The asthenosphere low-velocity layer

B. Gutenberg

Over 100 years ago, Airy and Pratt independently concluded that the weight of a rock column of a given cross section above a depth of the order of 100 km is about the same everywhere regardless of the elevation of the surface of the earth. Airy assumed that the greater height of mountains is compensated by greater thickness of the relatively light crustal material below them, while Pratt believed that the density of the crustal material under mountains is smaller than that under lowlands. Dutton (1889) introduced the expression isostasy for the " tendency to bulge ", where the lighter material exists, and, " where the denser matter existed, ... to depress the surface ". Barrell (1914) pointed out that the maintenance of isostasy requires that below the relatively strong, mainly crustal " lithosphere " there is a layer with smaller yield strength, which he called " astenosphere " (1914, p. 659). He assumed that below the asthenosphere the yield strength is greater again, and that processes requiring gradual movements take place mainly in the asthenosphere. His ideas have been extended by Daly (1940) and others.

At least as early as 1907, v. Wolff had concluded that at depths near 100 km the temperature should be close to the melting point of ultrabasic material. Gutenberg (1926) investigated if earthquake waves show a corresponding appreciable decrease in velocity. He studied the amplitudes of longitudinal waves of shallow earthquakes as a function of distance up to 3000 km and found that starting at a distance of about 200 km from the source the amplitudes of longitudinal waves decrease roughly exponentially with distance and reach a minimum near an epicentral distance of about 1700 km, where they increase suddenly by a factor of roughly 100. From these results he concluded that there is a slight decrease in the velocity of longitudinal waves at a depth of about 75 km, but no indication of molten material. This " asthenosphere low velocity layer ", as he called it later to distinguish it from

lithosphere low velocity layers, has been ever since the subject of investigations with more and more improved methods and results.

Byerly (1926) noticed that the travel time curve of longitudinal waves consists of several portions with different slope, and that at a distance of about 20° " there appears to be a rather abrupt increase in the apparent surface velocity ". Subsequently, this discontinuity in the slope of the travel time curve, which has been found for many regions, as well as a discontinuity in the earth which has been assumed to cause this sudden change in slope have been called the "twenty degree discontinuity ". The depth of this hypothetical discontinuity in the mantle has been investigated repeatedly by Jeffreys, by Miss Lehmann and by others (see, e. g., Jeffreys, 1959, pp. 98-102), who found for its depth values of between about 200 and 500 km. On the other hand, investigations of travel times of longitudinal waves by Gutenberg and Richter (1935, p. 339) confirmed Gutenberg's earlier conclusion that at the depths which are reached by the rays emerging at distances of less than 14° " the velocity may be closely constant, or may even decrease with depth ". If it decreases by more than the critical rate, given by $dV/dr = V/r$, " the range from 6° to 14°, with its small amplitudes, corresponds to a so-called shadow zone, while the large amplitudes just beyond 14° may represent a focal point " (Gutenberg and Richter, 1935, p. 339); compare Figure 1. They found no indication of a discontinuity.

Gutenberg and Richter (1939, p. 531) realized that " additional information bearing on this problem can be derived from a study of seismograms of deep shocks recorded at short distanced ". They selected earthquakes originating at depths between 50 and 250 km which had been recorded at Huancayo, Peru, at distances of between 2 and 23 degrees. They found that the shadow zone at the surface has its greatest extent, when the source is at a depth of about 80 km, and that it disappears, if the source is deeper than roughly 200 km. They concluded (1939, p. 537) that this confirms and extends "the conclusion that a slight decrease in the velocity of longitudinal waves occurs at a depth of about 80 km. Quantitative data have then been given by Gutenberg (1945, p. 126; see also Gutenberg, 1959b, p. 85) when he discussed the determination of magnitudes of deep focus earthquakes on the basis of amplitudes of P and S. He concluded (1945, p. 129) that " the shadow zone for P and S waves at epicentral distances near 10°, indicating a slight minimum in wave velocity at a depth near 100 km, has been confirmed ".

In an " attempt to find more accurately this decrease in wave velocity with increasing depth " Gutenberg (1948) investigated seismo-

grams of earthquakes which had originated at distances of between 4°
and 25° from Pasadena. In addition to a confirmation of the earlier
results about the amplitudes in and near the shadow zone, he concluded
that the travel time curves of P and S " consist of two branches, the
first almost a straight line ending in a shadow zone, the second beginn-
ing with relatively large amplitudes and with a delay relative to the
first branch. . . . . . All observations are explained on the assumption that
the velocity of P as well as S waves decreases somewhat (between ½ and
3 per cent?) at a depth. . . . . near 80 to 100 km ". He also pointed out
" that such a low-velocity layer has now been established for the solid
earth, the ocean, and the atmosphere, providing shadow zones for
elastic (sound) waves in all three. . . . . Only a relatively small decrease
in velocity is required to produce a rather extensive and pronounced
shadow zone ". This has been discussed in more detail by Gutenberg
(1954 b).

A new method to find the velocities in the asthenosphere low-
velocity layer was introduced by Gutenberg (1953). He pointed out
that the wave velocity V at the depth $h = R - r$ (R = radius of the
earth, r = distance of the source of the earthquake from the earth's
center) of an earthquake can be found from the apparent velocity V^*
at the point of inflection of the travel time curve from

$$V = V^* r / R \qquad [1]$$

Gutenberg applied this equation to values of V^* found from travel time
curves of 82 shocks originating at depths of between about 50 and 600 km,
mostly in or near Japan. He constructed curves for the velocity as
function of depth for P and S, which show a clear decrease with depth
below the Mohorovičić discontinuity " with a minimum at a depth of
roughly 100 km for longitudinal waves and 150 km for transverse waves.
Poisson's ratio increases from about 0.26 at a depth of 50 km to 0.29 at
250 km. . . . . Small local differences in the rate of the velocity decrease
result in noticeable local differences in amplitudes and perhaps even
in travel times at epicentral distances between about 5° and 20° ".

If a low-velocity layer exists below the crust, the corresponding
travel time curve may be interrupted (Figure 1). In such instances, the
arrival times of P and S which are reported by the stations at distances
inside the shadow zone may belong to diffracted or to channel waves.
Especially, in instances where the source is in the low-velocity layer
(Figure 1d), waves leaving the source nearly horizontally do not reach
the surface, and the point of inflection of the travel time curve cannot

be observed. Usually, such instances are not detected on the basis
of the reported times, since it is not possible to decide whether a small
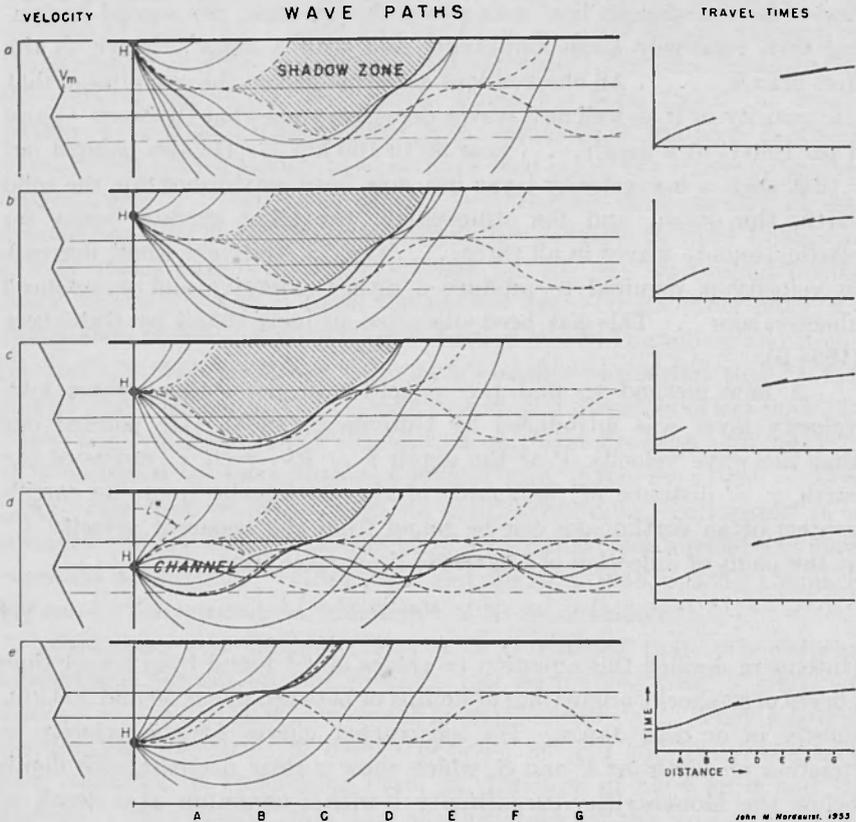beginning of P on a record belongs to a small direct wave or to a

VELOCITY   WAVE   PATHS   TRAVEL TIMES

Fig. 1. – Sketches of waves paths, travel times and shadow zones for various
       focal depths if a low-velocity layer exists. The velocity-depth curve
       at left has been assumed and does not correspond to conditions in
       the crust and upper mantle (After Gutenberg, 1954b).

diffracted wave or to movement at the surface produced by a channel
wave propagated within a few wave lengths below the surface (Figure
1d). In such instances, a continuous travel time curve is frequently
drawn on the basis of all reported times although these may refer
to various types of waves. If the apparent velocity at the point
of inflection of such a spurious travel time curve is used, equation [1]
leads to a result which still is close to the actual velocity at the depth

of the source. If, however, such a fictitious travel time curve is used
to calculate the wave velocity as a function of depth, assumptions may
have to be made, including discontinuities (as in the case of the " twenty
degree discontinuity ") and a spurious velocity-depth curve may be
obtained.

A very important discovery was made by Caloi (1953, 1954) when
he observed and explained correctly two new phases in records of earth-
quakes originating at depths of between 20 and 200 km. He found
that these phases, which he called Pa and Sa are propagated to great
distances with constant velocities of about 8.0 and 4.4 km/sec respec-
tively. Caloi explained these phases as waves guided by the low-velocity
layer in the asthenosphere. Båth and others consider the possibility
that channel waves, especially Lg, are surface waves of higher modes
(I. U. G. G. Chronicle No. 22, 1959, pp. 223, 224). Pa and Sa have
been reported independently by Press and Ewing (1955) who suggested
tentatively " a mechanism of transmission involving ' whispering gallery '
propagation in the mantle by multiple grazing reflections from the
Mohorovičic discontinuity ". The velocities observed by them are 7.98 to
8.24 km/sec for Pa-waves, while they observed two Sa-wave types with
velocities of 4.58 and 4.4 km/sec respectively. They, too, observed
these waves in shocks originating at depths of as much as 200 km.
Waves originating at foci over about 60 km deep are no longer grazing
the Mohorovičic discontinuity and should loose an appreciable amount
of energy at each reflection at the discontinuity, so that the "whispering
gallery " explanation can not apply at least in these instances. Guten-
berg (1955b) found Pa and Sa waves in records of the Kern County,
California, earthquake of 1952 with velocities of 8.05 and 4.44 km/sec
respectively at distances to almost 100°. " There is general agreement
that Pa and Sa waves are found which have travelled through oceanic
[subcrustal] layers . . . . .   There is no clear evidence that the astheno-
sphere channel is missing in any of the larger units of the earth's sub-
crustal layers " (Gutenberg, 1955b, p. 293).

The Kern County, California, earthquake of 1952 has furnished
excellent data for the travel times of waves at the epicentral distances
involved in the study of the effects of the asthenosphere channel. Guten-
berg (1954a) combined these travel times with those from several other
well recorded shocks in Southern California (Figure 2). In the interim,
it had been well established from travel times of waves generated by
earthquakes and by artificial explosions that the velocities of waves
refracted immediately below the Mohorovičic discontinuity are between

8.0 and 8.2 km/sec for P and between 4.6 and 4.8 km/sec for S (see, e. g.,
Gutenberg, 1955a, p. 23; 1959b, pp. 32-35). Finding of lower velocities
than 8.0 and 4.6 km/sec respectively in the upper portion of the mantle
by Jeffreys and others, e. g. increase in the velocity of P from 7.75
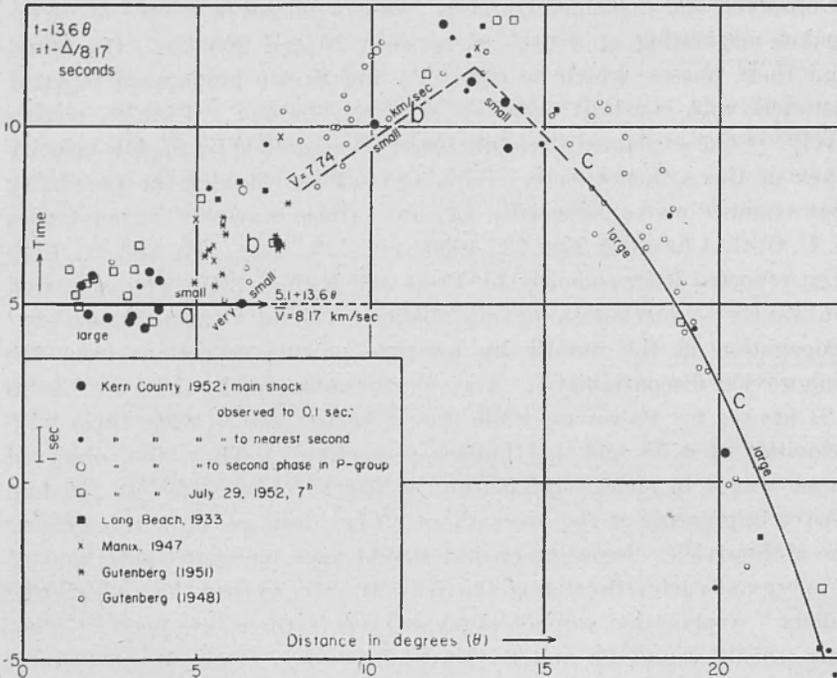km/sec below the crust to 7.95 at 100 km and in that of S from 4.35

Fig. 2. – Observed travel times of P minus 13.6 times distance in degrees
(subtracted to permit use of larger time scale) for earthquakes in South-
ern California (based on Gutenberg, 1954a; see also Gutenberg, 1959b,
pp. 78 and 79).

km/sec below the crust to 4.45 at a depth of 100 km (reproduced, e. g.
by Jeffreys, 1959, p. 122 and also by Bullen, 1953, p. 211) contradict
the many reliable values obtained from refracted (or diffracted) waves
directly below the crust unless a low-velocity layer is assumed.

Longitudinal waves having a velocity of between 8.1 and 8.2
km/sec are found in California (Figure 2) at epicentral distances between
about 1° and 7° with rapidly decreasing amplitudes. Between about 7°
and 13° the first waves are small and scatter about a straight line cor-

responding to a velocity of only about 7.7 to 7.8 km/sec. Starting at about 14° longitudinal waves with amplitudes increasing rapidly near about 15° and indicating a rapid increase in velocity below a depth of about 200 km are found. Again, these results have been interpreted by Gutenberg (1954a, p. 346) as an effect of the asthenosphere channel. He concluded that the channel " is probably due to a greater effect of the increase in temperature with depth than that of the increase in pressure at the depth where the melting point of the material is appoached, while above and below this channel the effect of the increase in pressure with depth prevails ". Gutenberg (1954b) has also sketched the rays originating from earthquakes at various depths in and near the channel, the resulting forms of the travel time curves and of the shadow zones (reproduced in Figure 1). Theoretical computations related to this problem have been carried out by Båth (1957).

Attempts to evaluate theoretically the temperature gradient necessary to produce a low-velocity layer in the asthenosphere have been made by Valle (1956). He concluded that a gradient in excess of 14°/km is necessary for a low-velocity layer for P waves, and in excess of 11°/km for S waves. The actual gradient just below the crust has been estimated by some geophysicists to be greater, by others to be smaller than the critical values calculated by Valle. It may be more accurate now to calculate the temperature gradient from the decrease in wave velocities than vice versa. The relationship of these two quantities has also been studied by Birch (1958) but he has applied his equations only to crustal layers. While on his assumptions the theory does not indicate a decrease in velocity with depth for granite in the crust, a slight decrease is indicated by his curve for gabbro below a depth of about 20 km (Birch, 1958, p. 167). However, all such numerical results depend on the rather uncertain values of the temperature as a function of depth (Figure 3).

Experimental data on the effect of temperature and pressure on rocks have been published by Hughes and Maurette (1957) and by Birch (1958). On the assumption of plausible values for pressure and temperature in the earth's crust Hughes and Maurette find that in granites and in several basic igneous rocks the wave velocities can be expected to increase with depth in roughly the upper ten km of the crust and to decrease with depth below about 10 km. On the same assumptions they find that in dunite which is considered sometimes to be representative of the material in the asthenosphere, the velocities of both wave types should reach a maximum already under the conditions at a depth of about 5 km.

The investigation of records obtained from atomic explosions again confirmed the existence of the shadow zone at distances between roughly 1000 and 2000 km from the source (Anonymous, 1958). On the basis
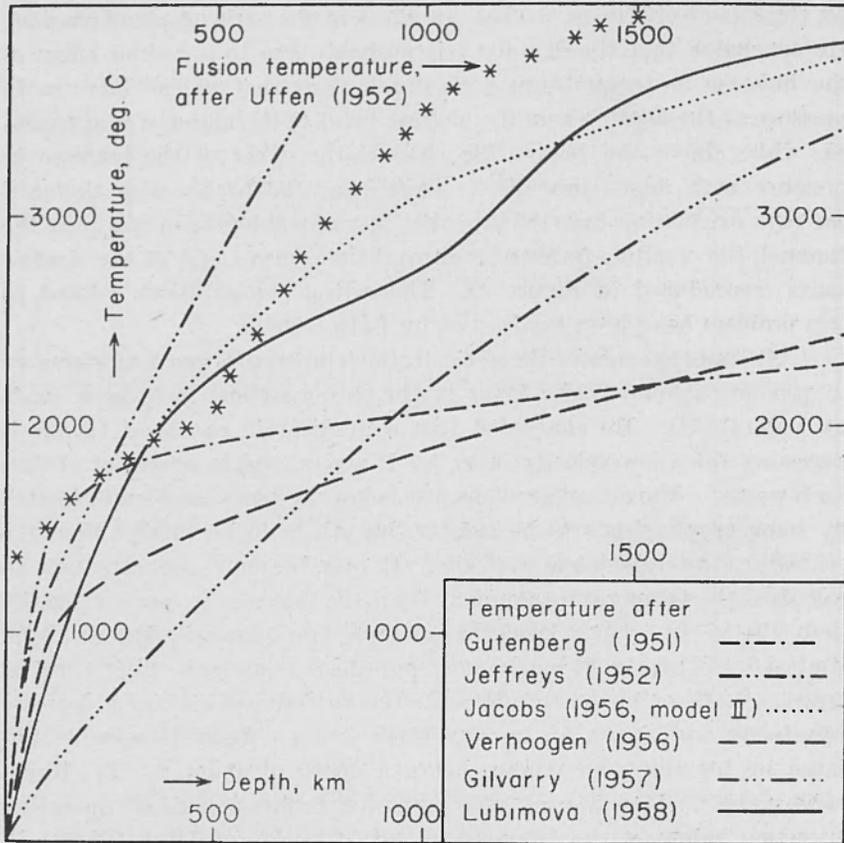
Fig. 3. – Temperature in the upper portion of the earth's mantle according to various authors; fusion temperature after Uffen (1952) based on the velocity of seismic waves.

of the increasing evidence for the asthenosphere channel, Gutenberg (1958) has suggested to identify Bullen's region B with the asthenosphere and to put its lower boundary tentatively and somewhat arbitrarily at a depth of 200 km. Problems connected with the " twenty degree discontinuity " have been discussed in more detail by Gutenberg (1959b, pp. 75-89).

As has been pointed out it had been found over 50 years ago that the temperature in the upper mantle must be close to the melting point. There are still no reliable data available, how close the temperature comes to the melting point; Figure 3 gives some estimates of the temperature as a function of depth as well as a curve for the melting point which has been calculated by Uffen (1952) by use of seismic data on the basis of the Einstein-Debye theory of solids and Lindemann's theory of fusion. Figure 3 shows that at depths of between 100 and 300 km the temperature in the earth may indeed be close to the melting point of the material. Additional important evidence has been given by Gorshkov (1958) who reported that transverse waves from Japanese earthquakes are not recorded in Kamtchatka, if the waves have to travel under the belt of volcanoes, while they are well recorded at about the same distances at stations to which the wave paths are to the west or east of this belt. Gorshkov concluded that there seem to be magmatic foci at a depth of roughly 60 km. Their "occurrence . . . . . in the upper part of the mantle is probably far from being casual and is connected with a velocity decrease of seismic waves " (Gorshkov, 1958, p. 28). " The most probable reason for this phenomenon is a transition from a crystalline state into an amorphous condition " (Gorshkov, 1958, p. 109). These results give additional weight to the over fifty year old concept of the asthenosphere as a relatively weak layer, in which gradual movements, subcrustal currents, movements to make isostasy possible, and movements of extended portions of continental blocks relative to the main portion of the mantle may take place more easily than in the lithosphere.

More recently, observations of surface waves have added new evidence for the asthenosphere low-velocity channel. Intependent investigations of Takeuchi, Press and Kobayashi (1959) and of Dorman, Ewing and Oliver (1959) indicate that the observed dispersion curve for Rayleigh waves agrees better with calculated curves, if a low-velocity layer of the type discussed here is assumed than on the assumption that the velocities increase with depth and have a discontinuity. Moreover, according to Landisman, Sato and Ewing (1959) the dispersion curve for Love waves indicates a low-velocity layer for shear waves at depths of roughly 100 to 200 km under continents.

Press (1959) has found that "G-wave velocities for continental and oceanic paths do not differ by more than about 2 per cent. Since their velocity is controlled by the asthenosphere low-velocity zone, this must be present under contients and oceans ". Landisman, Sato and Ewing

(1959) find that " under oceans the region of low shear velocities rises
to depths of about 50 km; the low-velocity zone is thicker, and the
velocities are lower than under continents ".   Unfortunately, velocities
of longitudinal waves found from refraction shooting under oceans do not
help much in investigations of the low-velocity layer under oceans, since
the velocities depend on the locally relatively much differing tempera-
tures in the ocean bottoms at a given depth; moreover, there is some
doubt how far the material under the Mohorovičic discontinuity under
ocean bottoms differs from that under continental areas and also be-
tween various oceanic areas (compare Gutenberg, 1959a).   Moreover,
immediately under the relatively shallow Mohorovicic discontinuity
in oceanic areas the wave velocities are possibly lower than at some-
what greater depths.

Vesanen, Nurmia and Porkka (1959) have used time differences
pP–P to find the depth of the low-velocity channel.   They obtain values
of 45 km for Alaska, 80 km for North Japan, 95 km for the Tonga region
and 120 km for South America.

The values for the velocities at depths between the Mohorovicic
discontinuity and about 70 km have been more uncertain than those for
greater depths.   Gutenberg (1959a) has studied the wave velocities
which have been calculated from refracted (or diffracted) waves at the
shortest distances where they can be observed on records of earthquakes
or of artificial explosions.   He has found that these reported velocities
decrease with increasing depth of the Mohorovicic discontinuity.   Ap-
plication of the method of least squares to velocities reported during the
past ten years for continental areas give

$$V = (8.08 \pm 0.0131) - (0.011 \pm 0.0016) (h - 40) \text{ km/sec} \qquad [2]$$

$$v = (4.60 \pm 0.044) - (0.004 \pm 0.0056) (h - 40) \text{ km/sec} . \qquad [3]$$

The rate of decrease in the longitudinal velocity V is statistically signi-
ficant, that in the transverse velocity v is significant only if it is consider-
ed in connection with that for the longitudinal velocity.

To obtain more data on the velocities in the upper portion of the
mantle, it seemed advisable to apply equation [1] to travel time curves
of P and S for additional earthquakes.   The accuracy of this method
has been tested by applying equation [1] to the travel time curves of
Jeffreys and Bullen (1940), who have calculated travel times of P and S
for various focal depths h.   Jeffreys and Bullen express the depth h by
the ratio of the radius $r = (R - h)$, where R is the earth's radius, to the

radius of the average top of the earth's mantle. Since Jeffreys has assumed that in the average the crust is 33 km thick, the depth 0.00 of Jeffreys-Bullen corresponds to an actual depth of 33 km or to a radius of 6371 — 33 = 6338 km; generally, their depths h correspond to 6338h + 33 km below the earth's surface; for example, h = 0.01 corresponds to a depth of 96 km. Jeffreys and Bullen have included in these tables for each focal depth the time differences corresponding to consecutive distances; up to 10° epicentral distance, these differences give the time difference in sec/5°, beyond 10° in sec/10°. In Table 1, for each focal

Table 1

$h - 33 \over R - 33$	h	$R \over r$	$1 \over V^*$ sec $\over$ 10°	V^* km $\over$ sec	V km $\over$ sec	$1 \over v^*$ sec $\over$ 10°	v^* km $\over$ sec	v km $\over$ sec	V v assumed Jeffreys (1952)	
0.00	33	0.9948	144	7.72	7.68	254	4.37	4.35	7.75	4.35
0.01	96	0.9848	139	7.99	7.87	247	4.50	4.43	7.94	4.44
0.02	160	0.9749	134	8.29	8.08	240	4.63	4.51	8.13	4.54
0.03	223	0.9648	130	8.55	8.25	231	4.81	4.64	8.33	4.64
0.04	287	0.9549	124	8.96	8.56	224	4.96	4.74	8.54	4.74
0.05	350	0.9451	120	9.26	8.75	217	5.12	4.84	8.75	4.85
0.06	413	0.9352	116	9.58	8.96	209	5.32	4.98	8.97	4.96
0.07	477	0.9251	107	10.38	9.60	197	5.64	5.22	9.50	5.23
0.08	540	0.9152	102	10.89	9.97	187	5.94	5.44	9.91	5.46
0.09	603	0.9053	99	11.22	10.16	178	6.24	5.65	10.26	5.67
0.10	667	0.8953	95	11.70	10.48	171	6.50	5.82	10.55	5.85

Velocity V of longitudinal waves and v of transverse waves in km/sec at the depths h found by applying equation [1]. V^* and v^* are the reciprocals of the respective maximum values $(dt/d\Delta)^*$ given by Jeffreys and Bullen (1940) for focal depths h in km (second column). In the first column, the corresponding values for the depth are given which are used by Jeffreys-Bullen and in the International Seismological summary. R = 6371 km, radius of earth, $r = R - h$, Δ = epicentral distance.
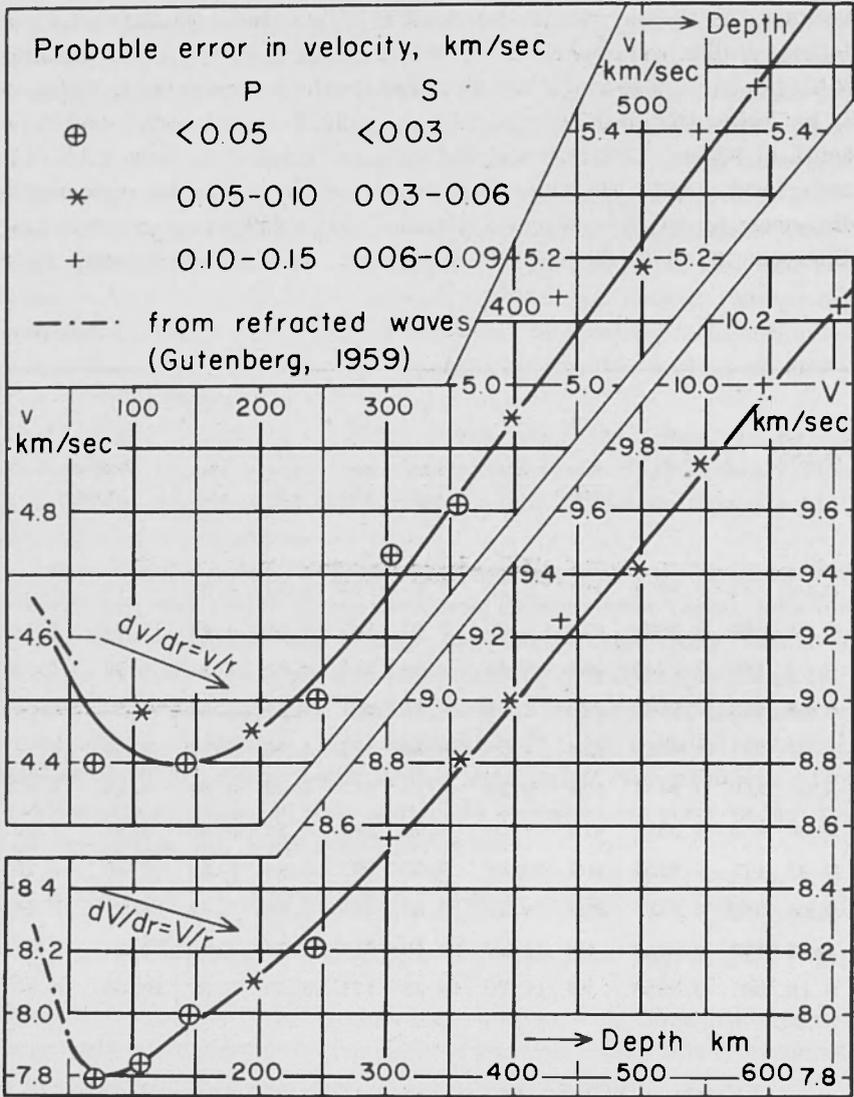
Fig. 4. – Velocities V of longitudinal and v of transverse waves under continents. The beginning of the curves correspond to observations of waves refracted just below the Mohorovicic discontinuity which have been recorded after shallow earthquakes or artificial explosions (Gutenberg, 1959a); data for the individual points are from Table 3.

depth their minimum difference, corresponding to the apparent velocity
at the point of inflection of the travel time curve for each of their focal
depths, is listed (always in sec/10°). This is followed by the correspond-
ing apparent velocity and then multiplied by r/R, which is also given in
the table. According to equation [1], the result is the actual velocity
at the depth h. In the last two columns of Table 1 the velocities origi-
nally assumed for the calculations by Jeffreys and Bullen are listed
from Jeffreys (1952, p. 115, or 1959, p. 122). The good agreement be-
tween these values and the corresponding values found from equation [1]
shows that the use of equation [1] does not introduce appreciable errors,
and that the travel time curves for various depths calculated by Jeffreys
and Bullen (1940) correspond closely to the assumed velocities as fun-
ction of depth.

It has been pointed out already that for the upper portion of the
mantle these velocities are noticeably smaller than the well determined
velocities immediately below the Mohorovičic discontinuity (see, e. g.
Gutenberg, 1959a; 1959b, pp. 32-35). It follows that closely below the
discontinuity there must be either a gradual or a sudden decrease in
velocity to the values of Jeffreys-Bullen, if these are correct. On the
other hand, the data obtained from refracted waves for the neighborhood
of the discontinuity and those for greater depths, based on equation [1]
fit one smooth curve (Figure 4).

The velocities found by Gutenberg (1953) had been based on travel
times of earthquake waves for the period 1928 to 1940. No later issues
of the International Seismological Summav (ISS) were available in 1953.
Moreover, during the war, data of many stations were less reliable than
before 1941 or missing. More plentiful and accurate data start again
with 1947; at the time of writing the ISS is available as far as the first
quarter of 1951. Consequently, the ISS for the period Januay, 1947
to March, 1951 has been searched for earthquakes for which (a) the
focal depth could be reliably determined, and (b) a number of arrival
times had been reported near the point of inflections of the travel
time curves of P or S which were sufficient to determine the values
of V^* or v^*. In Table 2 these shocks are listed with the resulting calcu-
lated velocities for the respective depths h. The procedure was the
same as in the earlier paper.

Next, average velocities for certain depth ranges have been calcu-
lated. The results for a combination of the old data (1953) with those
of the present Table 2 are given in Table 3. In section (a) of Table 3,
weights have been used as follows: for 1928 to 1940 a weight 2 for results

Table 2

Date			Region	Lat.	Long.	h	Velocity and Q	
							P	S
1947	Jan.	30	Hindu Kush	36 N	71 E	200	7.6 C	D
	Feb.	4	N. Japan	42 N	143 E	60	7.7 B	4.3 C
	Feb.	18	Marianne Is.	33 N	137 E	440	9.1 B	5.4 B
	Sept.	1	Off Italy	40 N	15 E	250±	8.2 B	4.5 C
	Oct.	3	Marianne Is.	31 N	139 E	380	8.6 C	4.8 B
	Oct.	29	N. Japan	44 N	145 E	110	7.9 B	4.6 C
	Nov.	14	N. Japan	44 N	143 E	180	8.0 B	4.5 B
	Dec.	7	Hindu Kush	37 N	70 E	230	8.0 B	4.5 B
1948	March	13	Rumania	46 N	27 E	190	8.1 B	D
	March	15	Marianne Is.	32 N	138 E	370	8.9 B	D
	April	11	Marianne Is.	33 N	136 E	430	9.2 B	D
	April	29	Rumania	46 N	27 E	160	8.2 C	D
	May	29	Rumania	46 N	27 E	140	8.2 C	D
	July	18	Hindu Kush	37 N	71 E	220	8.4 B	D
	Sept.	5	Marianne Is.	30 N	139 E	400	8.8 B	5.05 B
	Sept.	7	Hindu Kush	36 N	71 E	210	8.0 B	4.35 C
	Sept.	11	East Mediterr.	37 N	23 E	80	7.9 B	4.4 C
	Nov.	10	Marianne Is.	30 N	141 E	110	7.8 B	D
	Nov.	15	Marianne Is.	29 N	139 E	440	9.0 B	4.95 C
1949	Jan.	13	Kermadec	26 S	178 E	670	10.1 C	5.8 C
	March	11	Hindu Kush	37 N	71 E	240	8.25B	4.5 C
	March	19	S. Japan	31 N	130 E	160	7.9 B	D
	May	10	Hindu Kush	36 N	71 E	140	7.9 B	D
	June	12	S. America	27 S	63 W	600	10.0 C	5.5 C
	Aug.	17	N. Japan	43 N	146 E	80	7.9 B	4.35 C
	Sept.	23	Siberia	44 N	134 E	430	8.8 B	5.05 B
	Oct.	25	Japan	36 N	141 E	100	7.6 C	D
	Nov.	3	Kurile Is.	48 N	154 E	160	7.7 C	D
	Nov.	17	Marianne Is.	32 N	137 E	430	9.1 C	D
	Dec.	18	New Zealand	35 S	180 E	200	8.2 B	D
	Dec.	21	S. America	19 S	64 W	610	10.2 B	D
	Dec.	26	Rumania	46 N	27 E	140	8.1 C	D
1950	Jan.	12	N. Japan	42 N	142 E	110	7.9 B	4.55 B
	Jan.	13	Japan	38 N	142 E	60	7.5 B	4.3±C
	Jan.	16	Rumania	46 N	27 E	130	8.0 C	4.5 C

Table 2 (continued)

Date	Region	Lat.	Long.	h	Velocity and Q P	S
Feb. 3	N. Japan	42 N	144 E	60	7.6 C	4.3 C
Feb. 21	Off Sakhalin	47 N	146 E	390	9.1 B	4.7 B
Feb. 23	Off Sakhalin	49 N	148 E	540	9.6 B	5.4 B
Feb 28	Off S. Sakhalin	46 N	147 E	340	8.6 A	4.7 B
March 28	Japan	36 N	141 E	60	7.9 B	D
April 20	Off N. Japan	44 N	151 E	60	7.7 B	D
May 17	Sea of Japan	39 N	130 E	580	9.7 C	5.3 C
May 17	Hindu Kush	37 N	71 E	230	7.9 C	4.45 C
May 20	Hindu Kush	37 N	71 E	130	8.05B	4.25 C
May 28	Marianne Is.	32 N	139 E	290	8.5 B	4.75 C
June 26	Hindu Kush	38 N	74 E	120	7.7 B	D
July 12	Off Sakhalin	46 N	144 E	300±	8.3 C	4.75 C
Aug. 3	Mexico	18 N	100 W	90	7.7 C	D
Aug. 3	Japan	40 N	142 E	50	7.6 B	D
Sept. 16	S. Japan	33 N	132 E	120	7.9 B	4.4 B
Oct. 10	Hindu Kush	37 N	71 E	220	8.2 B	4.5 B
Nov. 9	Kurile Is.	47 N	151 E	180	8.1 B	4.5 C
Dec. 23	Japan	37 N	141 E	55	7.7 B	. D
Dec. 24	Hindu Kush	37 N	71 E	230±	8.3 C	4.6 C
1951 Jan. 4	Central Asia	38 N	74 E	140	8.1 C	4.4 C
Jan. 6	Hindu Kush	36 N	71 E	230	8.2 A	D
Jan. 28	Hindu Kush	37 N	71 E	120	8.1 C	D
Feb. 7	Marianne Is.	31 N	140 E	100	7.8 B	4.5 C
March 5	Riu Kiu Is.	28 N	128 E	180	7.8 B	4.4 B
March 10	Japan	41 N	142 E	80	7.5 B	D
March 23	Kermadec	31 S	180 E	250	8.4 C	D
1954 March 29	Spain	37 N	4 W	640	10.4 A	5.5 C

Fundamental data for shocks used in the determination of wave velocities V (longitudinal) and v (transverse) in km/sec as function of the depth h of focus (in km). Qualities Q decrease from A (good) to D (uncertain).

of quality A, 1 for B; for the new data, 3 for A, 2 for B and 1 for C.   In
section (b) of Table 3, all results have been given equal weight.   The
first line of section (a) contains data for depths between 70 and 90 km,
both inclusive, while for the corresponding line in section (b) also data
for $h = 60$ km are included.   Otherwise the data used in (a) and (b) are
the same.   The differences between the two solutions are well within the
limits of errors.   The probable errors of the results (b) do not permit a
conclusion, how large the systematic errors are, especially for focal
depths in the channel.   On the other hand, they include the effect of the
changes in the velocities with depth within each range of depths, usually

Table 3

(a)				(b)						
Aver. h	V	Aver h	v	Aver. h	n	V	Aver. h	n	v	
82	7.80	78	4.44	72	16	7.79 ± 0.04	70	10	4.40 ± 0.03	
108	7.85	108	4.47	109	14	7.85 ± 0.04	108	8	4.48 ± 0.03	
148	7.98	145	4.39	147	21	7.98 ± 0.04	143	9	4.39 ± 0.03	
199	8.11	194	4.46	199	12	8.10 ± 0.07	194	5	4.45 ± 0.05	
246	8.21	248	4.49	246	17	8.21 ± 0.04	245	9	4.50 ± 0.02	
303	8.60	305	4.72	303	4	8.56 ± 0.11	303	4	4.73 ± 0.03	
357	8.80	357	4.80	357	11	8.81 ± 0.07	356	8	4.81 ± 0.03	
395	8.90	394	4.92	396	11	8.97 ± 0.09	397	9	4.85 ± 0.06	
436	9.27	437	5.18	436	10	9.26 ± 0.10	437	8	5.14 ± 0.08	
500	9.48	501	5.19	499	7	9.42 ± 0.10	502	6	5.18 ± 0.05	
544	9.74	552	5.38	545	4	9.75 ± 0.09	545	4	5.40 ± 0.07	
597	10.05	590	5.50	595	4	10.00 ± 0.15	590	3	5.47 ± 0.09	
648	10.33	655	5.65	655	2	10.25 ± 0.16	655	2	5.65 ± 0.15	

Average velocities V (longitudinal) and v (transverse) in km/sec
for various depth ranges (average h in km) based on combined data
from Gutenberg (1953, Table 1) and from Table 2 of present paper;
(a) using weights (see text), (b) with weight 1 for all data; n = number
of shocks.

$\pm$ 20 km. The probable error for V obtained from a single shock is about $\pm$ 0.2 km/sec, that for v about $\pm$ 0.1 km/sec.

The results of the solutions (b) have been plotted in Figure 4 with indication of the probable errors for each average depth. As mentioned, smooth curves for V and for v as function h can be drawn which include the results obtained from travel time curves of waves refracted just below the Mohorovičic discontinuity (Gutenberg, 1959a). The minima of both curves in Figure 4 are slightly smaller than the velocities found for Pa and Sa respectively, as it has to be expected if these are channel waves. The slope of the beginning of both velocity-depth curves is greater than that given by $dV/dr = V/r$ and $dv/dr = v/r$ respectively, so that in the average continent the low-velocity chanel for P as well as for S begins at the Mohorovičic discontinuity.

Unfortunately, the distribution of epicenters of deep-focus shocks for which enough arrival times at nearby stations have been reported to find the apparent velocity at the point of inflection of the travel time curves is very uneven. The great majority of the epicenters which have been used are in the region of Japan, the Mariannas and the Kurile Islands. An attempt has been made to calculate velocities as function of depth for a combination of data obtained for earthquakes in the Hindu Kush region and in Rumania. The resulting velocities for P at depths of between 100 and 250 km are in the average about 0.05 km/sec smaller than those for the total (that is, mainly the Japanese area), those for S are about 0.1 km/sec smaller for depths near 120 km, about 0.02 km/sec greater for depths between 150 and 250 km. All these differences are within the limits of errors, considering the relatively small numbers of useful shocks (25 for P, 12 for S) in the Hindu Kush and Rumanian areas. However, there is no doubt that there are small local differences, especially under ocean bottoms. Unfortunately, no data exist or can be expected there to apply our method. However, the velocities found from waves refracted immediately below the Mohorovičic discontinuity leave no doubt (Gutenberg, 1959a) that the beginning of the velocity-depth curves for the mantle under oceanic areas differs somewhat from that under continents, although the channel exists under both.

In addition to small local differences in the velocity-depth curves, the varying depth of the Mohorovičic discontinuity can be expected to result in differences between the shadow zones (Figure 1) in different regions. Where, under otherwise equal conditions, the discontinuity is relatively deep, the channel can be expected to be relatively narrow; the velocity which exists immediately below the discontinuity there can

be expected to be reached again at a shallower depth than in regions
where the discontinuity is at a smaller depth (compare Figure 4).
Consequently, the low-velocity zone should be thicker under oceans
than under continental lowlands, as has been found by Landisman,
Sato and Ewing (1959) and should have a minimum thickness under
roots of mountains.

Table 4

d	V	v	s
40	8.08	4.60	0.260
60	7.87	4.51	0.256
80	7.80	4.45	0.259
100	7.83	4.42	0.266
120	7.89	4.40	0.274
140	7.94	4.39	0.280
160	8.00	4.40	0.283
180	8.06	4.42	0.285
200	8.12	4.45	0.285
250	8.30	4.54	0.286
300	8.51	4.66	0.286
350	8.75	4.81	0.283
400	9.00	4.95	0.283
450	9.25	5.09	0.283
500	9.49	5.22	0.283
550	9.74	5.36	0.283
600	9.99	5.50	0.282

Average velocities V of longitudinal and v of transverse waves
in km/sec and of Poisson's ratio s in the upper portion of the earth's
mantle under continents as a function of the depth d in km (September,
1959).

Poisson's ratio is close to 0.26 immediately below the Mohorovičic discontinuity, increases to about 0.28 in the lower portion of the channel (Table 4) and reaches a maximum of nearly 0.29 at depths of about 200 to 250 km; in the whole remainder of the mantle it is between about 0.28 and 0.30. It is interesting that below the channel it does not decrease again to the value of 0.26 which it has near the top of the mantle.

ABSTRACT

Results based on observations of various phenomena practically prove the existence of a world-wide low-velocity asthenosphere channel. Combination of new and earlier data on seismic wave velocities show that under continents the channel begins at the Mohorovicic discontinuity with velocities which are the smaller the greater the depth of the discontinuity. They are about 8.1 and 4.6 km/sec for longitudinal and transverse waves respectively, if the discontinuity is at a depth of 40 km. Longitudinal waves have a minimum velocity of about 7.8 km/sec near a depth of 80 km, transverse waves a minimum of about 4.4 km/sec near a depth of 150 km. The velocity which exists at a depth of 40 km under continents is reached again at a depth of about 190 km by longitudinal waves and at a depth near 270 km by transverse waves. The rate of decrease in velocity immediately below the discontinuity is in the average greater than the critical rate for formation of shadow zones. There is no indication of a discontinuity in the upper portion of the mantle. Poisson's ratio is nearly constant, 0.26, down to a depth of about 80 km, and increases in the asthenosphere channel to almost 0.29. Below the channel, it does not return to the value of 0.26 but remains nearly constant in most of the deeper portions of the mantle.

RIASSUNTO

Risultati basati su osservazioni di vari fenomeni provano praticamente l'esistenza di un canale astenosferico a flessione di velocità, diffuso in tutta la terra. L'unione di dati passati e recenti nella velocità dell'onda sismica mostra che sotto i continenti il canale comincia alla discontinuità di Moho-

rovicic, con velocità che sono tanto più piccole quanto più grande è la profondità della discontinuità.

Esse sono circa di 8.1 e 4.6 km/sec. rispettivamente per le onde longitudinali e trasversali, se la discontinuità è a una profondità di 40 km. Le onde longitudinali hanno una velocità minima di 7.8 km/sec circa intorno ad una profondità di 80 km, le onde trasversali una velocità minima di 4.4 km/sec circa intorno a una profondità di 150 km. La velocità esistente ad una profondità di 40 km sotto i continenti è nuovamente raggiunta ad una profondità di circa 190 km dalle onde longitudinali, ed ad una profondità di circa 270 km dalle onde trasversali.

L'ordine di decrescita nella velocità immediatamente al disotto della discontinuità è in media più grande che l'ordine critico per la formazione nelle zone d'ombra. Non vi è alcun segno di una discontinuità nella parte superiore del mantello.

Il coefficiente di Poisson è quasi costante, 0.26, sotto una profondità di 80 km circa, e cresce nel canale astenosferico a quasi 0.29. Sotto il canale, esso non torna al valore di 0.26 ma rimane quasi costante nella maggioranza delle parti più profonde del mantello.

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