GEOPHYSICS, VOL. XX, NO. 2 (APRIL, 1955), PP. 283-294, 5 FIGS.

CHANNEL WAVES IN THE EARTH'S CRUST*

B. GUTENBERG†

ABSTRACT

Three major low-velocity layers seem to exist in the earth's crust, of which two, the lithosphere channels, are found respectively in the "granitic" and "basaltic" ("gabbro") layers of the continents; a third channel extends from the Mohorovičić discontinuity downward into the asthenosphere. Several types of waves are guided by these channels, especially *Pa* and *Sa* by the asthenosphere channel, *Lg1*, *Lg2* and *Rg* by the lithosphere channels; waves guided by low-velocity layers in sediments must also be expected. Many records of the Southern California earthquake of July 21, 1952 show channel waves with periods and velocities as reported for other paths. The regular microseisms with periods of 4 to 10 sec have properties similar to those of the *Lg—Rg* group in earthquake records and are probably propagated by the same mechanism.

Most of these interpretations and conclusions are tentative; pertinent observations are scanty; and complications have thus far prevented development of adequate equations to calculate the amplitudes of waves guided by a given channel.

INTRODUCTION

The terms "channel wave" or "guided wave" indicated originally a body wave which, according to the classical ray equation based on Snell's law, is travelling completely inside a low-velocity layer (for example, *Lg2* in Figure 4). Frequently, surface waves which are appreciably affected by a low-velocity layer situated within a few wave lengths below the surface are also called channel waves. In the present paper this wider use of the expression "channel waves" is employed, especially since it is not yet certain which of the waves under consideration are body waves and which are surface waves. To a first approximation, the classical theory for surface waves is applicable to crustal structures which include a low-velocity layer. However, special theoretical investigations with better approximations are needed for the discussion of, and especially for the calculation of amplitudes for, channel waves of the body-wave as well as of the surface-wave types.

From various observations of amplitudes and travel times of body waves, Gutenberg (1954, with references to earlier papers) had concluded that there are at least two groups of low-velocity channels in the earth's crust. Above the Mohorovičić discontinuity are the "lithosphere channels" (see Figure 4, left) consisting probably of one channel in the so-called granitic layer,[1] another in the

* Manuscript received by the Editor December 6, 1954. Contribution No. 707, publications of the Division of Geological Sciences, California Institute of Technology, Pasadena, California.

† Seismological Laboratory, California Institute of Technology, Pasadena, California.

[1] The so-called granitic layer probably includes rock types differing appreciably from granite, possibly even basalt, and may be subdivided by discontinuities (see, e.g., Gutenberg, 1955). The layer below it extending downward to the Mohorovičić discontinuity is called basaltic layer by some, gabbro layer by others. Below the Mohorovičić discontinuity begins the "mantle" ("ultra-sima"). The lithosphere is the relatively strong outer part of the crust, the asthenosphere the deeper part where flow processes are supposed to make possible restoration of isostatic equilibrium.

"gabbro" layer; below the Mohorovičić discontinuity is the "asthenosphere channel," extending downward to a depth of about 150 km for longitudinal waves and to about 250 km for transverse waves (Gutenberg, 1955). Detailed investigation of these low-velocity layers is rendered difficult since their main effect consists in production of shadow zones. They were first suspected when it was observed that at certain distances the supposedly direct longitudinal or transverse waves arrive with very small amplitudes, also when conclusions based on various types of studies were found to contradict each other when observations were interpreted on the assumption that no low-velocity layer exists. The channel waves discussed in the present paper are among the rare instances of observable waves which are a consequence of low-velocity layers.

Press and Ewing (1951, 1952) have discovered two types of guided waves traveling in the "granitic" layer of North America, and have started the investigation of channel waves which will probably play an increasing role in the interpretation of records of earthquakes and of artificial explosions. Press and Ewing find that one type of the waves, *Lg*, has periods of 1 to 10 sec and a velocity of 3.5 km/sec, while the other, called *Rg* by them, which has periods of 8 to 12 sec, travels with an average velocity of 3.0 km/sec. Lehmann (1953), and Ewing, Press, and Oliver (1954) have extended these results. Detailed observations for Eurasia are reported by Båth (1954). He comes to the conclusion that *Lg* consists of at least two different wave types, *Lg1* and *Lg2*, and agrees with Gutenberg (1954) that the lithosphere channels make possible their propagation. Prevailing periods in Uppsala are 3 to 8 (occasionally 2 to 11) sec for *Lg1*, 3 to 12 sec for *Lg2*, and 3 to 14 (occasionally 16) sec for *Rg*.

Two additional types of channel waves not related to the preceding were first found by Caloi (1953) on records of earthquakes with focal depths of the order of 100 km. He called one *Pa* (periods between 5 and 12 sec, velocity about 7.9 to 8.0 km/sec), the other *Sa* (periods 10 sec or more, occasionally up to 30 sec, velocity about 4.4 km/sec). Press and Ewing (1954) have found these waves independently and have explained them as *Pn* and *Sn* waves with "numerous multiple reflections at near-grazing incidence from the underside of the Mohorovičić discontinuity, in direct analogy with the whispering gallery effect discussed by Rayleigh." However, the fact that Caloi, Press, and Ewing (1955) and the author have observed them in shocks with intermediate focal depth is in favor of Caloi's explanation (given independently by Gutenberg in a discussion of the results by Press and Ewing, 1954) that they are waves guided by the asthenosphere low-velocity channel which had been established earlier.

The question has been raised if sufficient energy from waves in channels at some depth below the earth's crust can be observed at its surface. The waves *Lg* and *Rg* through the lithosphere channel with its upper boundary at a depth of the order of 10 km have usually periods of 3 to 8 sec and a velocity of $3\frac{1}{2}\pm$ km/sec, resulting in wave lengths of the order of 10 to 25 km, so that the earth's surface is roughly one wave length or less above the boundary of the channel.

The same holds for the *Sa* waves through the lithosphere channel with its boundary probably 40 or 50 km below the surface, since these waves have periods which are usually between 10 and 20 sec and a velocity of about $4\frac{1}{2}$ km/sec. For sound waves through the main channel in the ocean, large amplitudes have been observed under similar circumstances, and elastic (sound) channel waves through the ocean as well as through the atmosphere have been generated by sources close to but outside the channel.

In the present paper, findings concerning various channel waves are discussed on the basis of records of the main Kern County, California earthquake of July 21, 1952, 11h 54m 14s, of some Pasadena records from other shocks, and of results published by various authors (see references).

TRAVEL TIMES AND VELOCITIES OF WAVES GUIDED BY THE LITHOSPHERE CHANNELS (*Lg*, *Rg*, $\bar{S}$)

At relatively short distances, the various waves guided by the lithosphere channels can rarely be studied in detail on records of short-period instruments as a consequence of their large amplitudes and of other phases with short periods which are superimposed. On the other hand, long-period instruments emphasize the long-period surface waves arriving at about the same time. Consequently, *Lg*-waves are usually clearest when recorded on medium-period instruments where, frequently, they are riding on top of longer surface waves (Figure 1).[2] On the records of the 1952 California earthquake written at Guadalajara and Puebla, which were marked for Figure 1 before the results of Båth (1954) were known, the "*eLg*" corresponds to *Lg1* of Båth, the "*Lg*" to his *Lg2*. In the other records of Figure 1, the phase marked "*Lg*" is his *Lg1*.

Arrival times of *Lg1* were established on seismograms of the main 1952 shock at 17 stations, all on the North American continent, except for Kingston, Jamaica. The times for the latter fit well, but the recorded motion may belong to some other phase. However, no *Lg*-waves were found at other stations outside of the North American continent. Since there is some doubt whether or not the large short-period "*iM*" phases observed at distances between 10° and 14° in the 1952 California shock are *Lg* waves, two least square solutions were made, one with inclusion of four such instances, the second without them. The resulting equations for the travel time t of *Lg1* to the epicentral distance Δ (measured in km) are:

All data: $$t = (4.6 \pm 4.9) + \Delta/(3.58 \pm 0.02) \quad (1)$$

Only $\Delta > 15°$: $$t = (9.5 \pm 10.0) + \Delta/(3.60 \pm 0.04). \quad (2)$$

Since the two equations agree within the limits of error, it is not unlikely that the short-period "*iM*" at distances of about 10° is actually *Lg1*. Observations of travel times of *Lg1* (zero line given by $t = 4 + 31\ \Delta°$ to permit use of a large scale) are plotted in the upper part of Figure 2, together with straight lines representing

[2] Figures have been arranged and drafted by Mr. J. Nordquist.

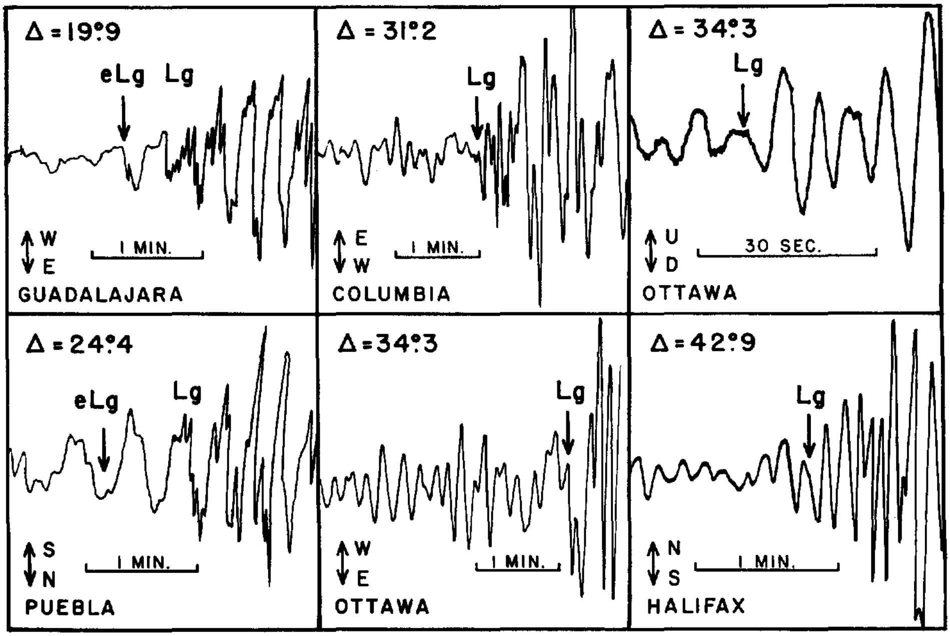

FIG. 1. Portions of selected records of the main Kern County, California, earthquake of July 21, 1952, showing *Lg* waves. Δ is the epicentral distance in degrees. Arrows indicate direction of ground motion.

equations (1) and (2). The velocities of 3.58±0.02 km/sec (equation (1)) and 3.60±0.04 km/sec (equation (2)) agree with those found by Press and Ewing (3.51±0.07), by Lehmann (3.57), and by Båth (3.54±0.07), especially if the effect of the constant term used in equations (1) and (2) is considered. The velocity of transverse waves near the top of the "granitic" layer in Southern California is 3.67±km/sec (Gutenberg, 1951, p. 145).

The phase called $\overline{S}$ or *Sg* at epicentral distances over 150±km (but not at distances less than 120±km where it is the direct transverse wave indicated now by *s*) is possibly *Lg*. On seismograms recorded in Southern California at distances between about 150 and 500 km, "the phase previously called $\overline{S}$ corresponds roughly to the beginning of the maximum in the *S* group. . . . It seems to move to later and later phases as the distance increases" (Gutenberg, 1951, p. 161). This group creates "the impression of waves traveling in a dispersive medium" and has been considered to be a wave guided by a low-velocity channel. At distances of about 300 km, a phase of the $\overline{S}$-group, called *Se*, is usually the dominant $\overline{S}$-phase in seismograms written in Southern California, but at distances beyond about 500 km, another, called *Sf*, carries the largest amplitudes. The travel time curves of these two waves are given (*l.c.*, p. 161) by

$$Se: \quad t = 4.9 + \Delta/3.64 \qquad Sf: \quad t = 4.3 + \Delta/3.52. \qquad (3)$$

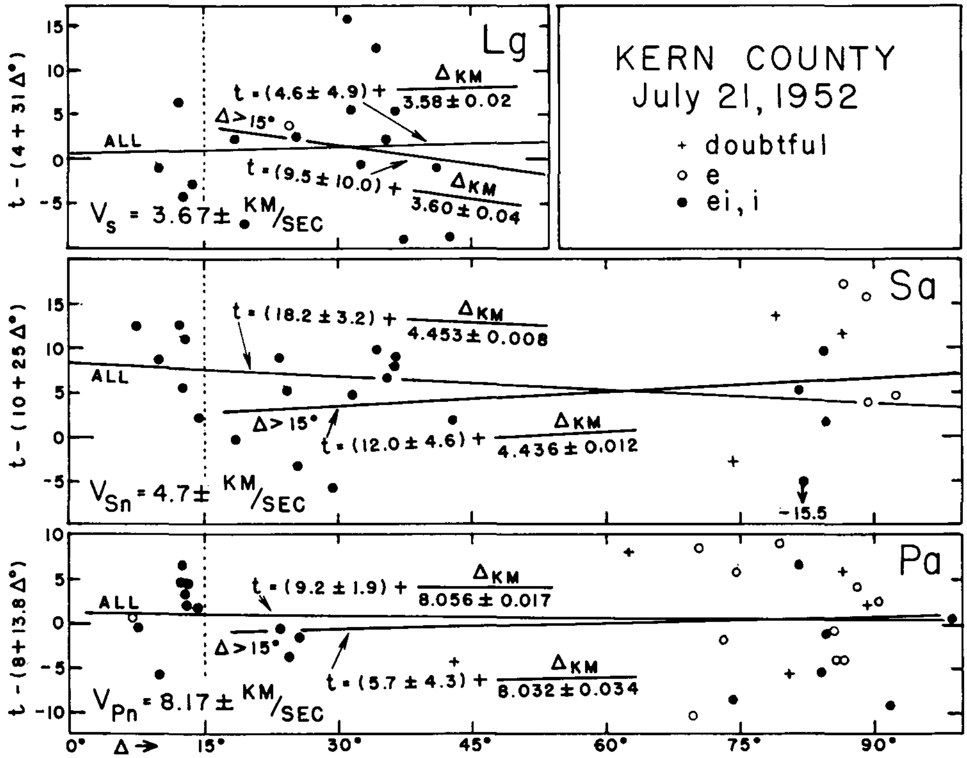

FIG. 2. Observed travel times t of *Lg*, *Sa*, and *Pa*. Quantities $a+b\Delta$ are subtracted from t to enable the use of a large time scale. The values of the constants a and b differ for each phase. Δ is the epicentral distance in degrees. Straight lines correspond to least-square solutions discussed in the text.

The velocities of *Se* and *Sf* agree well with those found for *Lg1* and, at distances of a few degrees, the $\overline{S}$-group may be identical with the *Lg*-group; the phase "*iM*", mentioned above for the 1952 California shock, may be the continuation of the $\overline{S}$-group at distances of about 10°.

Since there are not sufficient travel times for *Lg2* from records of the 1952 California shock to justify inclusion of a constant term in a solution using the method of least squares, it was assumed that the constant term is zero. The resulting velocity is:

$$v = 3.38 \pm 0.03 \text{ km/sec.} \qquad (4)$$

The velocity found by Båth for *Lg2* in Eurasia is 3.37 ± 0.04 km/sec.

Relatively few *Rg* waves can be identified on records of the 1952 California shock, partly as a consequence of the large motion, and most of the identifications are doubtful. The velocity of *Rg* found by Press and Ewing is 3.05 ± 0.07 km/sec, and that calculated by Båth is 3.07 ± 0.04 km/sec. Båth (1954, p. 303) points out

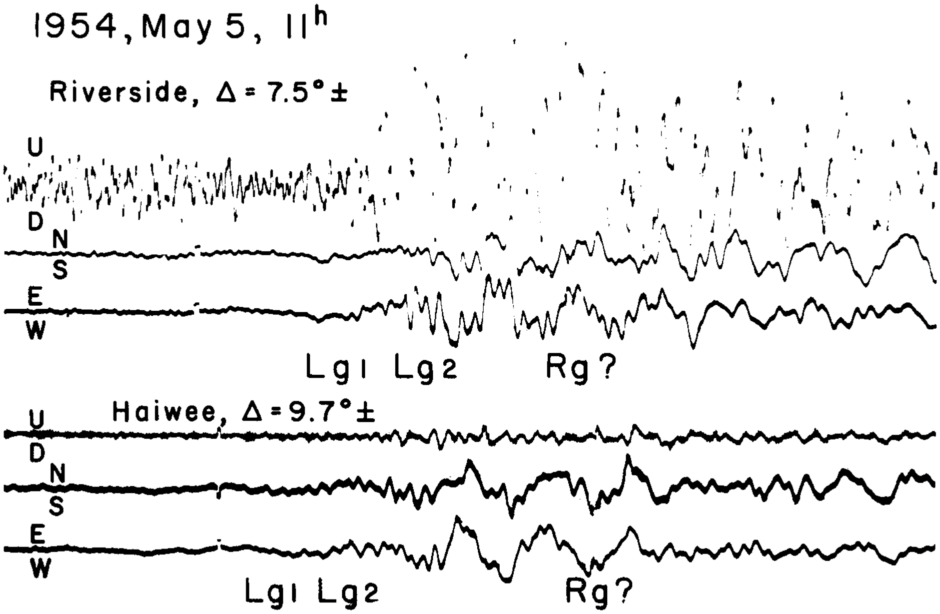

FIG. 3. Portions covering about $2\frac{1}{2}$ minutes of seismograms recorded at Riverside and Haiwee, California, from an earthquake which originated approximately at 27° North Lat., $112\frac{3}{4}$° West Long. (Lower California). Instruments: Benioff short-period vertical seismographs (U-D) and horizontal standard Wood-Anderson torsion seismographs (N-S; E-W). Δ is the approximate epicentral distance in degrees. Since the epicenter is to the SSE of the stations, waves of the *SH* type should be recorded in the same direction on both horizontal components, *SV* and *P* in opposite directions. *Lg* and *Rg* have relatively short periods; long-period waves are surface waves.

that the observed velocities of *Rg* scatter more than those for both *Lg* waves, and that *Rg*, too, may consist of two or more wave types.

The *Lg*-waves have been tentatively considered to be transverse waves with a prevailing *SH* component. However, Press and Ewing (1952, p. 221) state: "During the first few cycles the waves have approximately equal amplitudes on all three components, but the transverse horizontal rapidly gains amplitude . . . within about 30 seconds," and Miss Lehmann (1953, p. 249, 251) points out that the waves have a considerable vertical component. Båth (1954, p. 316) finds that at Uppsala and at Kiruna the vertical component of *Lg* is smaller than the horizontal. Since few stations recording *Lg* in the California shock of 1952 had two horizontal components with approximately the same constants, some Pasadena records of other shocks have also been investigated. Figure 3 shows two examples: the horizontal seismographs have about the same constants; however, the vertical records are written by seismographs with appreciably different characteristics.

From all records which have been studied, it follows that in *Lg*, motion perpendicular to the ray (*SH* type) prevails, but that not infrequently an appreciable vertical component is recorded; in portions of *Lg*, the motion is approximately in the direction of the ray (*SV*-type). Some irregularities may result from body or surface waves superposed upon *Lg*. Otherwise, motion consisting of a shear in the horizontal component accompanied by a relatively large vertical component is not predicted by the classic wave theory for either body or surface waves.

MECHANISM FOR TRANSMISSION OF THE *Lg*- AND *Rg*-GROUP

In a discussion of the mechanism for the transmission of *Lg*, Båth (1954, p. 321) states that not only the velocities but also their different relation to focal depth indicate that *Lg2* belongs to a lower layer than *Lg1*. He finds that in Sweden the *Lg2*-waves show a maximum energy (relative to *S*) for shocks originating at a depth of the order of 40 km, while the relatively largest *Lg1* waves are recorded for shocks with a focal depth of 20 km or less. He considers the possibility that *Lg1* is a wave travelling completely above the depth with maximum velocity (that is, roughly in the uppermost 10 km) with many reflections at the earth's surface, and that *Lg2* is a wave travelling in the low-velocity channel, that is at depth between roughly 10 and 20 km. However, he points out correctly (l.c., p. 322) that for foci below the level of maximum velocity, that is, for foci deeper than roughly 10 km, *Lg1* could then not exist if the classical wave theory holds, but that diffraction and scattering may be the mechanisms actually responsible.

The conclusion of Båth that *Lg2* seems to be propagated in a lower layer than *Lg1* seems to be doubtful, especially since the determination of focal depths in intervals of 10 km near a depth of 50 km (partly by Gutenberg and Richter) is not accurate enough for finding the relationship between focal depth and relative amplitudes of the phases from relatively few records of shocks supposedly slightly deeper than normal. On the other hand, use of the velocity ratio between *Lg* and *Rg* requires better knowledge of the wave types involved than we have now. Clues concerning the wave types may result from Båth's findings that the energy ratios *Lg2/S* and *Lg2/Rg* decrease with increasing magnitude, and that for shocks of a magnitude greater than 5 more energy passes into *Rg* waves than into *Lg2* waves, and more into *Lg2* than into *Lg1* waves.

Gutenberg (1954, 1955) has concluded that the hypothetical "gabbro" layer between the Conrad and the Mohorovičić discontinuities may be another low-velocity layer (Figure 4, left). The average velocity reported for this layer in a few regions on the basis of records of artificial explosions is roughly 3.8 km/sec (Gutenberg, 1955) and corresponds probably to the top of the layer. In Figure 4 some possible types of paths of *Lg* waves are sketched. The mechanism shown in the diagrams explains the long duration of the *Lg* group. The actual distribution of wave velocities with depth may differ appreciably from that indicated in Figure 4 which had been assumed on the basis of amplitude and travel time studies before the existence of *Lg2* was known. If these velocities and their changes

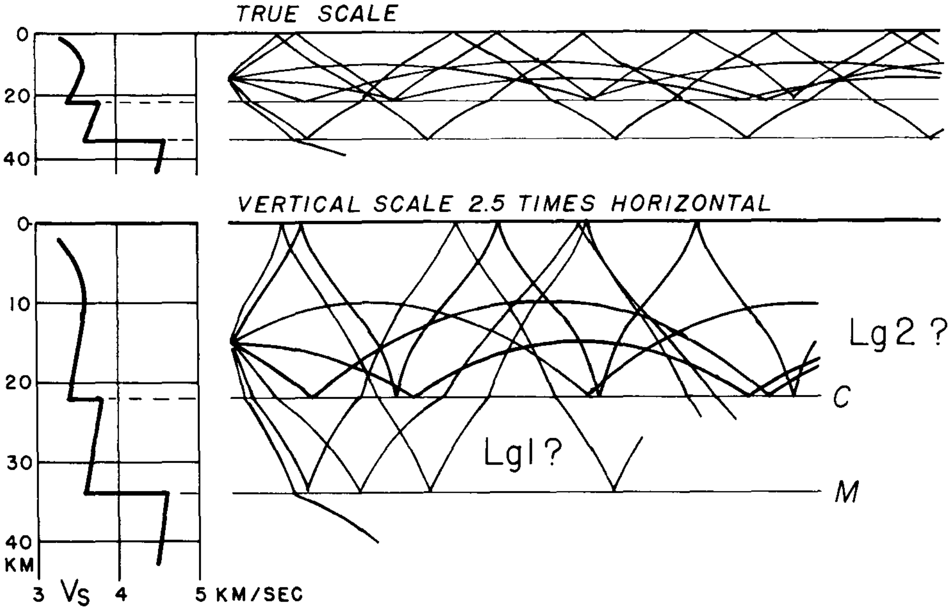

FIG. 4. Selected paths of transverse channel waves, constructed on the basis of the assumed velocity *Vs* of transverse waves as a function of depth, given at the left. *C* = Conrad discontinuity between "granitic" and "gabbro" layers. *M* = Mohorovičić discontinuity separating the latter from the ultrabasic mantle.

with depth are a good approximation to the actual values, it seems likely that *Lg2* is a true channel wave along the channel in the "granitic" layer, while *Lg1* in addition possibly enters the channel in the "gabbro" layer. The various types of paths can not be constructed with confidence before the velocities are better known as a function of depth.

The observed relative importance of *SH* (dominant), *SV*, and *P* (negligible) in these channel waves corresponds to the fact that total reflection is most likely for *SH*, less so for *SV*, and still less for longitudinal waves. The following is a summary of the conditions for total reflection at the earth's surface and at a discontinuity, where r = ratio of transverse velocity above to transverse velocity below; r is assumed to be between 0.6 ± and 1.0:

	SH	*SV*	longitudinal
surface	always	$i > 35° \pm$	never totally reflected
discontinuity	$i > \arcsin r$	$i > \arcsin r$	never totally reflected

There is little disagreement about *Rg*. Båth (1954, p. 307) points out that it shows dispersion of a type found theoretically by Haskell (1951) for Rayleigh waves propagated through surface layers including a low-velocity layer. At present, the conclusion that *Rg* is such a Rayleigh wave is preferred by all who have studied it.

EFFECTS OF THE STRUCTURE OF THE EARTH'S CRUST ON *Lg*-WAVES. MICROSEISMS

Below the ocean basins the "granitic" layer is usually missing, and the "gabbro" layer extends only to a depth of $10 \pm$ km below sea level. Consequently, there is probably no decrease in velocity with depth in the ocean bottom above the Mohorovičić discontinuity, especially since such a decrease would begin at a greater depth under the oceans than under the continents as a consequence of the lower temperature under the oceans. Therefore, no *Lg* or *Rg* waves can be expected with a path across an ocean bottom—if our explanations for their mechanism are correct. Actually, all authors agree that neither *Lg* nor *Rg* waves with paths through the crust under the deeper parts of ocean are observed. In addition, *Lg* and *Rg* waves are weakened or even disappear in crossing recent mountain chains, where probably the channels are too much distorted to permit transmission of guided waves (Båth, 1954, p. 300, 319; Ewing, oral communication). In California they seem to be more weakened in crossing the transverse ranges and the Sierra Nevada than along paths between Mexico and Southern California.

Gutenberg (discussion of the paper by Båth, Rome, 1954) has pointed out that in many respects the properties of the regular microseisms with periods of 4 to 10 sec show great similarity to those of earthquake waves guided by the lithosphere channel. This includes especially their velocity of about 3 km/sec, their periods, and the "barriers" to their propagation in bottoms of deep oceans and under young mountain chains. All recent results are in favor of the hypothesis (Gutenberg, 1954) that these microseisms are due to waves guided by the lithosphere channels.

THE PROBLEM OF $\overline{P}$ AS A CHANNEL WAVE

There is some doubt about the existence of longitudinal waves guided by the lithosphere channel, similar to the *Pa*-waves guided by the asthenosphere channel. The group of longitudinal waves including $\overline{P}$ and the following first motion in seismograms recorded at distances beyond about 150 km was the first which was suspected to be guided by the low-velocity channel in the lithosphere (Gutenberg, 1951, p. 162). In Southern California, the amplitudes in the $\overline{P}$-group have a maximum at a distance near 130 km, and then decrease about exponentially to a distance of at least 500 km, where for a shock of magnitude 7.6 the ground amplitudes of such waves with periods of the order of 1 sec should be roughly 100 microns on the basis of a study of many smaller shocks. Only very few seismograms of the California shock in 1952 at distances beyond about 8° [for shorter distances all amplitudes are too large] show waves possibly belonging to this group. Their amplitudes continue to decrease rapidly with the distance, and near

$20°$ the ratio of their amplitudes to the period does not exceed $3 \pm$. This agrees with the findings by Båth (1954, p. 316) that at greater distances there is no indication of longitudinal waves guided by the lithosphere channel.

The velocity of "$\overline{P}$" in Southern California and in other regions has been found to be $5.6 \pm$ km/sec. However, the maximum of the amplitudes moves gradually from one phase of the group to another. The group creates "the impression of waves travelling in a dispersive medium." (Gutenberg, 1951, p. 162), and the observed velocities are probably group velocities, which could differ from the phase velocity since the periods increase noticeably with distance (Gutenberg, 1936).

WAVES GUIDED BY THE ASTHENOSPHERE CHANNEL (*Pa* AND *Sa*)

The seismograms of the Kern County shock include only a few written at distances between about 45° and 70°. This explains the gap in the data (Figure 2) for these distances. In addition, *Pa* arrives close to *PP* at distances of between about 25° and 40°, and should be close to *SKS* or in the *SKKS*-group beyond about 100°, which limits the use of records in these ranges of distances for *Pa*. On records of the Kern County shocks, *Pa* is usually less definite than *Lg* or *Sa*. Thirty-three instances of *Pa* were found, of which 24 are at distances beyond 23°. At distances between 7° and 15°, the large longitudinal phase following the beginning of the record was measured. Again, two least square solutions were made, one including all data (minimum distance 7°), another for the phases arriving at distances greater than 23°. The following travel times result:

All data: $$t = (9.2 \pm 1.9) + \Delta/(8.056 \pm 0.017) \qquad (5)$$

$\Delta > 23°$ only: $$t = (5.7 \pm 4.3) + \Delta/(8.032 \pm 0.034). \qquad (6)$$

Residuals are shown in Figure 2 (bottom). Since constants in the two equations agree within the limits of error, the large longitudinal phase near the beginning of records at distances between 7° and 15° may be produced by the same mechanism as $\overline{Pa}$ at greater distances. Press and Ewing find that their "*Pn* at great distances" has about the same velocity as *Pn* at short distances ($8.1 \pm$ km/sec); Caloi finds velocities between 7.9 and 8.0 km/sec for his *Pa*.

In the records of the California shock of 1952, *Sa* is frequently rather clear (Figure 5). Its forms and periods are similar to those reproduced by Caloi (1953) for *Sa*-phases in shocks originating at intermediate depth. Since *Sa* is recorded near *SS* at distances of 30° to 40° ,and at greater distances is superimposed on the long surface waves, it is best found on medium-period instruments. At distances near $15 \pm °$, the arrival time of the frequently rather large transverse phase, following the expected *S* by roughly 20 seconds (depending on the distance) was measured on records of six stations. In addition, *Sa* was found on records of twenty-two stations at distances between 18° and 93°. The method of least squares gives the following travel times:

All data: $t = (18.2 \pm 3.2) + \Delta/(4.453 \pm 0.008)$ (7)

$\Delta > 18°$ only: $t = (13.0 \pm 4.6) + \Delta/(4.436 \pm 0.012)$. (8)

Residuals are shown in Figure 2 (center). The velocity of *Sn*, which probably travels near the top of the asthenosphere channel, seems to be about 4.7 km/sec (Gutenberg, 1954). Caloi as well as Press and Ewing find a velocity of about 4.4 km/sec for *Sa* (without assumption of a constant term). Caloi points out that

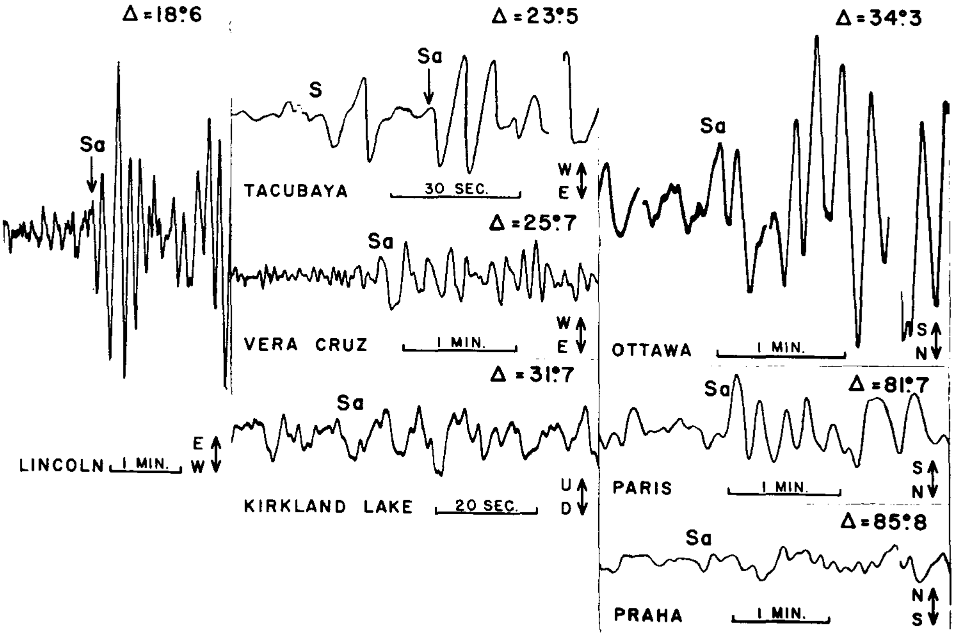

FIG. 5. Portions of selected records of the main Kern County, California, earthquake of 1952, showing *Sa*. Δ = epicentral distance in degrees; arrows indicate direction of ground motion.

Sa has a considerable vertical component, and this is confirmed by records of the California shock of 1952.

There is general agreement that *Pa* and *Sa* waves are found which have travelled through oceanic crustal layers. Both phases have been observed on European records of the 1952 California shock, and *Pa* was recorded (with some doubt) at Huancayo, La Paz, and Apia; *Sa* was recorded at Honolulu. There is no indication of *Pa* or *Sa* on records written in Japan or New Zealand, but this may be a result of not-well-suited instrumental constants in Japan and of large microseisms in New Zealand. Thus far, there is no clear evidence that the asthenosphere channel is missing in any of the larger units of the earth's crust.

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