On *Lg* as read in North American records (*)

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Important studies of surface waves of earthquakes carried out in recent years, for the greater part by M. Ewing, F. Press and other workers of the Lamont Geological Observatory, Palisades, New York, have led to remarkable advances in the understanding of these waves. Dispersion as observed is now found to be concordant with the dispersion derived theoretically, and new important phases have been recognized.

Propagation over oceanic paths has long been known to differ distinctly from propagation over continental paths. In recent years the structure of the bottom of the deep ocean has been determined in various places by means of refraction measurements and with approximately the same results everywhere. When these results were used and the water layer taken to be active in Rayleigh wave propagation, dispersion curves could be calculated to which the dispersion as observed had a very good fit. The long duration of the Rayleigh wave train, extending to group velocities smaller than the velocity of sound in water, could now be explained. Theory predicted the presence of an Airy phase which was actually found in the records. Another oceanic phase, the T phase, was also discovered.

In the continental trains of surface waves two new phases were also found. These are the *Lg* and the *Rg* phases appearing in the Love wave train and the Rayleigh wave train respectively (Press and Ewing 1952). While the T phase which develops in the ocean may travel part of its way in continental structure the *Lg* and *Rg* phases are exclusively continental and are absent when a deep ocean path intervenes. Therefore the short-period *Lg* phase which is often quite conspicuous and more easily identified than *Rg* can be used to distinguish between continental and oceanic structure in water covered areas (see Oliver et al. 1955). Thus the discovery af the phase has provided an important tool for the exploration of the crust.

The period of *Lg* is usually from ½ sec. to 6 sec. and the phase was first discovered

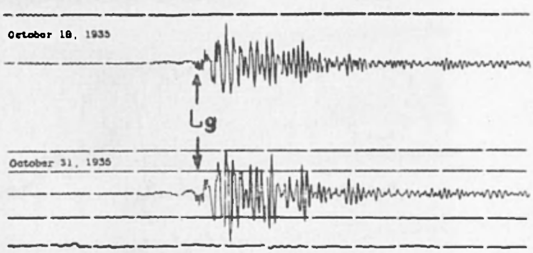

Fig. 1. – Pasadena records of Helena earthquakes, October 1935

in the records of special Palisades instruments having particularly good response in this range. *Lg* was quite often the largest phase in the records and therefore very conspicuous. But, once discovered, the phase was found to be present in the records of other instruments, also the long-period ones. Here the short-period *Lg* waves have smaller amplitude, but at small and moderate distances they arrive at the time when the long-period Love waves are about to rise to maximum amplitude so that the combined phase has great prominence. In fig. 1 are seen the Pasadena linear strain records of the Helena earthquakes of Octo-

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ber 19, and 31, 1935. The epicentral distance is 13°.2.

Press and Ewing read the onset of *Lg* in about 40 Northamerican records, most of them Palisades records, and found the velocity 3.51 ± 0.07 km/sec. This is close to the S velocity in the upper part of the crust and since the motion is mainly transverse although it has also longitudinal and vertical components it is likely to be produced by *S* waves confined to the upper crust. The

Fig. 2 shows the record of a shot fired on the bottom of water 60 feet deep. There are 7 channels fed from 3 pick-ups and with the imputs magnified and filtered in such a way that high and low frequency response of varying intensity is obtained. What is seen to happen is this: first low frequency ground waves arrive and after an interval of time a high frequency wave sets in abruptly. It has the velocity of sound in water and is termed the water wave. Ewing

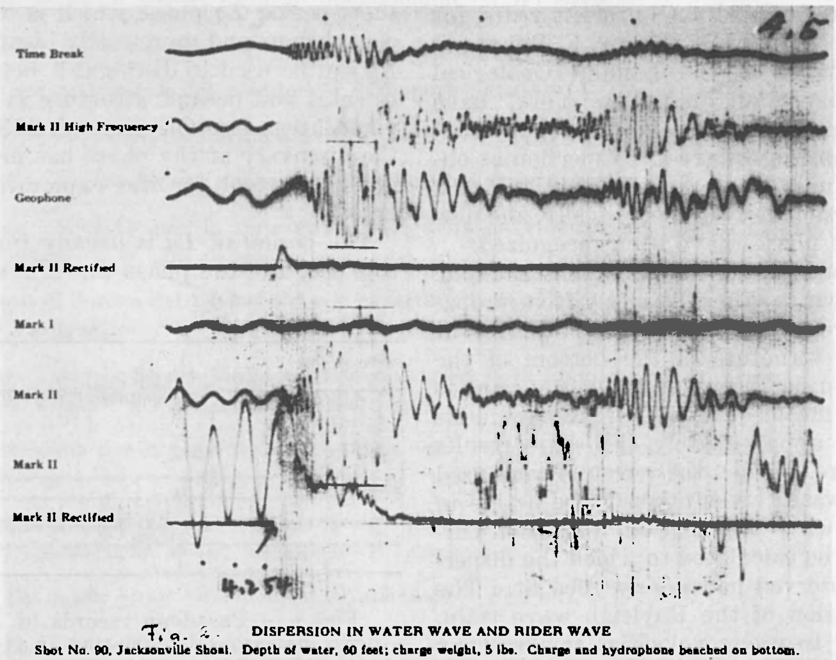

Fig. 2. – Dispersion in water wave and rider wave

mechanism is not yet fully understood, but as was immediately noticed by Press and Ewing, the combined *Lg* and Love wave phase resembles the records obtained of explosions in shallow water and therefore is likely to be explainable in a similar way. The explosion results are described in the work: " Propagation of Sound in the Ocean " (Ewing, Worzel and Pekeris, 1948) which may be considered an introduction to the Palisades investigations of surface waves. Some of the pertinent results will be described briefly.

observed that it had inverse dispersion, i. e. its frequency decreased with time. This is seen more clearly in fig. 3.

Pekeris in the second part of the work mentioned works out the complete theory of the propagation of sound set up in a liquid layer resting on a liquid bottom in which the velocity of sound is greater. Waves will be guided through the upper layer by multiple reflections at the surface and at the bottom, setting up at the same time waves in the bottom. Ewing and Worzel determined the wave velocity in the bottom

— or bottom layers — by the usual refraction methods and Pekeris, using their results, was capable of predicting the events actually found to take place. Working out the dispersion in the ground and water waves on his theory he found agreement with observation. He found also that the water wave would arrive riding on a ground wave, the rider wave, and that the frequency of the rider wave increased while that of the water wave decreased until they had the same frequency, the Airy frequency, which remained constant. The amplitude reached its maximum during the process and in the Airy phase decreased. The rider wave and Airy frequencies as well as the cut off frequency of the first arriving ground wave all depend on the depth of water and the velocity in the bottom and, when known, aid in the determination of these quantities.

Fig. 4 shows Pekeris' model dispersion curve. One branch represents the dispersion curve of the ground wave. Frequency increases with decreasing group velocity. The other branch is that of the water wave, indicating inverse dispersion. At the lowest point corresponding to the smallest group velocity we have the Airy frequency. The rider frequency is the frequency of the low frequency branch corresponding to the frequency of the first arriving water wave.

Comparing figs. 1 and 2 we see that there is a similarity in the records in as much as we have high frequency waves arriving riding on low frequency waves the frequency of which increases while the amplitude reaches a maximum. As to the possible decrease of frequency in the high frequency *Lg* wave it would not be observable in this record where the time scale is too narrow. But, actually, the analogy is not complete. The *Lg* train of waves does not combine with the Love wave in the way in which the water wave combines with the ground wave but is often seen to continue as a short-period wave while the Love waves rise to a maximum and fall off.

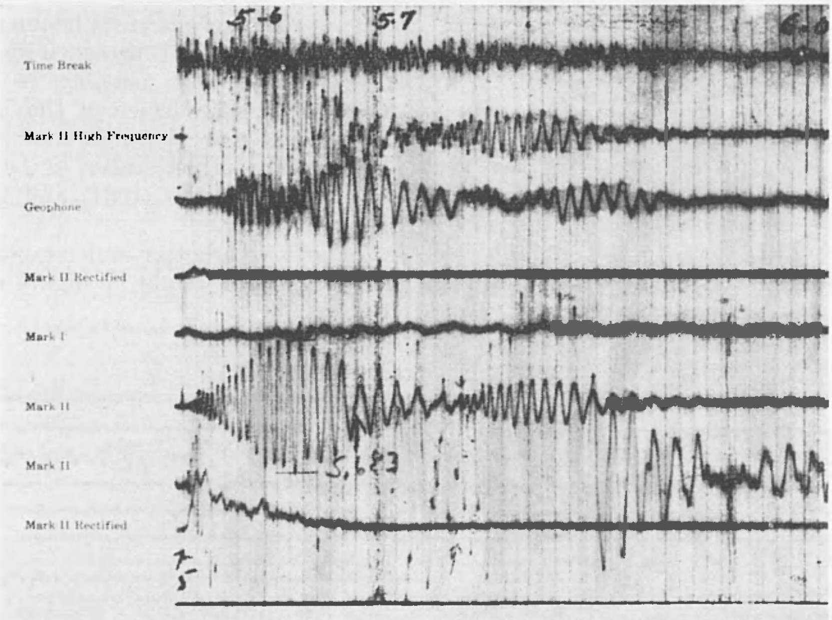

Fig. . DISPERSION IN WATER WAVE
Shot No. 91, Jacksonville Shoal. Depth of water, 60 feet; charge weight, 5 lbs. Charge and hydrophone beached on bottom.

Fig. 3. – Dispersion in water wave

There is, nevertheless, no doubt that *Lg* is guided along some kind of channel in the crust, and a very effective channel, since

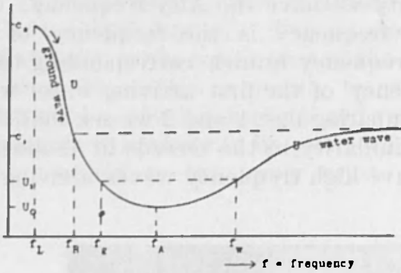

Fig. 4. – Dispersion curve
(Pekeris, 1948, p. 5)

the phase is known to have been observed at epicentral distances up to about 80°. The channel is not likely to be provided by the entire crust down to the Mohorovicic discontinuity, partly because the coupling

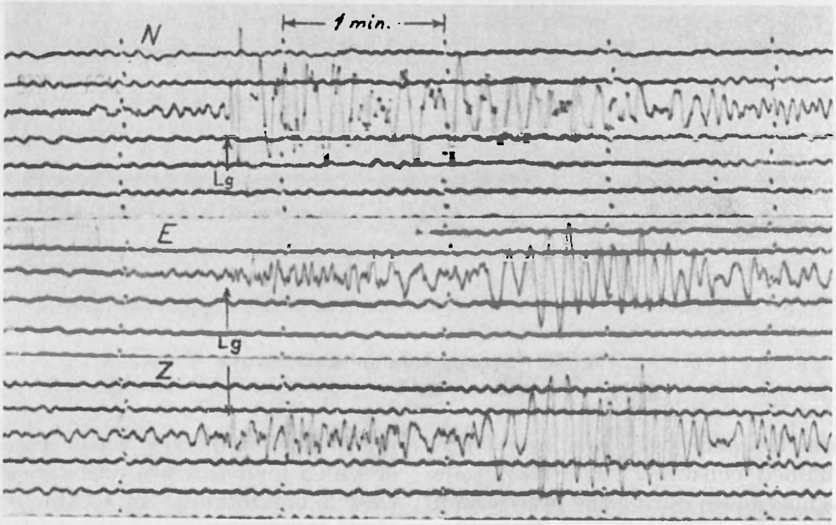

Fig. 5. – Palisades long-period records of Oklahoma earthquake

with the Love wave is not complete, partly because its velocity indicates that it is confined to the upper part of the crust. No sharp discontinuity seems to exist within the crust, at least not in Northeastern America, wherefore *Lg* is not likely to be set up by normal mode propagation in a channel having definite boundaries. But other wave guide actions have been considered. There is the SOFAR channel in the deep ocean discovered by Ewing and Worzel and described and explained by them in the third part of " Propagation of Sound in the Ocean ". Here a highly efficient wave guide is effected by velocity gradients above as well as below the axis of the channel. The oceanic T wave seems to be guided in a similar way. Wave guide action may also be set up in a channel having a discontinuity surface for one of its boundaries while the other boundary is provided by a velocity gradient. *Lg* may possibly be *S* waves reflected at the surface of the Earth and refracted at some depth in the crust. The problem of the mechanism of *Lg* has been considered by Båth (1954) and Gutenberg (1955).

To achieve a better understanding of the nature of *Lg* it would be highly important to have the complex wave motion analysed in a way similar to that in which the wave motion is analysed in the explosion experiments, but very few observatories have instruments of sufficiently varied response. The Palisades observatory, however, possesses a variety of instruments in the records of which a study of the components of the wave motion can be made. It has three components long-period instruments with a response similar to that of the Galitzin-Wilip; the three component electronic seismograph (pendulum period 12^s) previously mentioned with particularly good response to waves of period less than 5 sec.; three-component Benioff short- and long-period seismographs and a vertical electronic seismograph with pendulum period 1 sec. Since 1952 it has had in addition a vertical seismograph with pendulum period 10 sec. and galvanometer period 75 sec. and since 1953 two horizontals to match it.

In the Oklahoma earthquake of April 9, 1952 *Lg* was very well recorded on the various Palisades instruments at an epicentral distance of 19°.6. The epicentre being practically due west of Palisades, the east-west component instruments recorded longitudinal motion, the north-south components transverse motion. This is apparent in the long-period records of fig. 5. On the north- south component record marked *N* the first arriving surface waves are rather large Love waves of long period on which the short-period *Lg* waves are superposed. On the *E* and *Z* records the long-period waves are lacking, but the short-period *Lg* waves are clearly present. Later long-period waves, evidently Rayleigh waves, arrive on *E* and *Z*; the two wave trains are remarkably similar.

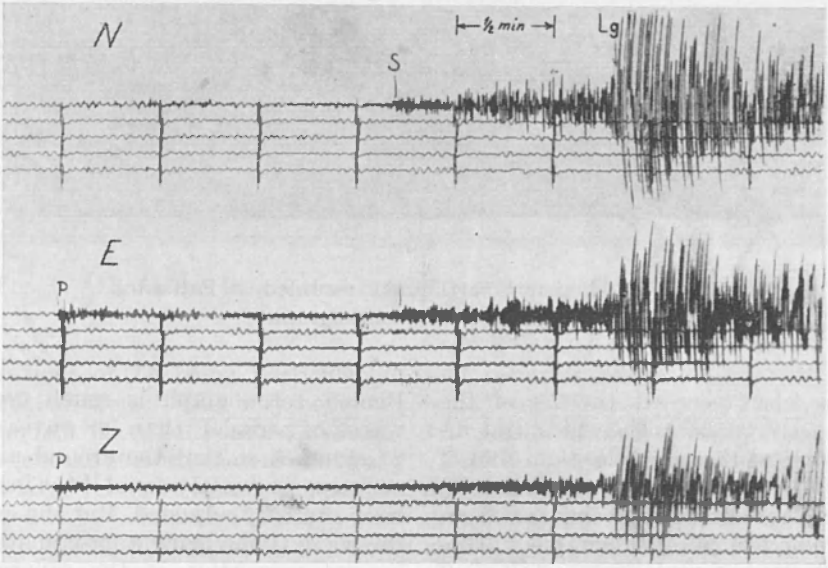

Fig. 6. – Oklahoma earthquake recorded at Palisades on 12^s electronic seismograph.

The records of the 12 sec. electronic seismograph present an entirely different picture (see fig. 6). Here the short-period *Lg* waves are very large and form a large group of waves in each record while the long-period Love waves are absent. *Lg* is largest on the *N* record, the transverse component, but also very large on the other two. The motion was too strong for the pens to record it distinctly and periods cannot be measured, but they are seen to be short.

The Benioff short- and long-period seismographs also record strong *Lg* with impulsive beginnings. On the short-period *N* record the movement is too strong to be readable. The 1^s electronic seismograph has a large but nevertheless distinct record of *Lg* (see fig. 7). Here the period is seen to vary between 1^s and $\frac{1}{2}^s$. Dispersion is not discernible in this record and short-period movement persists all during the long-period Love wave phase and longer. The *Lg* velocity as determined from the Palisades records is 3.59 km/sec.

much larger than at Pasadena. The dominant period at Palisades is 1^s, the maximum amplitude about 30 mm. while at Pasadena the period is about 2^s and the maximum amplitude 6 mm. The magnification of the

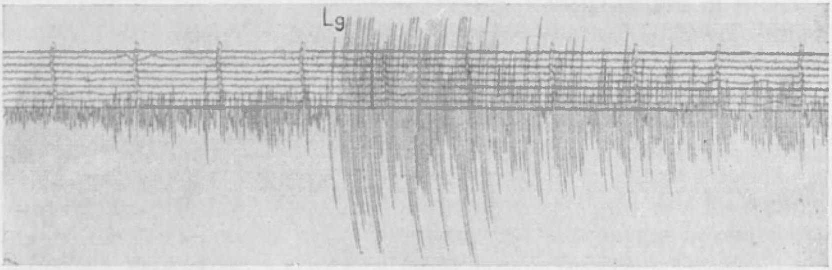

Fig. 7. – Oklahoma earthquake recorded at Palisades on 1^s electronic seismograph

Pasadena also recorded the *Lg* of the Oklahoma earthquake. The epicentral distance is $16^{\circ}.8$ and thus smaller than that of Palisades. The instrumentation is different, the only similar instruments being the short-period Benioff vertical seismographs. There are three-component long-period Benioff seismographs at both stations, but at Pasadena the galvanometer period is 90 sec. while at Palisades it is 75 sec. All the Benioff seismograph is much greater for waves of period 1^s than for waves of period 2^s, so much so that the ground amplitudes were approximately equal if the instruments were similarly adjusted. But the energies of the wave trains being approximately in the inverse proportion of the squares of the periods when the amplitudes are equal, the energy of *Lg* would then have been far greater at Palisades than at the nearer

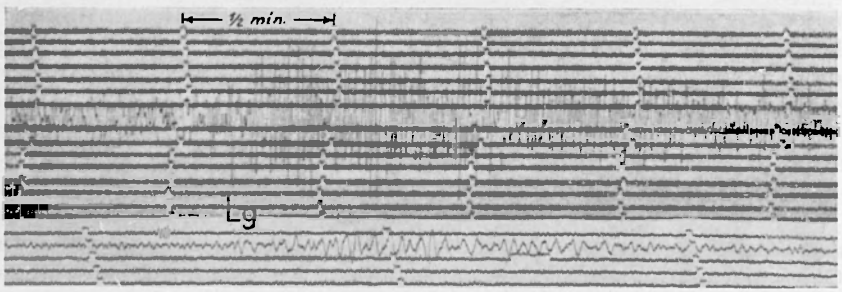

Fig. 8. – Oklahoma earthquake recorded at Palisades and at Pasadena on short-period Benioff vertical seismographs

Benioff records of Lg are much smaller at Pasadena than at Palisades. The two short-period vertical records are seen in fig. 8. They differ greatly, the period of the Palisades *Lg* being shorter and the amplitudes Pasadena station. It is not known whether the adjustment of the instruments actually is the same, so no definite conclusion can be reached, but it does not seem unreasonable to suppose that the movement was damped in passing the mountain chains on its way to Pasadena. *Lg* of the Oklahoma earthquake is known to have been strong, not only at Palisades, but also at other stations in Northeastern America, and there are other indications of *Lg* being particularly well transmitted there. This may bear relation to the fact that the areas over which many of the shocks of Northeastern America are felt are large relative to their epicentral intensity (see e. g. Lehmann 1955).

The Pasadena long-period records are seen in fig. 9. In the N record, the transverse component, the short-period *Lg* waves are superposed on the long-period Love waves, but they are not much larger than in the *E* record; their onset, however, is more distinct. The phase builds up gradually and is read later in the longitudinal and vertical records. The Pasadena standard torsion seismometers also recorded *Lg* as seen in fig. 10. Using the earliest onset in the Pasadena records we find the velocity 3.58 km/sec.

Lg was clearly recorded at the other stations in Southern California, but the first waves were small and the onsets not very distinct. Velocities as calculated ranged from 3.49 to 3.53 km/sec.

In a number of records of the Oklahoma earthquake collected at Saint Louis for a special study *Lg* was found to be a very well recorded phase. Velocities as determined for Harvard and Pittsburgh were 3.58 and 3.57 km/sec respectively.

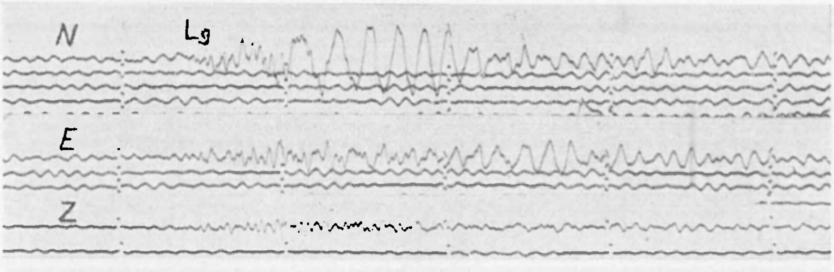

Fig. 9. – Pasadena long-period Benioff records of Oklahoma earthquake

In the large southern California earthquake of July 21, 1952, the Arvin-Tehachapi shock, *Lg* was very well recorded at many stations. Gutenberg (1955) finds the velocity 3.58 ± 0.02 km/sec. At Palisades, at an epicentral distance of 35°.7 the movement was still very strong, too strong indeed to be well recorded by instruments of all types. The 12 sec. electronic instrument had very large records. In the *N* record large swings of varying period continued from *S* onwards. *Lg* was taken to be the group of waves of period of about 3½ sec. marked by the second arrow in fig. 11. The period of the movement immediately preceding it is about 12 sec. The first arrow marks a group of waves of relatively short period, 5-6 sec. with surface velocity 3.9 km/sec. Sometimes in strong European

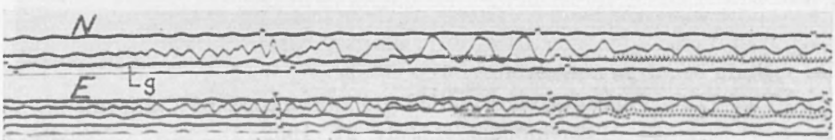

Fig. 10. – Pasadena torsion seismometer records of Oklahoma earthquake

earthquakes similar wave-groups with about the same velocity have been observed. Since the period of *Lg* in this case is relatively long, the amplitudes of the short-period Benioff records are not excessively large.

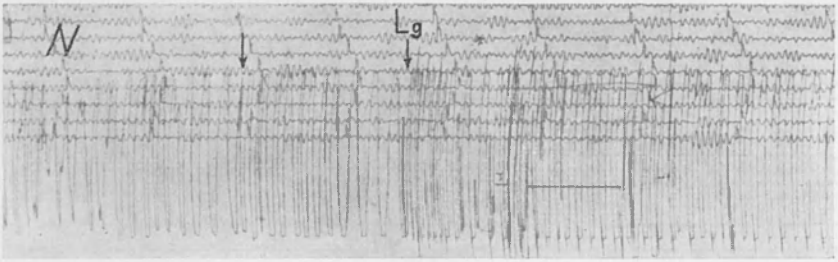

Fig. 11. – California earthquake of July 21, 1952 recorded at Palisades on 12ˢ electronic seismograph.

Fig. 12 shows the very clear Benioff short-period *N* record. The velocity as determined from this record is 3.59 km/sec. The

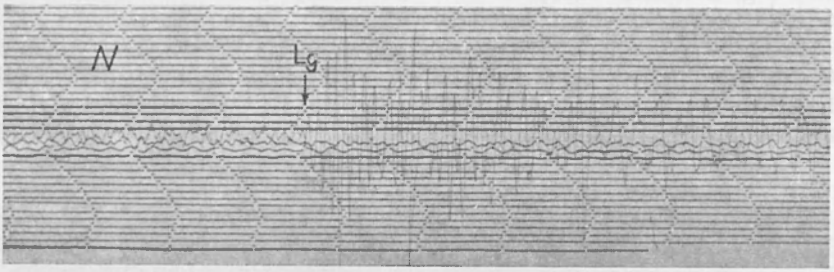

Fig. 12. – Palisades short-period Benioff record of California earthquake of July 21, 1952.

onsets are a little later on the *E* and *Z* records.

In the two large aftershocks of July 23. *Lg* was also recorded, but in most of the Palisades records the onsets were not so clear because *Lg* was preceded by movement similar to it. In the record of the 10ˢ,75ˢ vertical seismograph (fig. 13) *Lg* seems to arrive at the second arrow with the velocity 3.60 km/sec.; it is seen to ride on Rayleigh waves of long period. The first arrow marks the arrival of a group of small waves of period 6ˢ having the velocity 3.97 km/sec. It possibly corresponds to the early short-period waves of the main shock.

In these few examples we have seen that the period of *Lg* is not always the same and that one and the same shock may produce *Lg* waves of different period in different localities; the period, therefore, evidently depends on the path. The nature of the onsets was also found to differ, those of the Oklahoma *Lg* being much larger and sharper at Palisades than in Southern California. It would be of interest to study more closely how the character of *Lg* depends on the various factors which determine it.

* * *

During a stay at the Lamont Geological Observatory I was given the opportunity to study *Lg* in the records collected there of four North American earthquakes. The object was to see whether *Lg* could be distinguished in the various records, most of them from instruments the response of which to the short-period *Lg* waves was not particularly good, and, if so, to determine the velocity. While *Lg* was found to be clearly present in most of the records, the determination of the velocity of *Lg* met with some difficulty. The results first arrived at (Lehmann 1953) have here been modified.

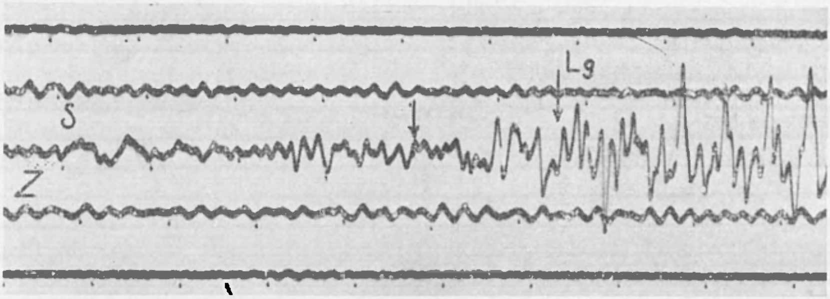

Fig. 13. – California earthquake, of July 23, 1952 recorded at Palisades on 10^s, 75^s vertical seismograph.

The earthquakes were:

The Timiskaming earthquake of November 1, 1935

the St. Lawrence earthquake of October 19, 1939

the Helena earthquake of October 31, 1935

the Grand Banks earthquake of November 18, 1929.

Since the velocity of *Lg* was wanted the epicentres and times of occurrence of the earthquakes had to be known. They were all in The International Seismological Summary (I. S. S.) and also in the " Seismicity of the Earth " (Gutenberg and Richter 1949) but the two sets of values did not quite agree. Besides, depths of 60 km and 40 km were assigned to the Timiskaming and the St. Lawrence earthquakes respectively in the " Seismicity of the Earth " while in the I. S. S. all the earthquakes were taken to be normal. This latter point required attention, for it is of importance for the understanding of the nature of *Lg* to find out whether earthquakes deeper than normal are capable of producing such strong *Lg* waves as those found to be recorded in these two earthquakes. A redetermination of their epicentres and depths was therefore attempted. In the course of the ensuing study it was found that the *P* and *S* travel times for small distances of these and other Northeastern American earthquakes differed from those of current tables but agreed with those found in explosion work and in the most recent studies of European earthquake travel times for *P* (Lehmann, 1955. For other references see this publication). Trial time-tables were therefore constructed and epicentres and times of occurrence adapted to them. There was then no longer any indication of depth greater than normal. The times of occurrence and epicentres adopted were:

Nov. 1,1935	6:03:35	46°.8 N 79°.1W
Oct. 19,1939	11:53:56	47°.8 N 70°.0W

We shall consider first *Lg* as recorded in these two earthquakes.

There were several good records available of the Timiskaming earthquake for epicentral distances from 2° to 13° and some for greater distances. Except at the smallest distances *P* is rather small, but it is clearly marked; *S* is much larger. Two distinct groups of surface waves were very well recorded at several stations, but at some the trace was too faint or had vanished because the movement was too strong. The

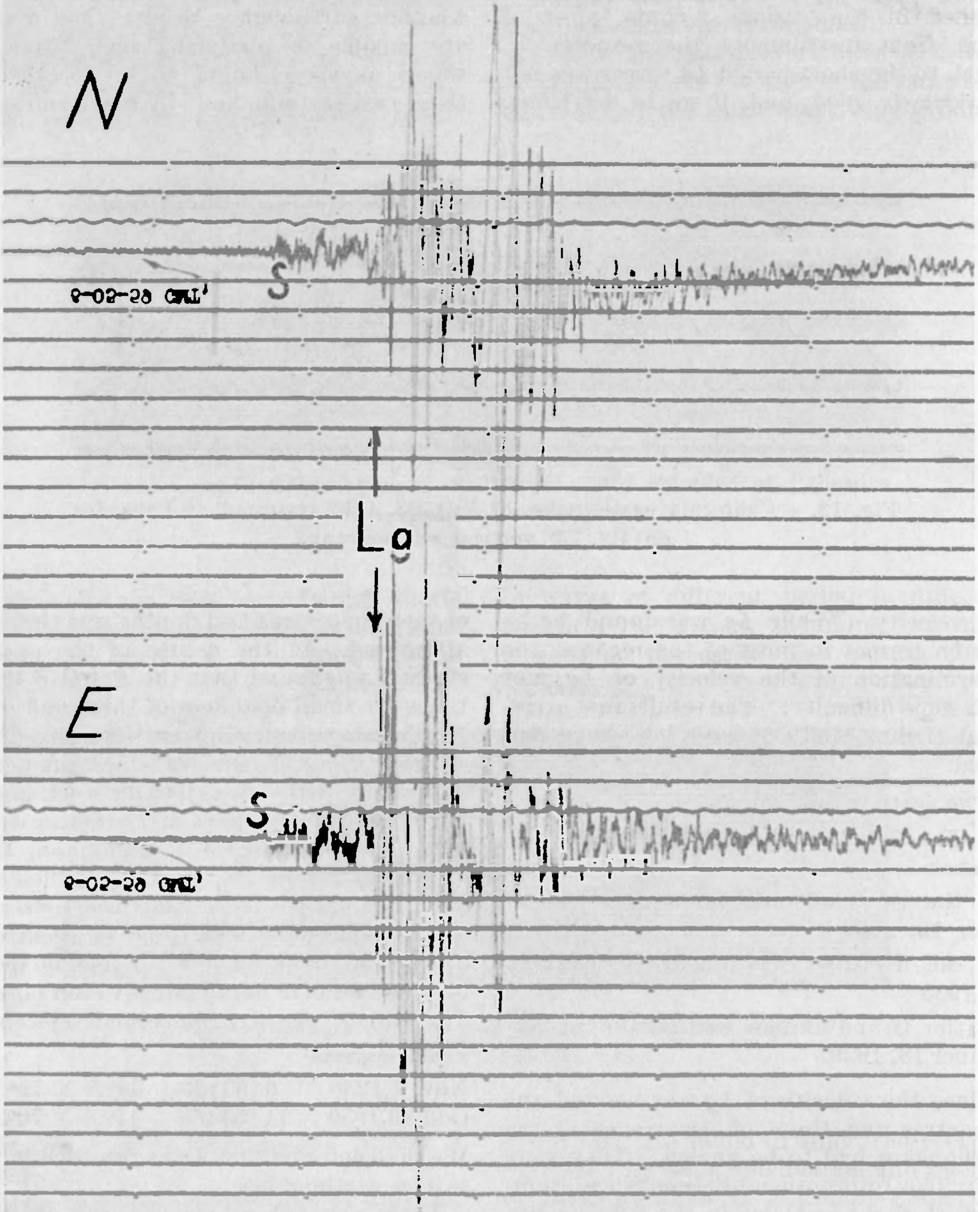

Fig. 14. – Philadelphia records of Timiskaming earthquake

Philadelphia record of fig. 14 was obtained at an epicentral distance of 7°.2. *S* is seen to be rather large and have a sharp onset. Subsequent to *S* the movement is somewhat irregular, but a little more than ½ min. after *S* large swings of period of about 2 sec. set in abruptly, and this is where *Lg* is assumed to have begun. The waves are larger on the *E* component record than on the *N* component, so, since the epicentre is NNW of Philadelphia, they are mainly transverse. Long-period waves, undoubtedly Love waves, are also present, but it is uncertain where they begin. The complex motion of the group subsides quickly so as to allow the second group, the Rayleigh wave group, to be clearly recorded. The period of the waves of this group is at first about 10^s, but it quickly decreases to 5^s.

At small distances the amplitudes of the short-period Lg waves are large as in the Philadelphia records, but they decrease with distance until they appear as ripples on the underlying long-period Love waves as in the Bozeman records of fig. 15. The epicentral distance is 22°. Bozeman is west of the epicentre and here we have the two groups of surface waves even better separated than in the Philadelphia records. In the Bozeman *N* record the Love waves are very clear and the Rayleigh waves form a large and regular group of waves in the *E* record. Short-period Lg waves are superposed on the long-period Love waves in the *N* record; they are also present in the *E* record, and they are, perhaps, more conspicuous here where the long-period movement is smaller.

The onsets of Lg were read in all the records and the travel times tabulated in table 1. They were plotted against epicentral distance and a straight line drawn " through " the points. Its slope was 31.1 sec/degree which corresponds to the velocity 3.57 km/sec. In the table the residuals O-C_1 are against a line through the origin having this slope. There are 5 very large residuals. That of Columbia is possibly due to incorrect identification; there is a 29 sec. later onset which may be the true Lg. At Florissant and Saint Louis large movement sets in where the readings were made, but the subsequent trace being too faint to be readable, Lg may have been taken too early. At Tucson and Seattle the short-period Lg waves form small ripples on the long-period waves, and the earliest swings may have been lost. 3 stations have 2 readings. At Buffalo and Philadelphia the earlier readings are from more sensitive instruments than the later ones, and indeed from more sensitive instruments than most of the observations at hand. At Charlottesville the earlier reading is from the E component instrument, recording transverse motion.

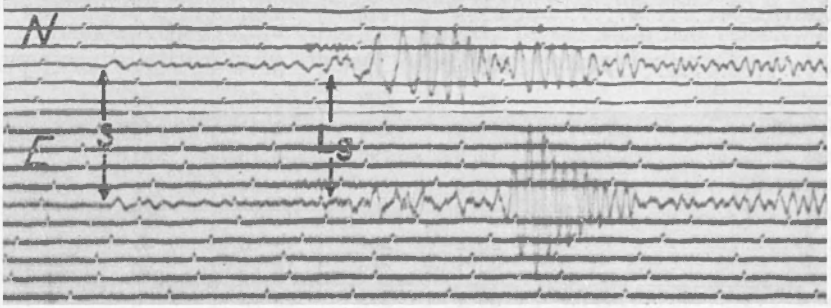

Fig. 15. – Bozeman records of Timiskaming earthquake

The 5 observations having very large residuals were neglected, and so were the early readings at Philadelphia and Buffalo and the late reading at Charlottesville. The remaining travel times were smoothed on a straight line by the method of least squares. The resulting line had a slightly greater slope than the one first found and it had its point of intercept below the origin. This could possibly indicate that the actual time of occurrence of the earthquake was earlier than the time adopted. This time was found by fitting the transmission times of P for small distances to the linear expression $7^s.37 + \Delta^\circ \times 13.58$ sec/degree (see Lehmann,

Table 1

1935, November 1, 6 : 03 : 35 46°.8 N 79°.1 W

Station	Δ	Az.	Lg	O-C	O-C	vel.
	°	°	m s	s	s	km/sec
Buffalo	3.9	177	1 51	–10		3.90
			59	– 2	1	3.64
Ithaca	4.8	155	2 26	– 3	– 2	3.65
Ann Arbor	5.6	218	2.9	0	1	3.58
Philadelphia	7.4	155	3 39	–11		3.75
			45	– 5	– 3	3.65
Chicago Loyola	7.9	234	4 10	4	6	3.51
Chicago Univ.	7.9	234	4 12	6	8	3.48
Charlottesville	8.8	176	4 27	– 7	– 4	3.66
			36	2		3.54
Halifax	11.1	95	5 39	– 6	– 3	3.64
Florissant	11.5	229	5 30	–28		
Saint Louis	11.6	229	5 45	–16		
Columbia	12.9	187	6 10	–31		
Saskatoon	18.6	297	9 34	– 4	0	3.60
Bozeman	22.0	280	11 27	3	8	3.56
Tucson	28.2	251	14 58	21		
Seattle	29.1	279	15 25	20		
Tinemaha	30.4	273	15 45	0	7	3.57
Sitka	35.2	309	18 18	3	12	3.56

Table 2

1939, October 19, 11 : 53 : 56 47°.8 N 70°.0 W

Station	Δ	Az.	Lg	O-C	vel.
	°	°	m s	s	km/sec
East Machias	3.5	153	1 48	0	3.60
Ottawa	4.6	243	2 21	– 1	3.62
Weston	5.5	190	2 46	– 4	3.68
Halifax	5.5	125	3 04	14	3.32
Fordham	7.5	205	3.9	3	3.56
Philadelphia	8.7	207	4 15	–13	3.79
Georgetown	10.3	213	5 19	1	3.59
Cleveland	10.4	236	5 11	–10	3.72
Cincinnati	13.6	236	6 55	– 5	3.64
Chicago Loyola	13.8	252	7 07	1	3.59
Columbia	16.1	215	8 32	15	3.49
Saint Louis	17.3	248	8 59	5	3.57
Lincoln	20.2	261	10 27	4	3.58
Bozeman	28.0	258	14 36	12	3.55
Tucson	34.5	260	17 59	15	3.55
Pasadena	38.2	268	20 11	33	3.50

1955, p. 353) but, while the observations clearly fitted a line of this slope, there was no way of ascertaining whether the constant term was correct for this earthquake. However, Hodgson's result for the same region being $t(Pn) = 7^s.50 \pm 0^s.11 + \Delta$ km/(8.176 + 0.013) km/sec (Hodgson, 1953, p. 146) our T_o probably is not far wrong. Since we cannot accept an *Lg* line having its point of intercept below the origin we neglect the observations for distances greater than 20° which may be late because the phase is small at these distances, and we then fmd by trial and error that the straight line of slope 30.85 sec/degree through the origin fits the remaining travel times fairly well. The residuals are the O-C_2 of table 1. The corresponding velocity is 3.60 km/sec, not very different from the one first found. The method of least squares was not applied since it does not necessarily yield the "best" result when the distribution of the residuals is not normal. The velocities as calculated from individual observations were tabulated. They are seen to vary a great deal, especially at small distances where a large residual affects the velocity more than at greater distances.

Instruments such as the Palisades electronic instruments having an exceptionally good response to the short-period *Lg* waves probably would have yielded more consistent readings than those here obtained. It has to be left to future studies to find with what accuracy the *Lg* velocity can be determined and whether, within the limits of accuracy, it is the same everywhere, independent of the path.

Hodgson in his rockburst studies is at a loss to explain the phase he calls S_1 (Hodgson, 1953, p. 160) as it appears in the records of his 5 " distant " stations, the nearest at a distance of 451 km, the most distant 942.7 km away. He mentions the possibility of its being *Lg* which, he thinks, would explain its erratic beginning and the nature of the phase. It appears as a group of short-period waves of rather large amplitude not readily explainable as being due to body waves. He hesitates to identify it with *Lg* because its velocity is higher than the average velocity 3.51 km/sec found by Press and Ewing (1952) from earthquake observations, but there seems to me to be no escape from the conclusion that it is a wave of the nature of *Lg*. Hodgson found the velocity 3.71 ± 0.08 km/sec, where 0.08 is the probable error, so it is not significantly higher than the velocity found for the Timiskaming earthquake.

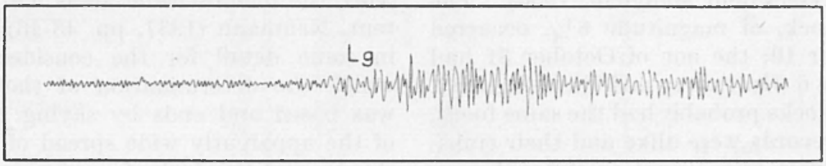

Fig. 16. – Tucson short-period record of St. Lawrence earthquake.

The St. Lawrence earthquake of Oct. 19, 1939 was much smaller than the Timiskaming earthquake. Only 4 stations at distances greater than 40° recorded *P*, and the forerunners were small also at the shorter distances. *Lg* is the largest phase in the records studied and at short distances its onset is sharp; some of the I. S. S. entrances are due to readings of this phase. The appearance of the records is peculiar, the period of the motion being unusually short throughout, in the forerunners and also in the surface waves. For this reason the trace is mostly faint and good reproductions not obtainable. Fig. 16 shows a tracing of the Tucson ($\Delta = 34°.5$) short-period vertical Benioff record. The phase is still quite large at this rather distant station. In all the records studied, also those of long-period instruments, the Rayleigh waves are small and not clearly separated from the first much larger group of surface waves.

The travel times of *Lg* are tabulated in table 2. The residuals are against $t(Lg)$ –

30.85 × Δ as taken for the Timiskaming earthquake. They scatter a great deal. but up to about 20° they have no apparent systematic trend. For greater distances the residuals are large and positive, possibly because small movement in the first part of the phase was lost. Actually, at Tucson, the phase is found to build up gradually. In fig. 16 the first part of *Lg* is not large, and there is a later increase. In another record there is a small beginning earlier than the one in this record. The velocities as calculated from individual observations were tabulated.

Lg was also recorded in the Helena earthquake of Oct. 31, 1935. It was read in the records of 10 observatories.

This earthquake was one of a swarm of earthquakes that began early in October 1935 (see Heck and Maughan, 1936). The largest shock, of magnitude 6 ¼, occurred on October 19; the one of October 31 had magnitude 6 (Gutenberg and Richter, 1949). The two shocks probably had the same focus, for their records were alike and their transmission times were equal or differed very little. The *P* and *S* transmission times tabulated in table 3 are mean values. The times are chiefly from the I. S. S., but some are from Gutenberg and Richter (1938) and a few are my own readings. When the difference of the times of the two shocks did not exceed 2^s, a × was affixed to the tabulated mean; if the difference exceeded 4^s, the mean was put in brackets. The suffixes 1 and 2 indicate that only readings of the first or the second shock respectively were available.

The two shocks were highly destructive in Helena, so the focus was evidently quite close to the city. F. Neumann (1937) determined the epicentre 46° 37' N, 111°58'W. For the I. S. S. the coordinates were rounded off to 46°.6N, 112°.0W, and this is the position here adopted. As Neumann points out, there is no way of determining the epicentre very accurately. There is only one near station, Bozeman; the next, Seattle, is 7° away, and the distribution in azimuth of the more distant stations is quite unsatisfactory. Neumann therefore based his determination on the records of the strong motion accelerograph operating in Helena from October 21. It had an automatic starter and was set working by the preliminary tremors of the large shock of October 31, but it did not record the onset of *P*. Once started it kept working for some time and therefore gave complete records of a number of aftershocks. These were found to have their epicentres from 3 to 6 km from the accelerograph and focal depths from 2 to 5 km. Because the small, completely recorded shocks, 10 in all, were so shallow it was concluded that the large shocks were also quite shallow. It seems questionable, however, whether this conclusion is warranted. There is nothing in the records to indicate that the shocks are exceptionally shallow; they look quite " normal ", and forerunners of both shocks were recorded out to an epicentral distance of 94°. Also, the macroseismic areas had great extent. Neumann (1937, pp. 43-46) accounts in some detail for the considerations on which the determination of the epicentre was based and ends by saying: " In view of the apparently wide spread of the aftershocks and the lack of precise information concerning the main shocks of the series an epicenter at 46° 37'N, 111° 58'W is adopted as representing as near as we know it, the central point of activity ". The Bozeman *Sg–Pg* was found to be $17^s.0$, and this corresponds closely to the distance from the adopted epicentre when the velocities of *Pg* and *Sg* are taken to be 5.5 km/sec and 3.2 km/sec respectively. The velocities are now believed to be higher, and the reading of *Sg* has evidently caused some difficulty, for different readings were reported later by Gutenberg and Richter (1938) and in the I. S. S. The Bozeman observation, therefore, cannot be said to give strong support to the determined epicentre. This however, is probably not far wrong.

We need also the times of occurrence, and the values obtained for them depend on the tables used. For the shock of October 31, Neumann has 18 : 37 : 47, Gutenberg and Richter 18 : 37 : 49, both values being derived from wave velocities at small distances no longer believed to be correct. I. S. S. has T_o = 18 : 37 : 56, but this is not supposed to be the time of occurrence. Taking $t(Pn) = 7^s.37 + \Delta^\circ \times 13.58$ sec/degree as for

Table 3

1935 October 19. 4 : 48 : 11

1935 October 31. 18 : 37 : 35

Station	Δ	Az.	P (mean)		O-C$_1$	O-C$_2$	S (mean)		O-C$_1$	O-C$_2$
	°	°	m	s	s	s	m	s	s	s
Bozeman	1.1	144		16×				37×		
Seattle	7.1	282	1	43_0		–1	3	(9)	5	9
Victoria	7.9	288	1	54_1		–1	3	17_2	–7	–3
Denver	8.5	141	2	8×		5	3	30	–9	–4
Tinemaha	10.6	208		31×		0				
Ukiah	11.1	232		38×		0	4	39×	–4	3
Haiwee	11.4	205		42×		0	5	3_1	13	19
Berkeley	11.6	224		45		0		(1)	6	13
Lick	11.7	221		48_1		2		4_2	6	13
Santa Clara	11.8	221	2	46_1		–2		8_1	8	15
Mt. Wilson	13.2	203	3	8×		1				
Riverside	13.2	200		7		0				
Pasadena	13.3	202		9×		1				
Santa Barbara	13.5	209		10_1		–1				
La Jolla	14.3	198		23×		1				
Tucson	14.4	176		19		–4	6	(10)	7	
Madison	16.4	94		48_2	1		6	55_2	5	
Florissant	17.7	108	3	57×	–5		7	16×	–3	
Sitka	17.7	315	4	(3)	–1			31×	12	
Saint Louis	17.9	109	3	58_1	–8			27_1	3	
Chicago	18.1	96	4	1×	–7			24×	–5	
Little Rock	19.0	122		(17)	–3		7	(50)	1	
Ann Arbor	20.6	92		34×	–3					
Cincinnati	21.4	100	4	43_1	–2		8	47_1	8	
Toronto	23.1	86	5	3×	1		9	12×	2	
Buffalo	23.7	94		8×	0			30×	9	
Ottawa	25.1	79		21	–1			(42)	–3	
Ithaca	25.5	85		28×	2		10	9	18	
Charlottesville	26.0	97								
Georgetown	26.6	95		34×	–2			5_1	–4	
Columbia	26.6	106						1_1	–8	
Vermont	27.1	80						15_2	–3	
Philadelphia	27.4	91		(45)	1			21×	–2	
Fordham	27.9	90		50_1	2			26_1	–5	
Harvard	28.9	82	5	53×	–4			49×	2	
Huancayo	67.0	140	10	48×	–3		19	33_1	–11	
Pulkovo	69.6	19					20	14_2	–1	
Stuttgart	72.1	36	11	21×	–1					
La Paz	74.2	136		(38)	4					
Vladivostok	74.3	316					21	9_2	1	
Granada	74.8	52	11	43_2	5					
Sverdlovsk	76.8	4					21	35_1	–1	
Chiufeng	83.6	325					22	49_2	2	
Grozny	88.3	16	12	43_2	–6					
Nanking	89.3	319					23	41_0	–1	
Tiflis	89.7	17	12	57_2	2		23	35_2	–11	
Tashkent	92.5	359	13	9_2	1		24	13_2	2	

the Timiskaming earthquake and fitting the Helena P transmission times for distances up to 14°.4 to it, we find for the T_o (= time of occurrence) of October 19, 4:48:11 and of October 31, 18 : 37 : 55. The resulting residuals are tabulated in table 3 under the heading O-C_2. They are very satisfactory, but there being no Pn observations for distances smaller than 7°, we do not know whether our straight line would fit also at the smallest distances. Actually the observations at hand fit the Jeffreys-Bullen tables just as well, but T_o has then to be taken 5^s earlier. Bozeman has probably recorded Pg. Taking the velocity of this wave to be 6.2 km/sec, the observation as quoted is about 4^s early and T_o should have to be taken 4^s earlier to fit it. However, neither the position of the epicentre nor the Bozeman observation is likely to be quite correct. We shall, therefore, leave our $T_o s$ as determined above, but have to admit that the uncertainty is considerable.

Although it is somewhat outside the scope of the present investigation we shall consider also the travel times of P for greater distances and the travel times of S. In table 3 the P residuals for distances greater than 16° are against the Jeffreys-Bullen (J-B) travel times for a surface focus minus 6^s, and so are all the S residuals headed O-C_1, whereas the O–C_2 are against the S times determined from $t(S) = 10^s + \Delta^o \times 24.0$ sec/degree found to fit the travel times up to 14° in Northeastern America. For small distances neither the one nor the other of the two sets of S residuals are at all satisfactory. For greater distances the P times have quite a good fit except in the range 17°.7 to 21°.4 where all the residuals are negative and some of them surprisingly large. The S residuals have much greater scatter, but no apparent systematic trend. Thus from about 22° onwards the Helena S–P agree, on the average, with the J-B S–P for a surface focus, whereas the S–P of Northeastern American earthquakes were always greater than the S–P of the same tables (see Lehmann, 1955, p. 365). This seems contradictory to the conclusion that the Helena shocks are exceptionally shallow.

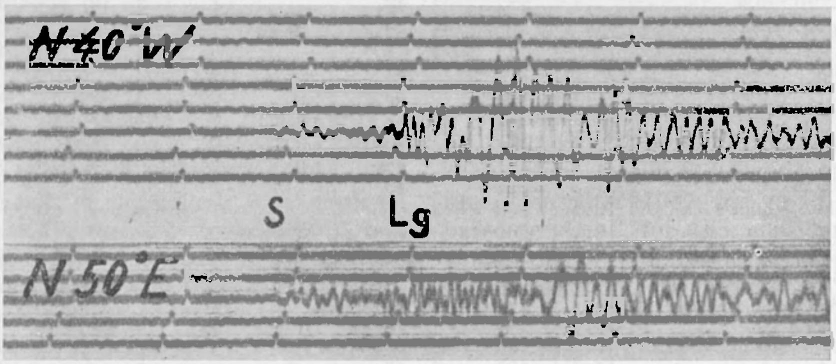

Fig. 17. – Ukiah records of Helena earthquake.

Lg was clearly present in the records studied. In the long-period records the short-period Lg wave arrived riding on the long Love waves. These sometimes began earlier as in the Pasadena record of fig. 1, and sometimes almost simultaneously with Lg. The Ukiah records are seen in fig. 17. The epicentral distance is 11°.1. The seismographs have orientation N 50°E and N 40° W. The azimuth of the epicentre in Ukiah being 59°, the latter component records transverse motion, the other one longitudinal motion. As a result the Love and Rayleigh wave groups are very clearly separated. Lg rides on long waves in the transverse component, but is more conspicuous in the longitudinal component where it is undisturbed. Fig. 18 shows a short-period Berkeley E record of the earthquake. Here *Lg* is clearly recorded with a well defined onset.

The *Lg* travel times are tabulated in table 4. When they are compared with the times calculated from $t(Lg) = \Delta^o \times 30.85$ sec/degree as taken for the other two earthquakes, we find the residuals headed *O–C*. They are as erratic as those of the other shocks. Individual velocities were calculated. Their mean is 3.59 km/sec while the velocity determined by the slope of the straight line representing the travel times is 3.60 km/sec. The "standard error" calculated in the usual way is 0.02 km/sec, but neither this quantity nor the mean itself has much significance since the distribution of the residuals shows no approach to a normal distribution.

In spite of the fact that the velocity as determined for the three shocks has no great accuracy there is no doubt about its exceeding the value 3.51 km/sec found by Press and Ewing. It is closer to the mean velocity found for the Southern California shock of 1952 (Gutenberg, 1955) and to the best determined velocities of the Oklahoma shock.

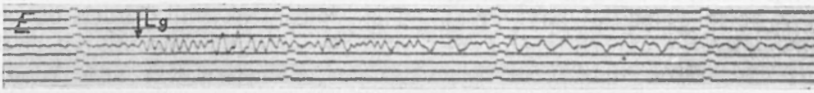

Fig. 18. – Berkeley short-period record of Helena earthquake.

There were only 5 records available of the Grand Banks earthquake of November 18, 1929 and they were not very useful for *Lg*. It was a stronger earthquake than the others considered and in the photographic records of the nearer stations the trace vanished where large swings commenced and *Lg* could not be distinguished. Ann Arbor at a distance of 20°.3 had a mechanically recording instrument, but the records were peculiar with short-period movement throughout and no distinct *Lg*. The Tucson *N* record of fig. 19 is from a distance of 44°.0, and here surface waves of short period are very conspicuous. Since Tucson is east of the epicentre it is the transverse component record. On the *E* record the onset is somewhat later but the short-period waves are also large. The epicentre of this earthquake could not

Table 4

1935, October 31, 18:37:55

Station	Δ °	Lg m s	O-C s	vel. km/sec
Victoria........	7.9	3 56	– 8	3.72
Ukiah	11.1	5 49	7	3.53
Berkeley	11.6	6 5	7	3.53
Pasadena	13.3	6 55	5	3.56
Tucson	14.4	7 29	5	3.56
Sitka	17.7	8 50	–16	3.71
Chicago	18.1	9 25	7	3.56
Toronto........	23.1	11 45	– 8	3.64
Ottawa	25.1	13 10	16	3.53
Columbia	26.6	13 45	4	3.58

be determined with much accuracy. Taking the position 44° N, 56° W and T_o = 20 : 31 : 58 as given in " The Seismicity of the Earth " we find that *Lg* as marked in fig. 19 arrives with the velocity 3.76 km/sec. There is a strong later increase of amplitude and the corresponding velocity is 3.48 km/sec. However, these velocities have great uncertainty owing to the uncertainty of the epicentre and T_o. Copenhagen is at very nearly the same epicentral distance as Tucson, and it is interesting to find that it has no trace of short-period surface waves, but records rather large regular Love waves of long period.

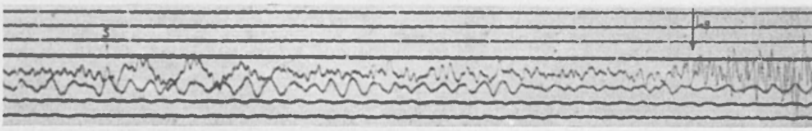

Fig. 19. – Tucson record of Grand Banks earthquake

* * *

Several other records obtained in North America were examined for *Lg*. At Pasadena a selection of records of deep Mexican earthquakes made for another study was examined. This was of interest in connection with the question as to whether or not *Lg* is present in records of deep shocks. The phase was not found in any of the records examined.

Some shocks originating in the Gulf of California were found to have very clear *Lg* in the Pasadena and other Southern California records. The onset of the phase, however, was not always distinct because the amplitudes of the first swings were small and there were short-period waves similar to the *Lg* waves in the movement immediately preceding the phase (most of the records are from short-period instruments). *Lg* was read in some of the records of the Southern California network of stations for the following earthquakes:

1. 1945, June 27, 13 : 08 : 20, 26°.8 N 111°. 8 W, M = 7.0

2. 1954. April 29, 10 : 49 : 27, 29°½ N 112°½ W, M = 7¼ — 7½

3. 1954, April 29, 11 : 34 : 34, 29°½ N 112°½ W, M = 7½ — 7³/₄

4. 1954, May 5, 13 : 09 : 46, 27°½ N 112°½ W, M = 6³/₄

The epicentres and times of occurrence of these earthquakes could not be determined very accurately because the recording stations and especially the near ones were not well distributed in azimuth. This may account for errors in the velocities as determined for *Lg* but not for the spreading of the values as found for stations that are all very nearly in the same azimuth. For the first earthquake *Lg* was read in the records of 6 Southern California stations and the velocities varied from 3 35 to 3.51 km/sec. For the second and third earthquakes the variation was still greater, from about 3.2 to 3.5 km/sec. The *Lg* onsets of the first of the two earthquakes of April 29, 1954 were smaller than those of the second and the readings more uncertain. There is a remarkable and very interesting difference between the records of these two shocks. The time difference between them is 45 min. Therefore corresponding phases are written immediately below each other by instruments having recording speed 60 mm/min. and this facilitates comparison. The Haiwee *N* record ($\Delta = 8^o$) is seen in fig. 20. *P* and the subsequent movement are equally big in both earthquakes, but about 2 min. after *P* long waves of period 25 sec arrive in the second earthquake while small movement continues in the first; the long waves increase and rise to large amplitudes. *Lg* sets in sharply $2^m\ 27^s$ after *P* in the second shock; in the first there is a small corresponding increase of movement which may and may not be *Lg*, while there is a clear onset of *Lg* waves 10 sec. later and again an increase of short-period waves 20 sec. later. For a little more than 3 min. the movement as recorded in the two earthquakes differs greatly, then the long-period waves of the second earthquake subside and the two records are again very much alike, the one no stronger than the other. This raises the question as to what is responsible for the production of long waves and *Lg*, the structure of the layer in which the earthquake occurs or possibly the mechanism of the shock. The two shocks here considered have been taken to have the same epicentre and normal depth, but small differences in location are not excluded. In the shock of May 5, 1954 we have again clear *Lg* waves but somewhat uncertain onsets and the velocities found vary a great deal.

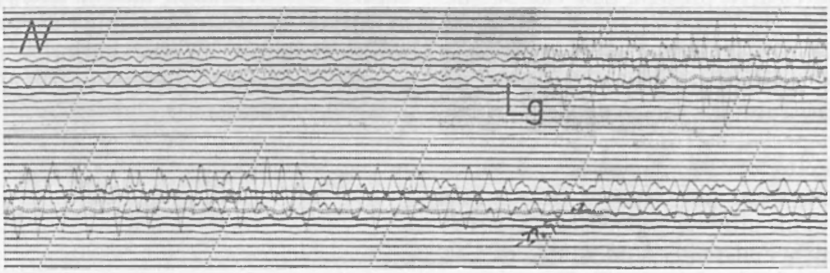

Fig. 20. – Two Gulf of California earthquakes recorded at Haiwee

Other Pasadena records were examined for *Lg*, and in normal shocks the phase was usually present, but as in the Southern California shocks, it often grew out of a background of short-period movement which resembled it a great deal except for being less regular. Then the readings were uncertain and the velocities as determined varied considerably. There were, however, also earthquakes in which the first *Lg* movement was large and the onsets sharp as in the Nevada earthquake of Febr. 8, 1940, 8^h. *Lg* was read in the records of 4 Southern California stations at distances from 4°.5 to 6°.5. The wave period was about 1^s. The Gutenberg and Richter (1949) epicentre and T_o were taken and *Lg* velocities from 3.27 to 3.39 km/sec found. However, when the *P* observations were fitted to the linear expression used earlier (see p. 11), a later T_o was found and the corresponding *Lg* velocities ranged from 3.42 to 3.50 km/sec.

The earthquake of Dec. 14, 1950, 13^h originated in the same region, and it also had clear *Lg*. Several records filed by the Seismology Branch of the U. S. Coast and Geodetic Survey at Washington were examined for *Lg* and good readings obtained for 7 stations at distances ranging from 700 km (Salt Lake City) to 3620 km (Washington). Butte at a distance of 900 km was the only station which had no clear *Lg*. The provisional epicentre and time of occurrence were taken and velocities ranging from 3.39 km/sec (for Washington, the most distant station) to 3.54 km/sec were found.

In the Manix earthquake of April 10, 1947 *Lg* was also recorded by the U. S. Coast and Geodetic Survey stations. Salt Lake City at a distance of 6°.8 had an *Lg* which increased gradually to become very large; it was found to arrive with the velocity 3.65 km/sec. At Lincoln (16°.7) and Chicago (23°.7) the *Lg* group was small and of short duration but clearly marked. The velocities found were 3.60 and 3.58 km/sec respectively. At Butte (11°.4) and Rapid City (13°.7) the onsets were less clear and the velocities as determined 3.41 and 3.45 km/sec respectively.

Very large *Lg* waves of quite short period (less than 1^s) were found in some of the U. S. Coast and Geodetic Survey records of the Montana earthquake of Nov. 23, 1947, but in many other records the phase was unreadable because the movement had been so strong as to make the trace vanish. The velocity found for *Lg* at Sitka (18°.2) was 3.38 km/sec, but microseismic movement masked the phase. The velocities found for three other stations were 3.42, 3.48 and 3.55 km/sec.

All in all a considerable number of North American records were examined for *Lg*. In normal shocks the phase was usually found to be clearly present and this is the chief result of the last section of this work. As to the velocities they are not very reliable for various reasons. Most of the records were from instruments the response of which to the short-period waves is not particularly good, and many of the epicentres and times of occurence were not known with much accuracy. This affects velocities at small distances rather strongly and many of the records were obtained at small distances.

A great many København records were also examined for *Lg*. It is intended to publish the results of this study later, but mention may be made here of the impression gained that on the whole the København *Lg*'s are less conspicuous than those recorded in North America; moreover they are quite often completely lacking also on continental paths. This is true also for Uppsala (see Båth, 1954). The average *Lg* period is longer in København than in North America. The periods of the *Lg*'s here studied have not often been given because they could not be measured with much accuracy, but they were mostly quite short, from about ½ sec to 3 sec. At København many *Lg*'s had a period of about 6 sec and most of the periods were in the range 4 sec to 6 sec. This is in accord with the result obtained for Uppsala where the maximum frequency is at 5-6 sec (Båth, 1954, p. 306). At Uppsala only Wiechert records were available, but at København there were in addition to records from long-period instruments many records from the short-period Benioff vertical seismograph. *Lg* was rarely clearly present in them while in North America good *Lg* records were obtained on these instruments (see e. g. figs. 8 and 16). Thus the difference observed in the periods is not of instrumental origin. At København there is often a background movement of a period close to that of *Lg* and this masks the phase, sometimes severely so. It may be difficult or perhaps impossible to say whether *Lg* is present.

While the well-determined velocities of North American *Lg* belong to the range 3.5-3.6 km/sec the velocities of the København *Lg* were usually found to be smaller. Whether this is significant cannot be said, for precise velocity determinations require well recorded *Lg*'s with clear onsets, and epicentres and times of occurrence more accurate than those available for the majority of Euroasiatic shocks (see Lehmann 1949).

A closer study of the phase *Lg* and precise determinations of its velocity should yield interesting and important results.

* * *

The work here reported on was done during a visit to seismological observatories in the U. S. A. and chiefly during my stay at the Lamont Geological Observatory, Palisades. I am indepted to Professor Maurice Ewing for bringing about this visit, for suggesting the subject of my research and for extending to me useful information and advice. My visits to the other observatories were made profitable through the kind way in which their facilities and records were made available to me. My thanks are due, in particular, to Dr. Beno Gutenberg, Dr. C. F. Richter, Dr. Perry Byerly, the late Reverend James B. Macelwane, Captain E. B. Roberts and Mr. L. M. Murphy.

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