Engdahl *et al.* solutions, primarily on account of their late origin times, resulting from the assumption of an exceedingly deep hypocenter (17 and 20 km, respectively).

Finally, we note that source parameters for the CHASE and IITRI series are also listed by Pulli *et al.* (2000). However, for Events 3, 5, and 6, the epicenters are between 610 and 984 km away from the relevant ISC locations, and the origin times of Events 5 and 6 are 48 and 21 minutes late, respectively (no origin time is given for Event 3). Most seriously, the origin time proposed for Event 5 (18:00:00) is noncausal, being more than 3 minutes later than the arrival at HLK shown in Figure 3B. The epicenter estimate itself is unacceptable, as indicated by the standard deviation of the residuals of our corrected *T* times, which range from 60.0 to 80.7 s. As we cannot explain the origins of these discrepancies, we elect hereafter to ignore all epicentral parameters given by Pulli *et al.* (2000), including those pertaining to Events 4 and 7, unreported by ISC. For the latter, which was recorded at Middleton Island, the standard deviation of *T* residuals for these authors' epicenter reached 354.8 s, or close to 6 minutes.

Location from *T* Phases Only

To further investigate the large negative *T*-phase residuals of the Vancouver shots, we attempted to locate the events from *T* phases only, using the propagation model of Levitus *et al.* (1994) and the Monte Carlo algorithm of Wysession *et al.* (1991) to define an uncertainty ellipse under the assumption of Gaussian noise in the data, with $\sigma_G = 1.5$ s. Figure 4 shows the results (labeled "*T*") in the case of Event 3. Essentially similar results are obtained for Events 5 and 6. Because of the very limited azimuthal distribution of receivers, there is poor control on distance, with the semimajor axis of the ellipse extending 95 km in the north-northeast direction. Nevertheless, our results imply that the source of the *T* waves lies at least 40 km to the west of the published ISC epicenter. Note that we could not relocate Event 4 (9 September 1969), which is not detected above noise level at RAR and AFI.

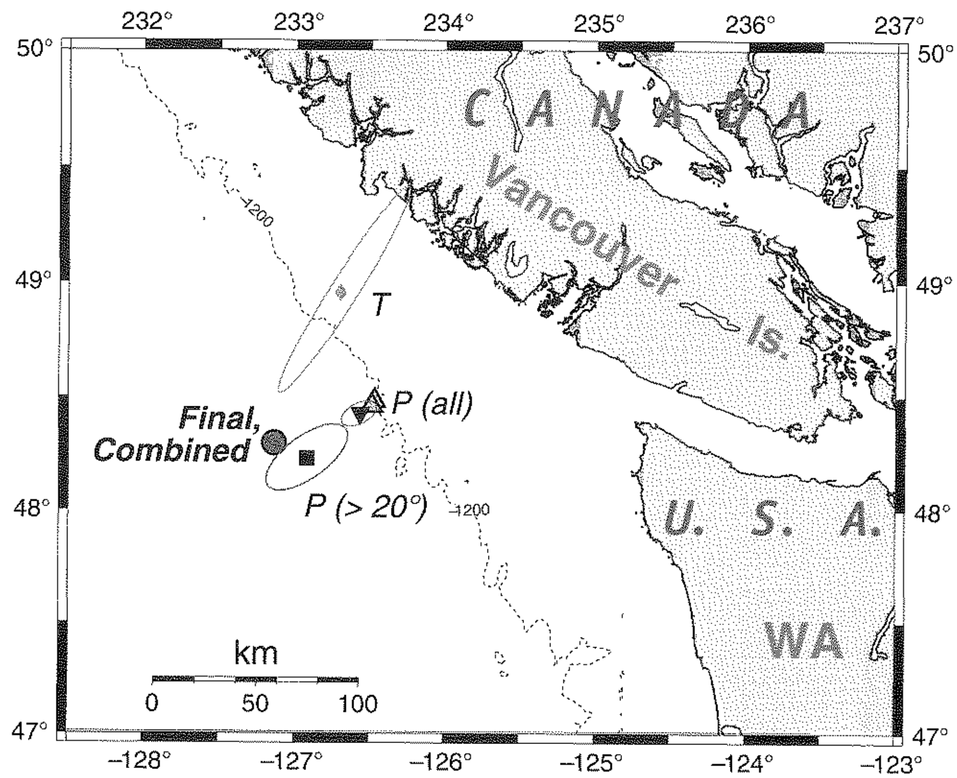
This incompatibility between the body-wave and *T*-phase epicenters is intriguing and its origin can be sought *a priori* in several scenarios: First, the source of the *P* and *T* waves could be physically distinct, both in space and time. We are motivated in this assumption by the detailed analysis of the space-time relationship of the origin of *T* waves generated by earthquakes at continental or island margins (Okal and Talandier, 1997, 1998; Gräber and Piserchia, 2003). Under this scenario, an earthquake located inland generates a *T* phase at the SOFAR interface on the slope, seaward of the shore and hence of the epicenter, resulting in negative *T*-phase residuals.

Given that both teleseismic and acoustic estimates of the epicenter are at sea, a fact supported by the clear explosive

TABLE 2
T-phase travel time residuals computed from ISC epicentral parameters

Station				Arrival Time (corrected)	Epicentral Distance (°)	Residual *r* (s)
Code	Name	Island	Network			
Event 3 — 13 AUG 1969, Average *r*: –34.0 ± 7.5 s						
HLK	Haleakala	Maui	HVO	16:57:20.6	36.51	–42.8
KIP	Kipapa	Oahu	WWSSN	16:57:48.9	36.91	–44.4
TPT	Tiputa	Rangiroa	RSP	17:34:16.0	66.04	–30.0
PPN	Papeeno	Tahiti	RSP	17:37:55.5	68.95	–29.0
RKT	Rikitea	Gambier	RSP	17:41:33.7	71.82	–22.7
AFI	Afiamalu	Upolu	WWSSN	17:44:12.1	74.06	–38.9
RAR	Rarotonga	Rarotonga	WWSSN	17:46:14.8	75.62	–30.3
Event 5 — 01 OCT 1969, Average *r*: –27.3 ± 4.4 s						
HLK	Haleakala	Maui	HVO	17:56:27.1	36.52	–30.8
KIP	Kipapa	Oahu	WWSSN	17:56:53.7	36.91	–34.1
MLO	Mauna Loa	Hawaii	HVO	17:57:15.0	37.15	–29.2
TPT	Tiputa	Rangiroa	RSP	18:33:17.5	66.06	–24.1
PPT	Papeete	Tahiti	RSP	18:37:02.9	69.05	–22.7
TVO	Taravao	Tahiti	RSP	18:37:11.6	69.17	–22.8
Event 6 — 28 MAY 1970, Average *r*: –24.8 ± 6.7 s						
HLK	Haleakala	Maui	HVO	18:23:52.1	36.43	–19.9
KIP	Kipapa	Oahu	WWSSN	18:24:06.1	36.82	–35.7
MLO	Mauna Loa	Hawaii	HVO	18:24:29.5	37.06	–29.0
TPT	Tiputa	Rangiroa	RSP	19:00:36.0	66.00	–22.3
RKT	Rikitea	Gambier	RSP	19:07:53.7	71.81	–17.1
Event 1 (CHASE V) — 24 MAY 1966, Average *r*: –1.6 ± 2.6 s						
HIL	Hilo	Hawaii	HVO	06:29:19.3	32.13	–1.8
HLK	Haleakala	Maui	HVO	06:29:30.1	32.21	2.4
PPT	Papeete	Tahiti	RSP	07:05:22.5	61.14	–2.0
RAR	Rarotonga	Rarotonga	WWSSN	07:14:31.8	68.44	–1.8
AFI	Afiamalu	Upolu	WWSSN	07:14:31.0	68.45	–5.7
Event 2 (IITRI III) — 06 SEP 1968, Average *r*: –17.8 ± 2.2 s						
AFI	Afiamalu	Upolu	WWSSN	03:28:51.7	65.61	–21.0
PMO	Pomariorio	Rangiroa	RSP	03:37:28.0	72.38	–14.8
PPT	Papeete	Tahiti	RSP	03:39:38.6	72.40	–17.5
RAR	Rarotonga	Rarotonga	WWSSN	03:40:26.3	74.83	–17.8

13 AUG 1969 —Event 3



▲ **Figure 4.** Relocation of Event 3. Seismic solutions: Upward-pointing triangles are USGS (back, partly hidden) and ISC (front) epicenters obtained from the full *P*-wave data set. Solid downward-pointing triangle is our relocation from the same data set, using the method of Wysession *et al.* (1991), with corresponding Monte Carlo ellipse (thin trace), computed for $\sigma_G = 1.5$ s. Solid square (with thick Monte Carlo ellipse) is the solution obtained from exclusively teleseismic arrivals ($\Delta > 20°$; note the ~40-km offset to the southwest. Solutions using *T* phases: Diamond is the solution obtained using exclusively the *T*-phase data set listed in Table 2. Note the elongated Monte Carlo ellipse, expressing the lack of distance control, due to the poor azimuthal coverage. Finally, the shaded circle is our preferred solution, resulting from the combined *T* and *P* ($\Delta > 20°$) data set. Note the small error ellipse, contained inside the symbol on the scale of this figure.

nature of the source (see below), we reject this scenario for the Vancouver shots, since it would lead to longer acoustic than seismic paths, and hence to positive residuals *r*. In principle, one could also envision the converse model of a *T* phase emanating at the acoustic epicenter, and generating the seismic (*P*) waves upon hitting the continental shelf, about 40 km to the east and 27 s later. This model is however highly unlikely, since it requires a strong spatial directivity of the original *T* phase, with the source sending little or no energy vertically to the ocean floor only about 2 km below.

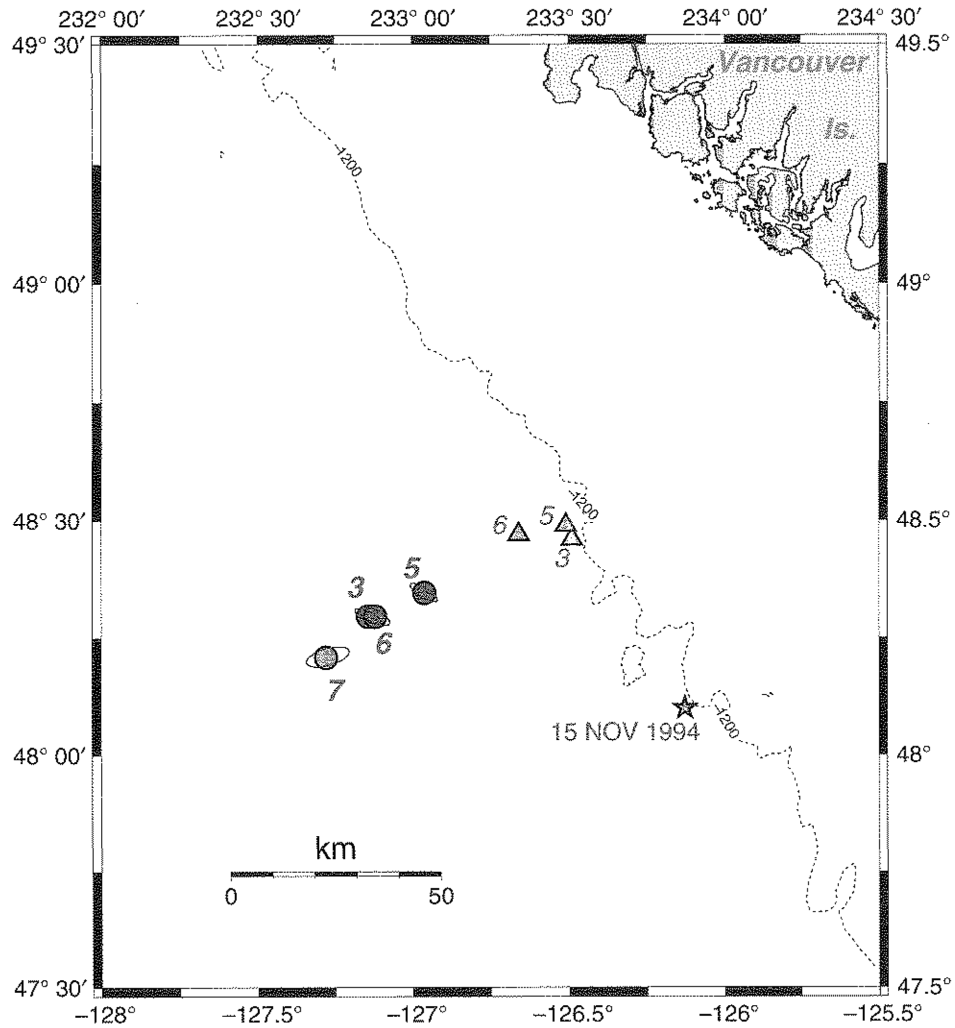
A more likely explanation of the disparity between acoustic and seismic epicenters may simply be a systematic error in the seismic epicentral location. We are motivated in this assumption by the reported systematic bias between epicentral locations proposed by worldwide reporting agencies and regional networks, which, for example, reaches 30 km for the 2001 Nisqually earthquake (Crosson *et al.*, 2001; Villaseñor *et al.*, 2001). Hyndman *et al.* (1978) have further documented a northeastward offset of up to 60 km between the true location of Explorer Ridge earthquakes, as detected by a local network of ocean-bottom seismometers, and their USGS/PDE estimates, for windows of time (1976) and magnitude (m_b 4.4) comparable to those of the Vancouver shots. Similarly, Dziak *et al.* (1991, 2000) have reported a 10- to 35-km bias in location between teleseismic estimates of earthquakes on the Blanco Ridge, and their counterparts obtained from *T* phases recorded on nearby hydrophone arrays.

This observed bias between regional and worldwide epicentral estimates in and around the Pacific Northwest is generally interpreted as expressing the inadequacy of global models to account properly for the propagation of regional *P* waves (principally P_n) crossing the complex structure of the subduction zone (Spence *et al.*, 1985; Zhao *et al.*, 2001).

Combined Relocation

To obtain an improved and, we hope, reliable estimate of the sources of the Vancouver shots, we performed relocations using data sets combining the acoustic and *P* arrival times but the latter only for stations at distances greater than 20°, this threshold being chosen to eliminate regional arrivals significantly affected by lateral heterogeneity. Results, listed in Table 1 and shown in Figure 5, suggest that the shots took place

ALL VANCOUVER EVENTS



▲ **Figure 5.** Map of the final relocations of Events 3, 5, 6, and 7. Triangles are the available ISC locations; solid dots (with Monte Carlo ellipses) are our preferred relocations obtained from combined T and $P (\Delta > 20°)$ data set. In the case of Event 7, no P times are reported, but the inversion uses the Middleton T phase, which results in a confidence ellipse of comparable dimension. The star shows the location of the reference nearby small earthquake used as part of the identification study in Section 4.

within 15 km of a central location at 48.29°N, 127.13°W. We also relocated Events 1 and 2 using the same technique, with somewhat different results: The epicenter of CHASE V is essentially unchanged, while that of IITRI-III is moved south-southwest approximately 12 km from the USGS location.

Event 7, 4 September 1970

This event was not located by any of the agencies, and thus no data set of P-wave arrival times is available. On the other hand, the event took place after the installation of station MID (Middleton Island), resulting in a much improved azimuthal coverage of the T-wave data set. As shown in Figure 5, the uncertainty ellipse for the T-wave epicenter is much smaller, with its semimajor axis (5.5 km) comparable to those of the combined relocations.

EXPLOSIVE NATURE OF THE SOURCES

The nature of the Vancouver sources was investigated by applying Talandier and Okal's (2001) algorithm. It consists of comparing the maximum amplitude e_{max} of the envelope of the ground velocity of the T phase recorded at an atoll station to the duration $\tau_{1/3}$, during which the envelope amplitude remains at least one third of e_{max}. The discriminant

$$D = \log_{10} e_{max} - 4.9 \log_{10} \tau_{1/3} + 4.1 \quad (1)$$

where $\tau_{1/3}$ is in seconds and e_{max} in μm/s, effectively separates explosions ($D > 0$) from earthquakes ($D < 0$). As shown in Figure 2, the T-phase paper record at station TPT, facing northeast on Rangiroa atoll, is heavily saturated, but its amplitude can be extrapolated from other stations on the atoll, as discussed in Talandier and Okal (2001) and Okal *et al.* (2003). For Events 1 and 2, predating the full development of the RSP subarray on Rangiroa, a further extrapolation is necessary between the high island of Tahiti and the atoll of Rangiroa. To estimate $\tau_{1/3}$, we use unclipped records at other stations on Rangiroa and extrapolate our measurements back to TPT, based on our experience with smaller explosions (principally from seismic refraction campaigns [Taber and Lewis, 1986; Nava *et al.*, 1988]), recorded digitally across the whole subarray. Such estimates are probably biased upward, since seismic propagation across the atoll can involve multipathing and result in additional complexity in the waveform and hence increased duration; we use a common value $\tau_{1/3} = 11 \pm 2$ s for Events 3–7. Our results, listed in Table 1, then clearly identify the Vancouver shots as explosions ($D = 0.60$ to 1.23 for Events 3–7; 0.24 to 1.65 accounting for error bars). The announced events 1 and 2 were similarly verified ($D = 1.65$ and 1.67). We also checked that a genuine local earthquake of comparable body-wave magnitude (15 November 1994; M_L 3.0; star in Figure 5) was clearly identified as such ($D = -3.33$).

Finally, we use our estimates of the envelope amplitudes e_{max} of Events 3–7 to propose values of the yields Y of the CHASE explosions. In Talandier and Okal (2001), we used a set of 44 announced underwater explosions at sea recorded on Polynesian atolls to derive the empirical relation

$$\log_{10} Y = 1.30 \log_{10} e_{max} + 2.64 \qquad (2)$$

between e_{max} (in µm/s) and Y (in kg of equivalent TNT). The resulting estimates of Y, listed in the "yield" column of Table 1, range from 53 to 346 tons of equivalent TNT. The similarity in paths and in receiver geometry allows a direct comparison of the relative yields of Events 3–7, which clearly confirms Event 3 as the largest shot (although this would not be apparent from its seismic magnitude) and Event 4 as the smallest (to the extent that it could not be reliably located). On the other hand, at most Pacific stations, the disparity in back-azimuth prevents a direct comparison of the records of Events 2 (IITRI-III at Amchitka) and 3 (Vancouver shot), for which similar yields are either reported or estimated in this study. Given the various approximations involved in their computation, our yield estimates must be regarded as mere order-of-magnitude estimates; a prudent conclusion is that the Vancouver shots (Events 3–7) involved yields in the range of 100 tons, with significant variation between events. This number appears reasonable when compared to the obviously larger CHASE V explosion, announced at 1 kt.

CONCLUSIONS

Our investigations of the old CHASE explosions in the Pacific lead to the following observations.

1. T phases were recorded at island seismic stations throughout the Pacific basin, at spectacular levels under favorable recording characteristics (Rangiroa, Maui), but could be detected above noise level even on stations running at mediocre magnifications (Rarotonga, Afiamalu).
2. Travel-time residuals relative to arrivals predicted from teleseismic locations (*e.g.*, ISC solutions) are exceedingly negative, suggesting that the latter are systematically biased, most probably as a result of the strong lateral heterogeneity affecting the propagation of regional phases, as observed by several independent studies of nearby epicenters, both on land and at sea.
3. Locations based exclusively on T-phase arrivals can reach a precision of ~5 km, given adequate azimuthal coverage at the source, as documented by the Middleton record for Event 7. Otherwise, a comparable precision can be achieved through the inversion of a combined data set of P and T arrival times. The 15-km offset between our locations of Event 7 (T phases only) and Events 3 and 6 (P and T combination) is comparable to the scatter between the latter (3 and 6) and Event 5 (all from combined relocations), thus suggesting that the two relocation methods are not systematically biased with respect to each other, and reaffirming the strong synergy achieved by regrouping data sets of different natures (Talandier *et al.*, 2002). P arrivals at regional distances ($\Delta < 20°$) should not be used, however, as they are strongly affected by lateral structure.

   Incidentally, it may appear as a paradox that a satisfactory relocation of an explosion could be achieved by combining T and uncorrected P times, while it clearly would fail for an earthquake source using P and uncorrected T times. The reason lies in the physical difference between the two source-side hybrid path segments. In the case of the seismic dislocation, the preconversion seismic segment can be tens of kilometers long, and the resulting timing error if interpreted as acoustic can reach tens of seconds; besides, T phases to different stations can have different conversion points along the coastline. In contrast, for an underwater explosion, the velocity contrast at the ocean bottom results in strong refraction, and P waves can be taken as radiating from a single point, namely the projection of the source on the ocean floor, with the uncorrected segment being essentially vertical and always short (at most 5 km); when that segment is interpreted as seismic rather than acoustic, the resulting timing error does not exceed 3 seconds.
4. Even though the paper records at the reference station TPT were heavily clipped, it is possible to estimate their amplitudes by extrapolation from other stations across the atoll. The application of Talandier and Okal's (2001) discriminant then clearly identifies all the Vancouver events as underwater explosions, and the values estimated for the envelopes of the ground velocity at TPT suggest yields in the 100-ton range, with a fluctuation of a factor of 6 between the largest shot (Event 3) and the smallest one (Event 4); these numbers are not unreason-

able when compared to that published for the larger CHASE V explosion (1 kt).

5. We publish in Table 1 our best estimates of the epicentral parameters of Events 3, 5, 6, and 7 and would welcome testing their accuracy against any ground-truth information which might become available in the future. Such a comparison could be taken as indicative of the probable level of performance of *T*-phase stations in locating a major explosion in the context of monitoring the CTBT.

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