A Study of *T* Phases Recorded at the Kiruna Seismograph Station

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Abstract

Five cases of clear *T* phases from earthquakes in the Norwegian Sea recorded at Kiruna since 1951 are studied. To the author's knowledge these observations are the first of this kind in Europe and also the first so far to the north – on the borders of the Arctic Ocean. It is verified that *T* propagates as a sound wave through the water. Different phases, constituting *T*, have been identified, travelling along the land path as P_g, S_g, and a third wave with a land velocity of 2.7 km/sec, probably an *S* wave in more superficial layers than granite. The conditions for sound-channel transmission in the ocean are studied. The efficient use which can be made of the *T* phase in epicentre locations is illustrated with a particular case. The period of *T* is remarkably constant and equal to 0.5 sec. By amplitude calculations it is shown among other things that different submarine earthquakes are not equally efficient in producing a *T* phase; reasons for this behaviour are discussed. A theoretical explanation is given for the fact that at the same time as the amplitude of *T* may be larger than that of *P* at Kiruna, there is no *T*, but a clear *P* recorded at Uppsala.

1. Introduction

T phases (from submarine earthquakes) were first reported as observed at a few seismological stations in North America, namely Weston, Fordham, Ottawa, and Harvard for Dominican Republic and North Atlantic shocks and at stations on or near the Pacific coast for Japanese and Aleutian shocks (see TOLSTOY and EWING, 1950, and LEET, LINEHAN, and BERGER, 1951). Observations of *T* phases at the Honolulu seismograph station and at two SOFAR stations in the Pacific for circumpacific-belt shocks have been studied by EWING, PRESS, and WORZEL (1952). *T* phases have also been observed at Brisbane and at Bermuda. Observations at Morne des Cadets, Martinique, of Dominican Republic shocks by COULOMB and MOLARD (1949) also refer to the *T* phase. Recently WADATI and INOUYE (1953) have reported observations of *T* phases at Japanese seismograph stations for earthquakes in the Japanese waters.

The present paper is a report of a study of records of *T* phases obtained by means of a short-period vertical seismograph, type Grenet-Coulomb, at the Kiruna seismograph station. Kiruna is situated in North Sweden at 67° 50'.4 N, 20° 25'.0 E (see map Fig. 1). The station was installed in the summer of 1951 and since that time five clear cases and one doubtful of

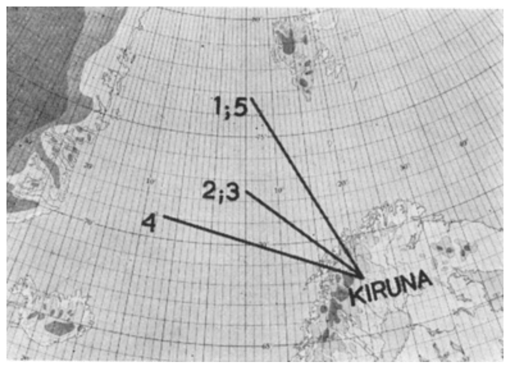


Fig. 1. Location of epicentres 1—5 (at the ends of the lines) and the paths to Kiruna.

Table 1. Observations of *T* phases.

rth-ake [o.	Date	Origin Time[1] G.M.T.	Epicentre	Distance to Kiruna km			*P* (Kiruna)	Observed arrival times of *T* (Kiruna)	Remarks
				Total path	Water path	Land path			
1	1951, Oct. 16	06.54.33* (USCGS) 06.54.30 (BCIS)	76° N, 5° E (USCGS) 77° N, 6° E (BCIS)	1,070	830	240	06.56.50	i_1 07.04.44 i_2 07.05.33	i_1 has a sharp beginning. Su: face waves very well deve oped. A position of epicent between the two determin: tions has been adopted.
2	1952, Feb. 10	06.10.05* (USCGS) 06.10.03 (BCIS)	72° 1/2 N, 2° E (USCGS) 73° N, 0° (BCIS)	780	530	250	06.11.52	i_1 06.16.47 i_2 06.17.07 i_3 06.17.45	Epicentre location has bee redetermined to be at 72° ½N 5° E.
3	1952, Apr. 28	01.15.15 (USCGS)	500 km NE of Jan Mayen	780	530	250	01.17.01	i_1 01.22.13 i_2 01.22.46	Same epicentre as in 2.
4	1952, Dec. 10	05.58.06 (USCGS)	71° N, 7° W (USCGS)	1,110	850	260	06.00.31	e_1 06.08.25 i_2 06.09.18 i_3 06.09.40	Beginning of e_1 uncertai disturbed by preceding m tion. Very well develope surface waves of long dur tion (up to 07*h*).
5	1953, Apr. 24	02.09.44* (USCGS) 02.09.42 (BCIS)	76° 1/2 N, 6° E (USCGS) 77° 1/2 N, 6° 1/2 E (BCIS)	1,070	830	240	02.12.07	e_1 02.20.08 i_2 02.20.39 i_3 02.20.45 i_4 02.21.16	Same epicentre as in 1 ha been used.

[1] In case more than one value is given, a star indicates which value has been adopted.

T phases have been obtained.[1] They are all due to submarine earthquakes in the area from Jan Mayen to Spitzbergen. To the author's knowledge these observations are the first of this kind in Europe and also the first so far to the north on the earth – on the borders of the Arctic Ocean.

2. Arrival times of *T*

Typical cases of *T* phases recorded at Kiruna are shown in Fig. 2. The *T* phase begins with small amplitudes, often gradually, and the largest amplitudes are later. In all cases several phases could be distinguished within the *T* group and they have all been measured. Table 1 gives the measured arrival times and other necessary data for the earthquakes. Reference is made to the Kiruna seismic bulletins for the usual analyses. All measurements of the *T* phases were made before I had any knowledge whatsoever where to expect them; this is of essential importance, especially in the case of gradually beginning phases.

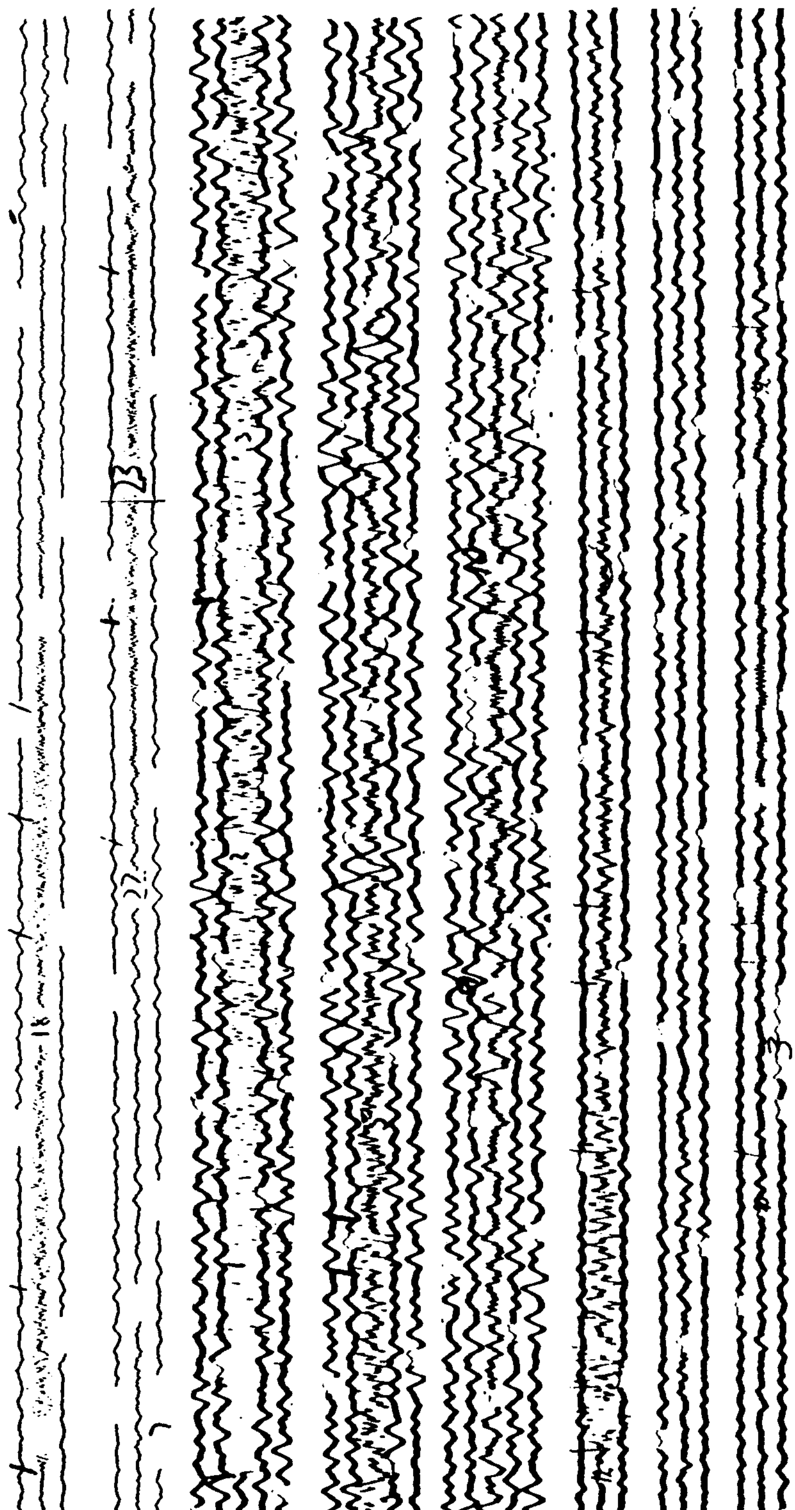
[1] Note added in proof: Another five cases of *T* phases have been recorded up to January, 1954.

The observed phases are definitely not *PcP*, *PcS*, or *ScS*, as the expected arrival times of these phases deviate by far too much from the observed times. Nor are the observed phases due to aftershocks or other earthquakes as in that case the very striking regularities in the arrival times and computed velocities (see below) would not be obtained.

It has been demonstrated conclusively by Ewing, Press, and Worzel (1952) that the *T* phase propagates as sound through the water and as body waves over the land path from the coast (or continental shelf) to the station. I therefore started with this assumption and computed the total travel times t simply from the equation

$$t = \frac{a_1}{v_1} + \frac{a_2}{v_2},$$

where a = the length of path, v = the corresponding velocity, and index 1 stands for water and 2 for land. The values of a_1 and a_2 are given in Table 1 as measured on a map (compare Fig. 1), assuming the transition to take place close to the actual coast. This seems justified as the sea is rather deep right up to the



a

b

c

Fig. 2. Examples of records of earthquakes with *T* phases at Kiruna. In each case the record begins with the *P* phase in the uppermost figure. The *T* phases are visible on the central line of the lowest figure in each case. For details compare *Table* 1.

a. April 28, 1952.
b. December 10, 1952.
c. April 24, 1953.

coast and the limit of the continental shelf coincides practically with the coast for cases 2—4; for 1 and 5 the path is to a large extent along the continental shelf, which passes from the northern part of the Norwegian coast towards NW just SW of Spitzbergen. For v_1 the value 1.45 km/sec was used. In our conditions this value is the most likely (see Table 5 below). Depending on the angle of incidence of the sound wave in water on to the solid ground, it is just as likely to get *S* waves as *P* waves in the crust. Detailed energy computations of interest in this connection have been made by ERGIN (1952). Therefore a_2/v_2 was computed both with $v_2 = 5.57$ km/sec for P_g and with $v_2 = 3.36$ km/sec for S_g (see BULLEN, 1947, p. 193). The arrival times computed in this way are given in Table 2, where the phases have been denoted in the same way as in Table 1 (i_1, i_2, etc) in order to facilitate comparison. The differences between observed and calculated times ($O - C$) are also given in Table 2.

Table 2. Comparison between observed (*O*) and calculated (*C*) arrival times of *T*.

Earthquake No.	Calculated arrival times of *T*	*O—C* sec
1	i_1 07.04.48 (P_g)	— 4
	07.05.16 (S_g)	
2	i_1 06.16.56 (P_g)	— 9
	(i_2) 06.17.25 (S_g)	(— 18)
3	i_1 01.22.06 (P_g)	+ 7
	i_2 01.22.35 (S_g)	+ 11
4	e_1 06.08.39 (P_g)	— 14
	i_2 06.09.09 (S_g)	+ 9
5	e_1 02.19.59 (P_g)	+ 9
	i_2 02.20.27 (S_g)	+ 12

In the computation of the arrival times of *T* the time it takes for a *P* or an *S* wave to travel from the focus to the ocean bottom (approximately 5 and 8 sec resp.) has been neglected as such a correction is completely within the limits of error.

The agreement between observed and computed arrival times is as good as can ever be required, regarding the following uncertainties:

1. Uncertainty of observed arrival times. This may amount to several seconds, as the phases are usually not as sharp as the direct body waves.
2. Uncertainty of the position of the assumed transition point water to land.
3. Local deviations from the assumed velocities of P_g and S_g. A computation, using a value of 6.0 km/sec for the P_g velocity, however, showed that this change is of no importance in this connection.

The good agreement is even more obvious from a calculation, following up point 2 above. It has been calculated how much the assumed transition point must be displaced along the great circle arc between station and epicentre in order to give perfect agreement between observed and calculated arrival times with the assumed velocities. If *t* is the observed travel time, it is easily shown that the water path constitutes a fraction *x* of the total path *a* given by

$$x = \frac{t - \dfrac{a}{v_2}}{a\left(\dfrac{1}{v_1} - \dfrac{1}{v_2}\right)}$$

The results are given in Table 3.

Table 3. Comparison between calculated and measured water and land paths.

Earthquake No.	Phase	Calculated paths km		Measured paths km	
		Water	Land	Water	Land
1	P_g	824	246	830	240
2	P_g	515	265	530	250
3	P_g	546	234	530	250
	S_g	562	218	530	250
4	S_g	877	233	850	260
5	P_g	845	225	830	240
	S_g	813	257	830	240

It is obvious that with a maximum displacement of not more than about 30 km, the differences are completely within the limits of error. The coast is actually not very well defined as it consists of a large archipelago.

If, on the other hand, with the measured values of a_1, a_2, and *t* and assumed value of v_2 we calculate v_1, we obtain the velocities given in Table 4.

The agreement with Table 5 below is perfect. The agreement between the velocities of *T* and of sound through the ocean in our case taken together with the same agreement found

Table 4. Computation of sound velocity in sea water.

Earthquake No.	Sound velocity (km/sec) calculated from	
	P_g	S_g
1	1.46	
2	1.48	1.52
3	1.42	1.41
4	1.49	1.43
5	1.43	1.42
Mean ($n = 9$) and standard error........	1.45 ± 0.01	

by EWING, PRESS, and WORZEL (1952) in conditions where both velocities were higher than here, is of decisive importance.

Thus, there seems to be no better explanation of the observed *T* phases than that according to which they propagate as sound waves through the water (either as a SOFAR channel wave or by multiple internal reflections) and as body waves over the land. We have found clear indications of *T* phases travelling as P_g over the land (observed in all cases 1 – 5); of *T* phases travelling as S_g over the land (observed in cases (2), 3 – 5);and of *T* phases travelling as a third wave with a velocity of 2.7 km/sec over the land (observed in cases 1 (i_2), 2 (i_3), [4 (i_3)], 5 (i_3) with land velocities 2.73, 2.66, [2.41], 2.70 km/sec resp.). The third wave may be an *S* wave in layers shallower than the granite. It will be denoted S_x in what follows. An explanation of the successive *T* phases as all being P_g but having different points of transition is excluded because that would require an inland transition point for the later phases which is not possible. A longer path than along the great circle in combination with lateral refraction may account for some of the later motion, but only in cases 2 and 3, and cannot be taken as an explanation of the later phases.

When the computed arrival times had been obtained (Table 2) they were compared with the records with the following results. In case 1 the computed S_g at 07.05.16 is confirmed by a phase, observed at 07.05.12. In case 2 the computed S_g at 06.17.25 corresponds to no marked phase at just this time, but there is an increase of the amplitudes at that time. The same is true at the computed time 01.22.35 of S_g in case 3. In case 4 a phase at 06.09.11 was on the point of being measured already in the first measurement, agreeing well with the computed S_g at 06.09.09. As the times, measured when the computed arrival times were known, cannot be attached the same weight as the originally measured times, they have been used in no calculations in this paper.

In addition to the clear cases reported in Table 1 also a somewhat doubtful *T* phase has been observed, where the possible *T* phase was disturbed by other short-period movements. This earthquake occurred on October 19, 1951, at 00.54,3 SW of Spitzbergen (about the same epicentre as 1 and 5) at an epicentral distance of 1,050 km from Kiruna (water path 810 km; land path 240 km). Kiruna recorded *P* at 00.56.38. A *T* phase of very small amplitude and short duration was observed at 01.04.22, which agrees well with the computed arrival time of 01.04.20 for $T(P_g)$.

The observations of *T* phases have particular interest in view of the recent investigations of long-range sound transmission by SOFAR channels by EWING and WORZEL (1948). Calculations of the depth of the SOFAR channel is not easy for our region, because data on water temperatures and salinities at greater depth usually seem to be limited to the summer months and because the conditions (temperatures and salinities) along the path through the water are rather variable. Every path crosses the Gulf Stream but passes also to a large extent through the drift-ice region. The vertical distribution of sound velocity has been computed, using three soundings by HELLAND-HANSEN and NANSEN (1912) in the sea west of Spitzbergen. The results are compiled in Table 5. New measurements in the Polar Sea have been given by WORTHINGTON (1953).

During winter the water temperature is probably not far from 0° C in this region with very little vertical variation. In these circumstances the minimum sound velocity is to be found in the surface layers of the ocean. With the indicated depths of the sound channel it is obvious that the sound travels right up to the steep coast before being transformed into body waves in the crust.

Table 5. Sound-channel transmission.

Date of sounding 1910	Location	Sound velocity at surface km/sec	Minimum sound velocity km/sec	Depth of SOFAR channel m	Remarks
July 18.....	78° 08′ N, 0° 35′ W	1.456	1.453	400—500	Location at the ice-edge
July 19.....	77° 41′ N, 0° 22′ W	1.463	1.452	200—500	Location at the ice-edge
July 19.....	77° 15′ N, 0° 35′ W	1.462	1.445	150	

3. Use of *T* in epicentre location

Due to the large difference in velocity of propagation of the direct *P* wave and the *T* wave, the difference in arrival times of *P* and *T* is very sensitive to changes in the epicentre location. Assuming only a water path for the *T* wave it is easily shown that within 15° from the epicentre the variation of the travel time difference between *T* and *P* with epicentral distance, $\frac{d\ (\delta t)}{d\ \Delta}$, is not less than 0.56 sec/km or 62 sec/1°. For greater epicentral distances this effect is still more pronounced due to the curvature of the travel time curve for *P*.

The use of *T* in epicentre location will be illustrated by case 2 above. In this case (but not in the others) a discrepancy was found between the epicentre location by BCIS and USCGS and the distance computed from the Kiruna record alone. BCIS gave $O = 06.10.03$ and epicentre at about 73° N, 0° long; USCGS gave $O = 06.10.05$ and epicentre at 72° $^1/_2$ N, 2° E. These epicentres correspond to the distances 940 km = 8°.5 and 870 km = 7°.8 resp. from Kiruna. The *S*—*P* difference for Kiruna (both *S* and *P* are very well defined in the records) is 1 min. 19 s, i.e. a distance of only 780 km = 7°.0. The travel time of *P* to Kiruna (assuming $O = 06.10.05$) is 1 min. 47 s, also giving an epicentral distance of only 780 km = 7°.0. The epicentre given by BCIS was corrected by means of a method introduced by Gutenberg and Richter (1937). There was no change of origin time, but the epicentre had to be displaced somewhat eastwards (to approx. 73° 1/4 N, 3° 1/2 E). But even this new epicentre, which gives the best fit to all station observations, does not agree with the observations at Kiruna alone; the distance to Kiruna from this point is 860 km = 7°.7. The arrival times of *T* computed for an epicentral distance of 860 km are 06.17.49 (P_g) and 06.18.18 (S_g), thus deviating by far too much from the observed times, whereas the agreement is good if an epicentral distance of 780 km = 7°.0 is assumed (see Table 2). This case clearly illustrates the efficient use which can be made of the *T* phase in epicentre locations and it also shows how records of sensitive instruments at relatively near stations may in cases give more trustworthy results than a value based on a large number of observations, many of which may be less certain. The most probable location of the epicentre is 72°1/2 N, 5° E (determined from the distance 7°.0 to Kiruna and the distance 13°.6 to Uppsala, obtained from the travel time of *P* to Uppsala).

4. Periods, amplitudes, and duration of the *T* phase

The periods of the observed *T* waves are 0.5 sec in all cases with only a few periods slightly lower (0.4 sec). Only in case 5 a couple of waves had a period of 0.7 sec. The period is also remarkably constant through the whole *T* phase in every case. The extent to which the observed periods depend on the response of the seismograph cannot be decided as that would require several instruments with different response. The short-period vertical seismograph (Grenet-Coulomb) at Kiruna had a maximum magnification of 8,500 at 0.3 sec period in cases 1—3 and a maximum magnification of 9,615 at 0.6 sec in cases 4—5. The fact that no change occurred in observed periods of *T* between the two instrumental set-ups indicates that the period 0.5 sec for *T* is of real importance.

Table 6. Amplitudes (A) and periods (τ) of P, surface waves, and T.

Earthquake No.	P		Surface waves						T(Z)						Duration of observable T phase min
	Z		E		N		Z		P_g		S_g		S_x		
	A μ	τ sec	A μ	τ sec	A μ	τ sec	A μ	τ sec	A μ	τ sec	A μ	τ sec	A μ	τ sec	
1	0.6	1	3.8	15	8.3	18	7.7	19	0.05	0.5	—	—	0.07	0.5	2.7
2	0.4	0.5	9.3	19	4.8	18	13	18	0.07	0.5	0.3	0.5	0.5	0.5	4.7
3	0.1	0.7	not recorded						0.07	0.5	0.1	0.5	—	—	3.8
4	2.7	1.0	56	17	40	17	—	—	0.03	0.5	0.07	0.5	0.08	0.5	5.0
5	0.4	1.0	8.8	17	7.0	17	—	—	very small ampl.		0.03	0.5	0.04	0.5	3.2

On one hand, the amplitudes of T may in some cases be larger than that of P at Kiruna (see case 2 in Table 6 below; this has also been found in some cases for other stations by other investigators). On the other hand, we observe that in the same case the T phase has disappeared before Uppsala is reached, whereas P is clearly recorded there. These circumstances, which at first sight may seem surprising, are easily explained by a consideration of the geometry of the energy distribution. The energy of the sound wave through the water diminishes essentially as the inverse first power of the distance, whereas for the body waves the energy decreases as the inverse second power of the distance. As a first approximation we assume homogeneous media and neglect extinction. The energy of the P wave at a station at a hypocentral distance r is then

$$(E_r)_P = \frac{c' E_0}{r^2},$$

where c' is a constant, equal to 1 if E_0 is the energy for $r = 1$. The corresponding energy at the epicentre (h = depth of hypocentre) is

$$(E_h)_P = \frac{c' E_0}{h^2}$$

A certain fraction α of this energy passes into the water as a sound wave. The energy of the T wave at a station (assumed coastal) is then

$$(E_r)_T = \frac{c\alpha\,(E_h)_P}{r} = \frac{c\alpha r}{h^2}(E_r)_P \qquad (1)$$

where the constant c depends on the energy transmission at the coast. From eq. (1) it is obvious that $(E_r)_T > (E_r)_P$, if $r > \frac{h^2}{c\alpha}$, which may easily happen. We then consider another station at distance R from the epicentre ($R > r$) and with a land path $= R - r$. The energy of the direct P wave is at that point

$$(E_R)_P = \frac{c' E_0}{R^2}$$

and the energy of the T wave, considering only the direct wave, is

$$(E_R)_T = \frac{c'\,(E_r)_T}{(R-r)^2} = \frac{cc'\alpha}{h^2}\,\frac{1}{r\left(1-\frac{r}{R}\right)^2}\,(E_R)_P \qquad (2)$$

From eq. (2) we find that $(E_R)_T < (E_R)_P$ at all points with epicentral distances

$$R > \frac{r}{1-\sqrt{\frac{cc'\alpha}{r\,h^2}}}$$

From eq. (2) we further find that

$$\lim_{R \to r} \frac{(E_R)_T}{(E_R)_P} = \frac{c\alpha r}{h^2}$$

in agreement with eq. (1). These simple considerations help to explain why the amplitudes of T may be larger than those of P at Kiruna, and in the same case there is no T phase but a clear P phase recorded at Uppsala.

The amplitudes of $P(Z)$, the seismic surface waves, and the various T phases as well as the duration of the observable T motion are given in Table 6. The amplitudes of the surface waves were obtained from the records of the Galitzin instruments, the others from the Grenet-Coulomb seismograph.

We find that the amplitudes of S_x are larger than those of S_g, and these in turn are larger than those of P_g. The largest amplitudes during the whole T phase belong to S_x. Comparing earthquakes at the same place (2 and 3; 1 and 5) we find that the amplitude of T increases with increasing earthquake magnitude; larger amplitudes of P are accompanied by T phases which are larger in approximately the same proportion. This is, however, not valid in comparison of earthquakes with different foci, even if the epicentral distance is the same. On the other hand, there is no obvious correlation between amplitudes of T and of the surface waves in 1 and 5. The duration (which is difficult to determine exactly) increases with increasing T amplitude in 2—3, but not in 1 and 5. If we compare earthquakes with different foci, we find that those NE of Jan Mayen (2, 3) are favourable for a large T phase; the ratio of the amplitudes of T to those of P as well as the surface waves are much larger than for the other earthquakes. The earthquake 4 at Jan Mayen is least effective in producing a T phase, whereas the Spitzbergen earthquakes (1, 5) are intermediate. The amplitudes of both P and the surface waves are about seven times as large in 4 as in 5 (same epicentral distance), whereas the amplitude of T is only twice as large in 4 as in 5. The same fact is clear from a comparison of 1 and 4.

The ratio of amplitude and period (A/τ) for T divided by the same ratio for P is given in Table 7.

Table 7. Ratios of A/τ for various T phases and P on the vertical component.

Earthquake No.	$\left(\frac{A}{\tau}\right)_T : \left(\frac{A}{\tau}\right)_P$		
	P_g	S_g	S_x
1	0.16	—	0.23
2	0.18	0.75	1.25
3	1.00	1.43	—
4	0.02	0.05	0.06
5	—	0.15	0.20

It does not seem excluded that a certain ratio of A/τ for T and P is characteristic for a certain epicentre location and a given station; this ratio is about 0.2 for 1 and 5; 1.3—1.4 for 2 and 3; 0.06 for 4. However, these numbers depend not only on the energy conditions but also on the lengths of the different water and land paths. From eq. (2) above we easily find that

$$\left(1-\frac{r}{R}\right)\sqrt{r}\,\frac{\left(\frac{A}{\tau}\right)_T}{\left(\frac{A}{\tau}\right)_P}=\sqrt{\frac{cc'\alpha}{h^2}}$$

where r = the lengh of the water path, R = the total path-length, h = the depth of the hypocentre, and c, c', α are constants depending on the energy transmission. The ratio of the factor $\left(1-\frac{r}{R}\right)\sqrt{r}$ for cases 2 and 3 to its value in cases 1, 4, and 5 is 0.7. Therefore comparative figures of $(A/\tau)_T : (A/\tau)_P$ for maximum amplitudes of T, when the influence of different path-lengths has been eliminated, are 0.23 for case 1, 0.88 for case 2, 1.00 for case 3, 0.06 for case 4, and 0.20 for case 5. These figures depend only on the efficiency of the different earthquakes in producing T phases. Still, cases 2 and 3 are the most efficient in producing a T phase. However, a much larger material than is now available is necessary in order to study these phenomena in detail.

By the way, the formula above indicates that the amplitude ratio of T and P is inversely proportional to focal depth. This is confirmed by observations of LEET, LINEHAN, and BERGER (1951, p. 136, fig. 4, showing ratios of amplitudes of T and S plotted against focal depth or actually against pP—P).

There are several reasons for obtaining different amplitudes of T phases even from earthquakes of the same magnitude and at the same epicentral distance. The reasons may be summarized as follows:

1. The conditions for the energy transmission to the water at the epicentre (flat or sloping bottom).
2. The ocean depth at the epicentre in relation to the depth of the sound channel. For 1 and 5 the ocean depth at the epicentre is about 2,600 m, for 2 and 3 about 2,500 m, and for 4 about 1,500 m.
3. The conditions along the water path and the incidence on the shelf or coast. Part of the path for 1 and 5 is almost tangential to the continental slope, causing partial loss of energy due to refractions. The incidence upon the coast itself is practically normal in all cases. Especially for 2 and 3 the incidence is also normal to the continental slope. As the con-

Table 8. Earthquakes in the Norwegian Sea and the Atlantic which have given no *T* phases at Kiruna.

Date	Origin time G.M.T.	Epicentre	Distance to Kiruna km	*P*(*Z*)		Surface waves					
						E		*N*		*Z*	
				A μ	τ sec	*A* μ	τ sec	*A* μ	τ sec	*A* μ	τ sec
1951, Dec. 4	08.50.50	54°½ N, 36° W	3,220	0.1	1.0	not recorded					
1952, Feb. 21	16.06.05	NE of Greenland	—	0.1	0.7	not recorded					
1952, Mar. 8	11.33.05	70°½ N, 15° W	1,400	—	—	4.9	17	2.3	15	5.2	16
1952, Mar. 8	11.36.57	70°½ N, 15° W	1,400	0.05	0.5	not recorded					
1952, Mar. 9	05.44.29	70°½ N, 15° W	1,400	—	—	6.8	13	4.6	15	—	—
1952, Mar. 12	12.13.10	64° N, 22° W	1,950	0.1	1.0	3.1	15	2.2	13	2.8	15
1952, May 16	14.32.3	63°½ N, 22°½ W	1,980	—	—	1.7	14	0.9	11	1.3	14

tinental edge is here concave towards the sea we may expect convergence on the land side, which may in part explain the large *T* amplitudes observed in these two cases.

No correlation between the occurrence of *T* phases and tsunami has so far been definitely proved (Ewing, Tolstoy, and Press, 1950; Leet, 1951; Wadati and Inouye, 1953). If this is proved or disproved it may give further information on the mechanism for production of *T* phases, as the mechanism producing a tsunami is known.

The station at Scoresby-Sund would have a favourable location for recording *T* phases from earthquakes in the Norwegian Sea, having in general an unbroken water path. The records of the short-period vertical seismograph (Grenet-Coulomb) were investigated for the cases available (1—4), but I found no trace of a *T* phase. This is probably due to low magnification of the instrument. The reason is most probably the same why Reykjavik has recorded no *T* phase in the cases 1—5; the Reykjavik Sprengnether records were examined. An examination of the Uppsala records of a short-period vertical Grenet-Coulomb instrument for the cases 2—5 (the only available for comparison) also showed that no trace of a *T* phase had been recorded. The reason cannot here be low sensitivity, as this is practically the same as at Kiruna, but the reason is instead the much longer land paths. Table 8 gives a list of earthquakes in the Norwegian Sea which have been recorded at Kiruna but without *T* phase, probably due to low magnitudes and to large epicentral distances and possibly to other unfavourable conditions for production of a *T* phase.

No earthquakes in other parts of the Atlantic Ocean have so far given *T* phases at Kiruna. Not even the strong earthquake on March 19, 1953, in the Antilles (14° N, 61° W, depth = 150—200 km, magnitude = $7^{3}/_{4}$, distance 74° to Kiruna) gave a *T* phase at Kiruna, although almost the entire path was oceanic. The path had to cross the Mid-Atlantic Ridge under a small angle and this may possibly prevent the transmission.

REFERENCES

BULLEN, K. E., 1947: *An Introduction to the Theory of Seismology.* Cambridge, 276 pp.

COULOMB, J., et P. MOLARD, 1949: Ondes séismiques au fond de la Mer des Antilles. *Ann de Géophys.*, **5**, pp. 212–214.

ERGIN, K., 1952: Energy Ratio of the Seismic Waves Reflected and Refracted at a Rock-Water Boundary. *Bull. Seism. Soc. Am.*, **42**, 4, pp. 349–372.

EWING, M., and J. L. WORZEL, 1948: Long-Range Sound Transmission. *Geol. Soc. Am. Mem.*, **27**, pp. 1–35.

EWING, M., I. TOLSTOY, and F. PRESS, 1950: Proposed Use of the *T* Phase in Tsunami Warning Systems. *Bull. Seism. Soc. Am.*, **40**, 1, pp. 53–58.

EWING, M., F. PRESS, and J. L. WORZEL, 1952: Further Study of the *T* Phase. *Bull. Seism. Soc. Am.*, **42**, 1, pp. 37–51.

GUTENBERG, B., and C. F. RICHTER, 1937: Materials for the Study of Deep-Focus Earthquakes (Second Paper). *Bull. Seism. Soc. Am.*, **27**, 3, pp. 157–183.

HELLAND-HANSEN, B., and F. NANSEN, 1912: *The Sea West of Spitsbergen. The Oceanographic Observations of the Isachsen Spitsbergen Expedition in 1910.* Christiania (Oslo), 89 pp.

LEET, L. D., D. LINEHAN, S. J., and PH. R. BERGER, 1951: Investigation of the *T* Phase. *Bull. Seism. Soc. Am.*, **41**, 2, pp. 123–141.

LEET, L. D., 1951: Discussion of "Proposed Use of the *T* Phase in Tsunami Warning Systems". *Bull. Seism. Soc. Am.*, **41**, 2, pp. 165–167.

TOLSTOY, J., and M. EWING, 1950: The *T* Phase of Shallow-Focus Earthquakes. *Bull. Seism. Soc. Am.*, **40**, 1, pp. 25–51.

WADATI, K., and W. INOUYE, 1953: On the *T* Phase of Seismic Waves Observed in Japan. *Proc. Japan Acad.*, **29**, 2, pp. 47–54.

WORTHINGTON, L. V., 1953: Oceanographic Results of Project Skijump I and Skijump II in the Polar Sea 1951—1952. *Trans. Am. Geophys. Un.*, **34**, 4, pp. 543—551.