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Tunneling of Low-Frequency Waves through the Subcrustal Lithosphere*

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Abstract. Tunnel waves are long period waves which on overcritical incidence have penetrated through thin high-velocity layers. The generation and propagation of these waves are studied by numerical experiments using synthetic seismograms. Tunnel waves may occur as secondary and primary arrivals, on retrograde and prograde branches. Observations from explosion and teleseismic data with low-frequency tunnel waves are presented. – A model of the lower lithosphere with thin high-velocity layers explains not only the occurrence of tunnel waves but also the high-frequency transmission of P_n and S_n waves to teleseismic distances. This model is in accordance with the observation of seismic anisotropy in the upper mantle and suggests refinements of existing petrological models of the lower lithosphere.

Key words: Tunnel waves – Teleseismic P_n, S_n transmission – Thin high-velocity layers in lower lithosphere – Synthetic seismograms – Model of lower lithosphere.

1. Introduction

Within the framework of the plate-tectonic hypothesis the lithosphere is usually regarded as a rigid plate on top of the low-viscosity asthenosphere. While the fine structure of the earth's crust has been revealed by explosion seismic investigations, the subcrustal or lower lithosphere, i.e. the region between the crust-mantle boundary and the asthenosphere, used to be visualized as a rather homogeneous layer in which the velocities of the seismic waves increase only slightly with depth.

Recently there has been increasing evidence that the subcrustal lithosphere possesses a rather complex fine structure with alternating layers of high and low

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** To whom offprint requests should be sent velocities (Hirn et al., 1973; Ansorge and Mueller, 1973; Mayer-Rosa and Mueller, 1973; Kind, 1974). In several parts of the globe unexpectedly high velocities up to 9.0 km/s have been observed both in the oceanic and continental lower lithosphere (Hales et al., 1970; Ryaboi, 1966; Kosminskaya et al., 1969, 1972). Kosminskaya et al. (1972) drew attention to the problem which arises in seismology if these high-velocity layers occur in the lower lithosphere. In this case seismic ray theory will not allow rays to penetrate below the lithosphere unless their phase velocity is equal or larger than the high velocities in the subcrustal lithosphere. Therefore ray theory cannot explain the occurrence of the same or of smaller velocities at dephts in or beneath the asthenosphere observed by body waves from earthquakes in the crust. Kosminskaya et al. (1972) proposed already a solution to the dilemma: high velocities should occur only in thin layers which are not seen by low-frequency body waves from earthquakes observed on seismic stations. They become visible if observed with high-frequency body waves from explosions.

It is the purpose of this communication to present the results of numerical experiments (Schulz, 1975) with the help of synthetic seismograms computed by the reflectivity method (Fuchs and Müller, 1971) on the low frequency-leakage of waves through thin high-velocity laminas. The main result is a strong low-pass filtering of body-wave signals on non-geometrical ray-paths. This is regarded as dispersion of body waves introduced by the presence of such laminas since multiple path signals with different frequency content arrive at the same location. Then explosion and teleseismic observations are reported which show such a dispersion. This provides new and independent evidence for the presence of thin high-velocity laminas in the lower lithosphere.

2. Numerical Experiments with Tunnel Waves

When a seismic body wave (e.g. P-wave) is incident from a low velocity medium upon the plane boundary of a high velocity medium beyond the critical angle, the wave does not only cause a refracted S wave and overcritically reflected P and S waves but also an inhomogeneous P wave in the lower medium (see. Fig. 1). The amplitude A of this inhomogeneous wave decays exponentially with distance z from the interface:

$$A \sim \exp\left(-\frac{\omega}{c}\sqrt{1-\frac{c^2}{\alpha_2^2}}\, z\right)$$

where ω is the angular frequency, c the phase velocity of the wave and α_2 the P-velocity of the lower medium. The amplitude decay increases with increasing frequency and decreasing phase velocity, i.e. with increasing of angle of incidence. If the inhomogeneous wave reaches the boundary to another low-velocity layer with sufficient amplitude, it will produce body waves propagating in the low-velocity medium (Fig. 1 lower part). The P-ray with phase velocity c tunneled into the lower medium with P-velocity α_1 is refracted with the angle of incidence $\gamma_1 = \sin^{-1}(\alpha_1/c)$.

Since the high-frequency parts of the wave decay more rapidly than the low frequencies, the latter will be predominant in the wave penetrating into the lower

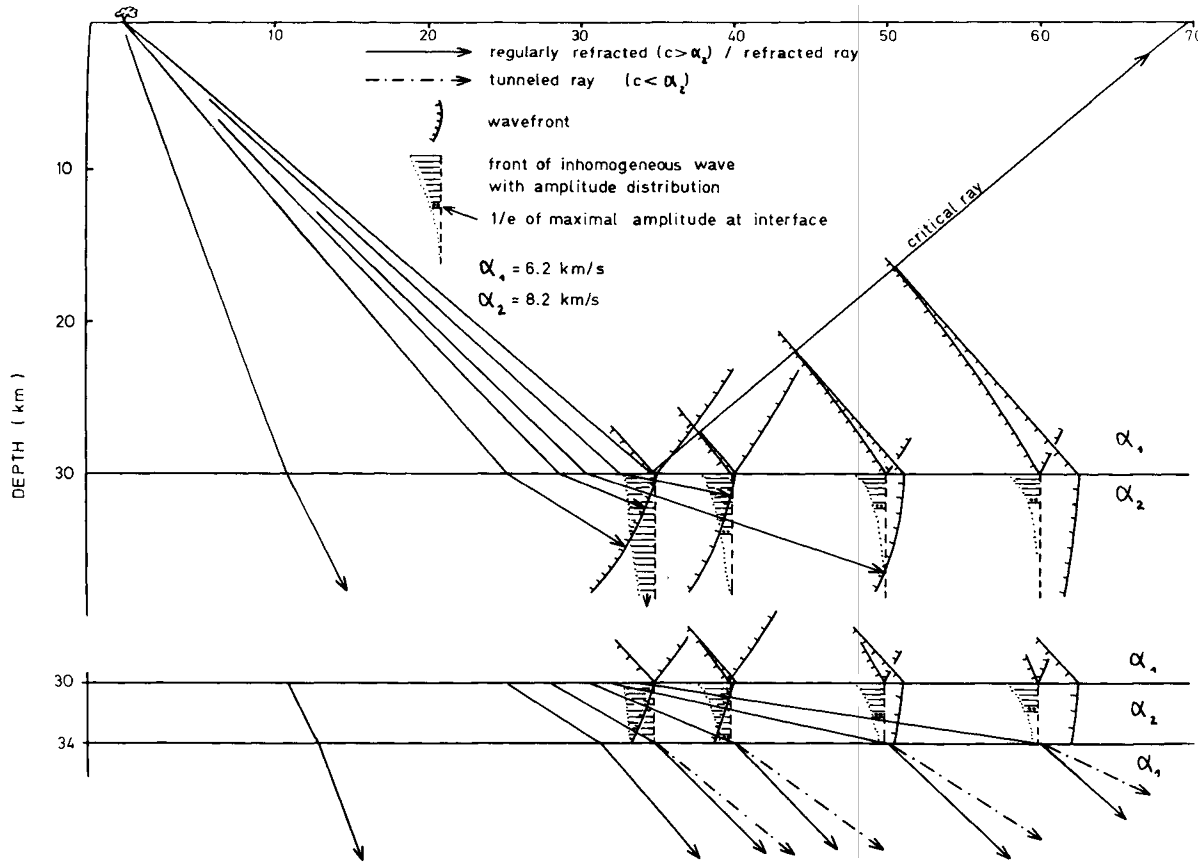

Fig. 1. Generation of a tunnel wave on the overcritical incidence on a thin high-velocity layer. Upper part: The overcritical reflection causes an inhomogeneous wave with amplitudes decaying exponentially with distance from the interface. Lower part: If the lower boundary of the high-velocity layer is reached by the inhomogeneous wave before its amplitudes are too strongly attenuated, a tunnel wave is generated in the lower halfspace. Solid lines: normally refracted rays. Dashed-dotted lines: tunneled rays

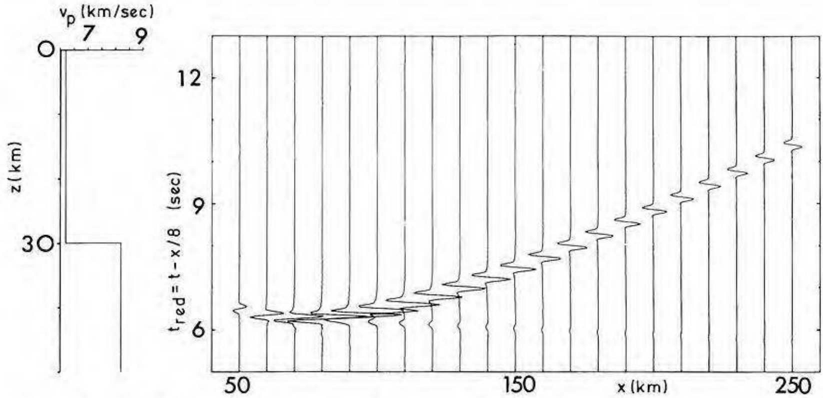

Fig. 2. Synthetic seismogram section of reflection from a first-order discontinuity. Explosion source with signal

$$f(t)=\begin{cases}\sin\frac{20}{3}\pi t-\frac{1}{2}\sin\frac{40}{3}\pi t & \text{for } 0\leqq t<0.3\\ 0 & \text{for } t<0,\ t>0.3\end{cases}$$

S-wave velocity $v_s=v_p/\sqrt{3}$; density $\rho=0.252+0.379\,v_p$

medium. In analogy to wave mechanics where a wave particle can penetrate or tunnel through a potential wall we will speak of a tunnel wave. This term has been used before by Richards (1973) for long-period body waves propagating in the lower mantle and leaking into the core, and by Mellman and Helmberger (1974) in the same sense as in this paper. The tunneling of low-frequency waves through thin layers has been treated by a number of authors. Brekovskikh (1960) discusses the energy penetrating into a high-velocity medium if a wave is incident from the low velocity halfspace at angles larger than critical (p. 275 and 297). Brekovskikh uses the term "lateral wave". Červený and Kozák (1972) and Červený et al. (1972) studied the occurrence of these waves, termed by them pseudospherical waves in a low-velocity channel with the aid of model seismic experiments using Schlieren-technique. Long-period waves tunneling through thin high-velocity layers have been used with success in seismic refraction experiments of exploration geophysics (Krey, 1957). One of the best examples of tunneled waves may be found in Figure 3 of Krey's paper.

We will now present a series of synthetic seismograms and begin with that for the P-reflection from a first-order discontinuity at a depth of 30 km (Fig. 2). The seismograms have been displayed with a reduction velocity of 8.0 km/s. If not stated otherwise the same reduction velocity will be used throughout this paper. The shape of the incident signal with a dominant frequency of 4 Hz is the same as that of the subcritical reflection. The reflection attains the largest amplitudes slightly beyond the critical distance where the head wave propagating in the lower medium separates from the overcritical reflection.

In the following Figures 3–5 a thin layer with thickness $h=1.25$, 0.3 and 0.1 km, respectively, is placed at a depth of 30 km followed by a low-velocity layer of 10 km thickness with the same mechanical properties as the upper halfspace.

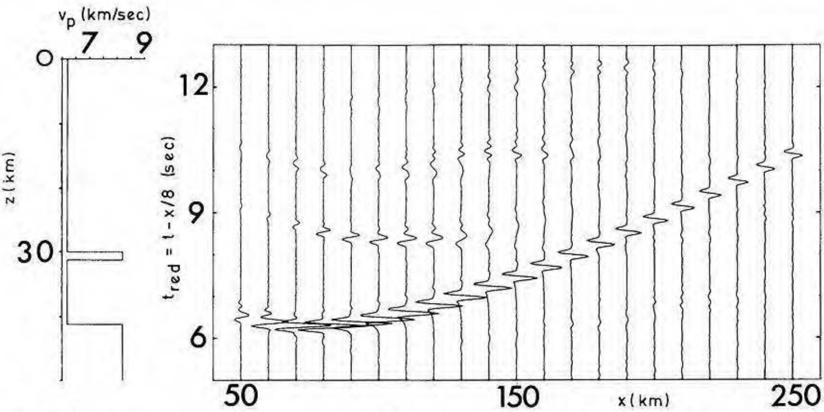

Fig. 3. Synthetic seismogram section of reflection from a thin layer ($h = 1.25$ km) over a low-velocity layer and a high-velocity halfspace. Other parameters as in Figure 2. The first signals following the strong reflection from the top of the thin layer is the undercritical reflection from the base of the low-velocity zone. Later signals are multiple reflections within the channel

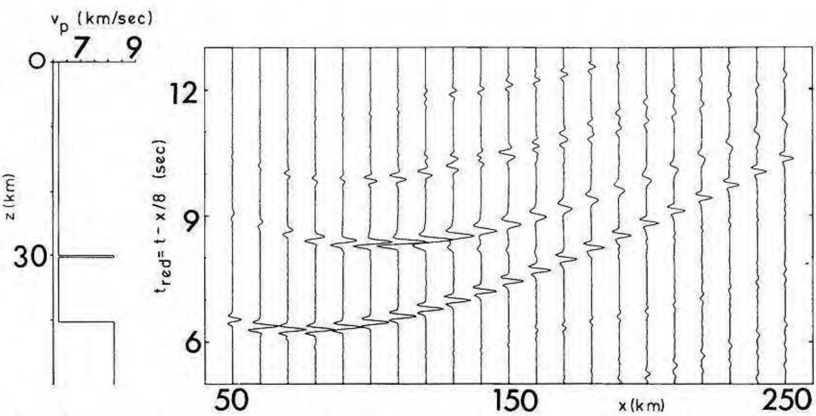

Fig. 4. Thickness of the thin layer $h = 300$ m, otherwise as Figure 3. The reflection from the base of the low-velocity layer beyond 130 km is overcritical and is formed by the wave which is tunneled through the thin layer. The amplitude of the reflection from the top of the thin layer has decreased especially in the neighbourhood of the critical distance

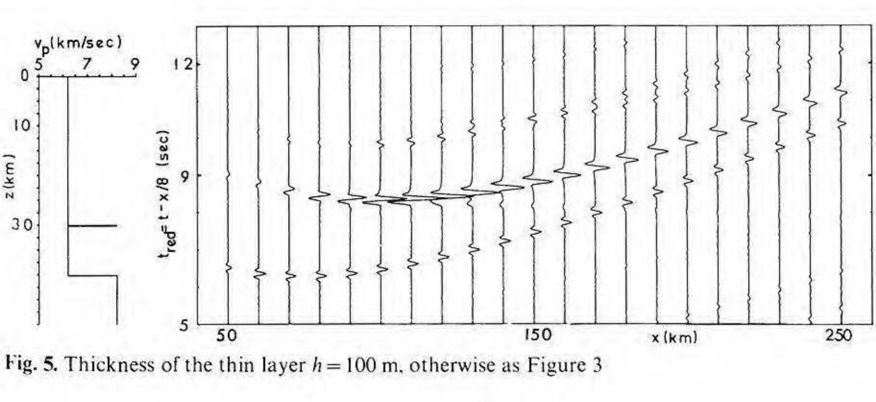

Fig. 5. Thickness of the thin layer $h = 100$ m, otherwise as Figure 3

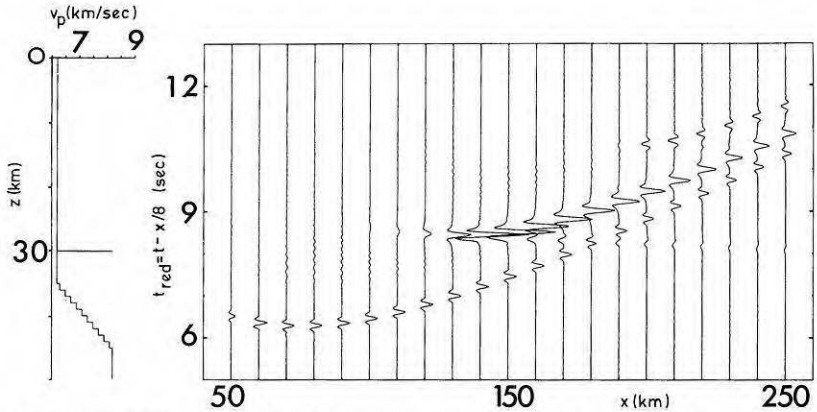

Fig. 6. Base of low-velocity layer is formed by gradient zone which suppresses undercritical reflections. Therefore the signal following the reflection from the top of the thin layer is a pure tunnel wave refracted from the gradient zone

The lower halfspace has the same properties as the thin high-velocity layer. The dominant wave length in the lamina is about 2 km.

In Figure 3 the reflection from the top of the thin lamina is followed by the subcritical reflection from the bottom of the low-velocity layer. Overcritical reflections from this interface are hardly seen since practically no energy with phase velocities less than 8.2 km/s is passed through the high-velocity layer. Only very low frequencies continue the subcritical reflection into the overcritical range. The amplitude distribution of the reflection from the top of the 1.25 km lamina is practically the same as from the first-order discontinuity (see Fig. 2).

In Figure 4 the thickness of the lamina is 0.3 km. In comparison with the previous figure the amplitudes of the overcritical reflections have increased strongly. At the same time the overcritical reflection from the thin layer has decreased, especially near the critical distance. This energy has penetrated the thin layer and appears now as overcritical reflection from the bottom of the low-velocity layer. This tunneling becomes even stronger if the thickness of the lamina is reduced to 0.1 km in Figure 5. The subcritical reflection from the bottom of the low-velocity layer is suppressed if the first-order discontinuity is replaced by a transition zone in Figure 6. The secondary arrival is now a pure tunnel wave reflected overcritically from the transition zone.

The frequency selection introduced by thin laminas becomes especially clear if a broad band signal is incident. In Figure 7 a wavelet with dominant frequencies of 1.6 and 6.5 Hz is used. The lamina has a thickness of 0.3 km as in Figure 4. Only in the subcritical range does the high-frequency component penetrate the high-velocity layer. In the overcritical range the reflection from the bottom of the low-velocity layer becomes abruptly a low-frequency signal. The high frequencies are screened off by the thin high-velocity layer. Such a layer can be regarded as a low pass-filter for tunnel waves and as a high-pass filter for overcritically reflected energy.

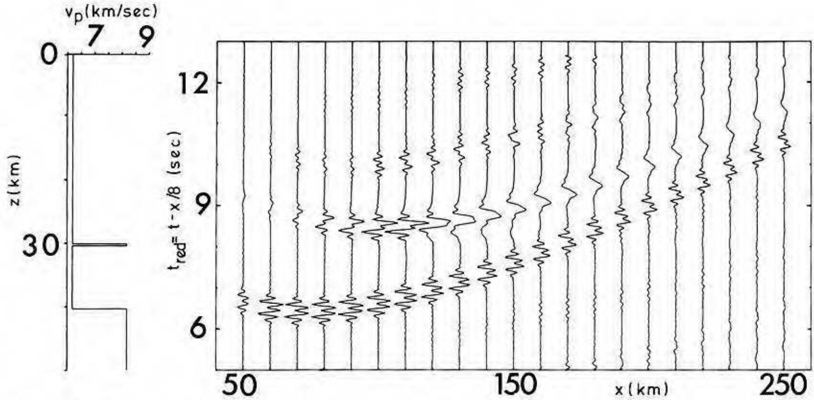

Fig. 7. Reflection of a broad band signal

$$f^*(t)=\begin{cases}\sin\frac{20\pi}{6}t-\frac{1}{2}\sin\frac{40\pi}{6}t+\sin\frac{80\pi}{6}t & \text{for } 0\leqq t<0.6\\ 0 & \text{for } t<0,\ t>0.6\end{cases}$$

from the same model as in Figure 4 ($h = 300$ m). High frequencies are only reflected undercritically but are screened off by the high-velocity layer for overcritical incidence. The tunnel wave can clearly be recognized from its low-frequency spectrum

The observation of high-frequency arrivals separated from low-frequency arrivals on the same seismogram in the P-field closely beyond the critical distance must be taken as strong indication for the presence of thin high velocity laminas. The tunnel waves occur with sufficient amplitudes only if the thickness of the lamina is less than 1/4 of the dominant wavelength. The velocity in the thin layer is higher or equal to the highest phase velocity of the tunnel wave.

3. Tunnel Waves in Explosion Seismic Observations

In the following figures we present evidence for the presence of tunnel waves in explosion seismic observations. Figure 8 is the record section of the crustal profile Florac 03 along the long-range profile in France (Sapin and Prodehl, 1973). In the distance range around 80 km the $P_M P$ reflection (phase 1) suddenly becomes clearly a low-frequency arrival while earlier phases on the same seismograms remain high-frequency signals. Therefore the low-frequency $P_M P$-wave cannot be caused by some anomalous behaviour of the near-surface material at the recording stations. The same low-frequency behaviour of $P_M P$ can also be observed at larger distances around 130 and 180 km. In this case also low-frequency $P_M P$ arrivals are clearly preceded by high-frequency signals. This frequency behaviour of the $P_M P$ branch is simulated by the presence of a thin high-velocity lamina overlying the crust mantle boundary in a synthetic record section in Figure 9. The phases 1, 2, and 3 in Figure 8 correspond to the phases C, B, and A, respectively, in Figure 9.

Figure 10a is a collection of similar examples of low-frequency arrivals preceded by high-frequency phases from explosion seismic observations on continents in

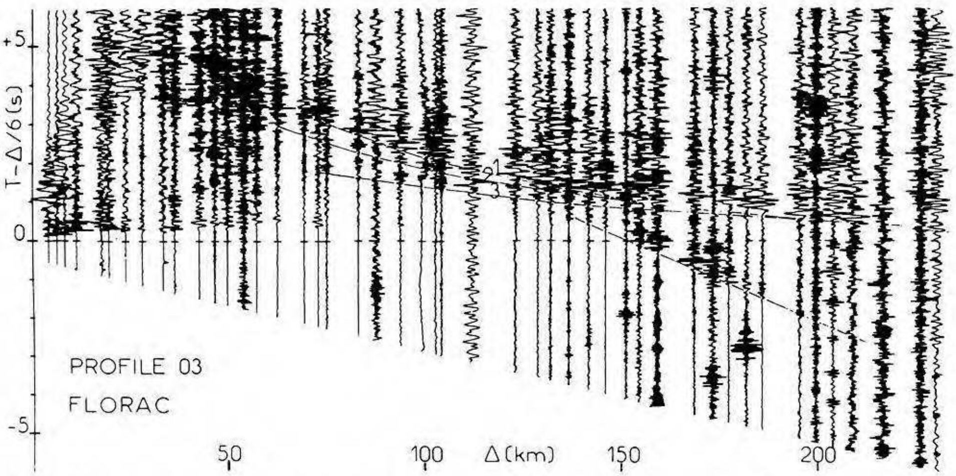

Fig. 8. Observed record section of the crustal profile FLORAC O3 in France (Sapin and Prodehl, 1973). Low frequency $P_M P$ phases can be recognized on seismograms around 80 km and 130 km with earlier high frequency arrival at the same station

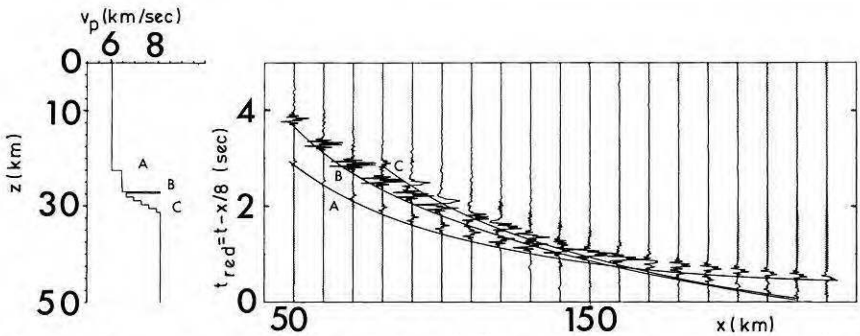

Fig. 9. Synthetic seismogram section for a model of the crust-mantle transition which simulates the observed frequency behaviour

several parts of the world: France, Canada, Portugal, Hungary, USSR. Our most extensive evidence for the presence of low-frequency arrivals in the presence of high-frequency precursors is taken from the LISPB seismic experiment in the British Isles (Bamford et al., 1975). Examples of the phenomena can be found on selected seismograms out to two or three hundred kilometres (Fig. 10b).

4. Tunnel Waves in Teleseismic Observations

In two studies (Walker and Sutton, 1971; Sutton and Walker, 1972) the authors present and interpret their observations of teleseismic *P*- and *S*-transmission through the upper mantle of the Pacific. They record events from the circum-

Pacific seismic belt by hydrophones at Midway, Wake and Eniwetok, and by SP-seismometers on Midway, Wake and Marcus. In Figure 11 their observations for events in the part north of 12° N of the Pacific are compiled.

The most striking feature of the hydrophone recordings (crosses in Fig. 11) is the absence of the normal refracted P-phase through the upper mantle which should be observed near the Jeffreys-Bullen travel time curve. Instead the hydrophones which are rather insensitive to 1–2 Hz frequencies receive seismic signals with dominant frequencies between 3–8 Hz lining up with an apparent velocity of 8.28 ± 0.03 km/s out to a distance of about 33°. From thereon the seismic arrivals recorded on the hydrophones follow closely J.-B. travel times. For distances smaller than 33° the hydrophones do not record high-frequency arrivals of normally refracted P-mantle phases with the exception of two deep focus earthquakes with hypocenters below 300 km. This indicates that highfrequency energy is not propagated through the lithosphere and below the asthenosphere unless the ray path is fairly steep in this part of the upper mantle as for deep earthquakes and for arrivals beyond 33°. – On the other hand the 1–2 Hz recordings of the SP-seismometers follow closely the J.-B. travel times.

Sutton and Walker interpret this frequency behaviour of the normally refracted P-wave as being caused by a low Q of the asthenosphere which affects the 3–8 Hz waves. Since the rays emerging beyond 33° plunge more steeply into the asthenosphere than those rays arriving at shorter distances, their total path through the low-Q asthenosphere is shorter. Therefore, high-frequency energy should be less attenuated at large distances.

Synthetic seismograms for a schematical model of the upper mantle with a prograde P-branch through the transition zone are shown in Figure 12. A zone of low $Q=100$ is built into the depth range of 30–40 km. The prograde P-branch refracted through the transition zone shows predominance of low-frequency energy at short distances while high-frequencies start to contribute only at large distances. This is the explanation of the frequency behaviour of teleseismic P-waves as proposed by Walker and Sutton scaled down to smaller depth and shorter distances.

However, the same frequency behaviour of P-transmission through the upper mantle can also be caused by a model of the lower lithosphere with thin high-velocity layers as will be shown in the following figures. The model of the upper mantle in Figure 13 is the same as in Figure 12 except for two differences: in Figure 13 Q is infinite, and a thin layer with thickness 625 m and a P-velocity of 8.2 km/s is introduced at a depth of 30 km. In this figure the wave tunneled through the thin layer forms a low frequency first arrival which is clearly to be recognized in the distance range 120–160 km. For larger distances the high frequencies of normally refracted rays start to appear. If the thickness of the layer or its wave velocities increase the high frequencies are stronger attenuated during tunneling through the thin layer. In Figure 14 its wave velocity is increased extremely to 9 km/s and is now larger tha the velocity in the lower halfspace ($v_p = 8.5$ km/s). The prograde P-branch is now a tunneled phase. Only beyond 250 km the high frequencies gradually start to contribute.

The frequency-distance behaviour of the refracted P-branch in Figures 12 and 14 is very much alike. A thin high-velocity layer acts as a low-pass filter in

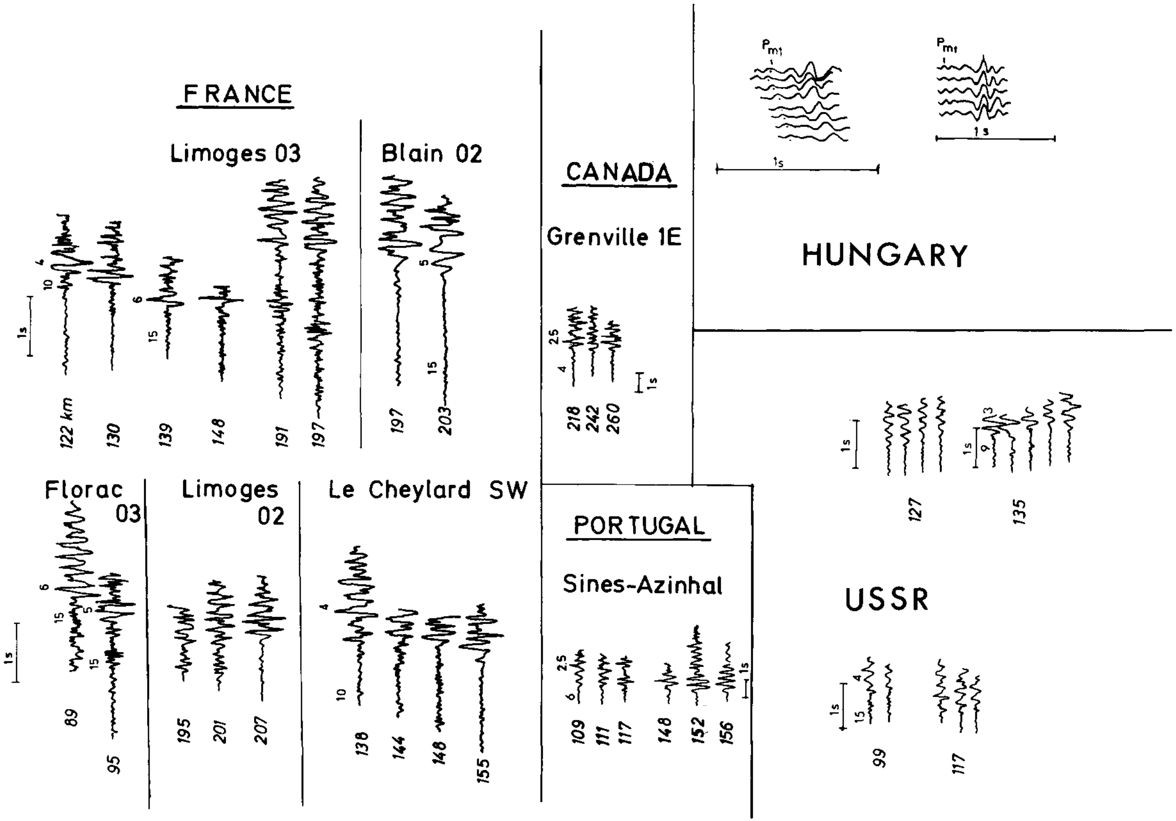

Fig. 10a. Examples of low-frequency $P_M P$ phases following high frequency precursors from various continental parts of the world. Dominant frequencies are indicated at the phases and distances at the beginning of every phase. Seismogram samples are form France (Sapin and Prodehl, 1973); Canada (Berry and Fuchs, 1973); Portugal (Mueller et al., 1973); Hungary (Mituch and Posgay, 1972); U.S.S.R. (Zverev and Tulina, 1971)

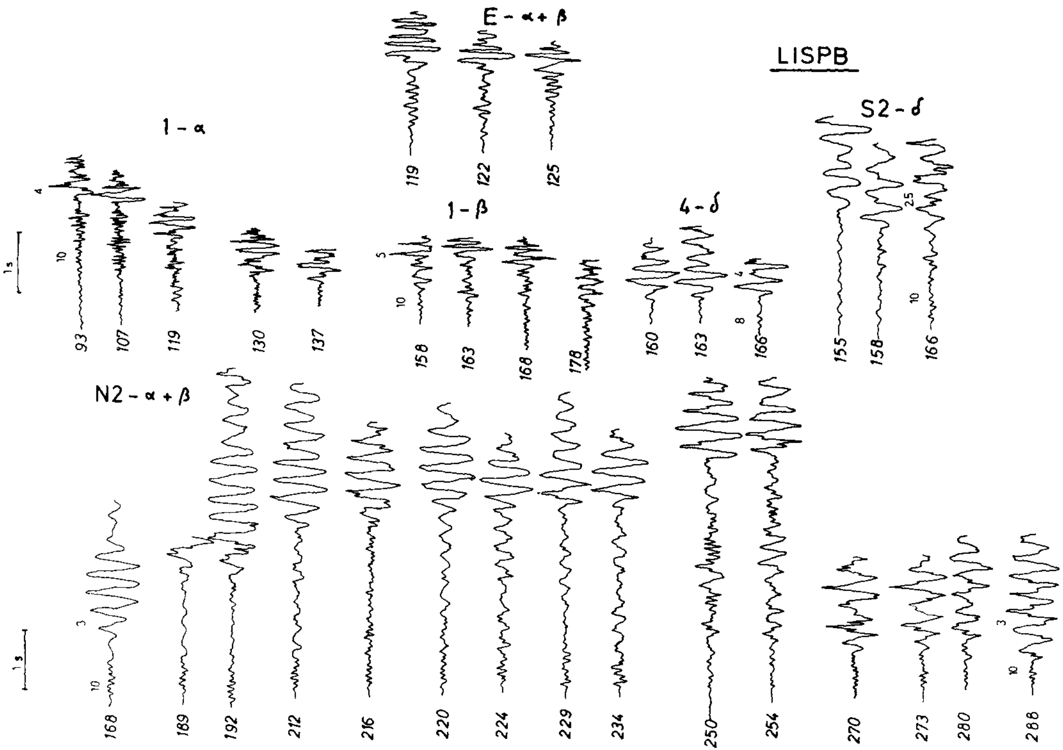

Fig. 10b. Examples of low-frequency $P_M P$ phases following high frequency precursors from the LISPB seismic experiment in Britain (Bamford et al., 1976)

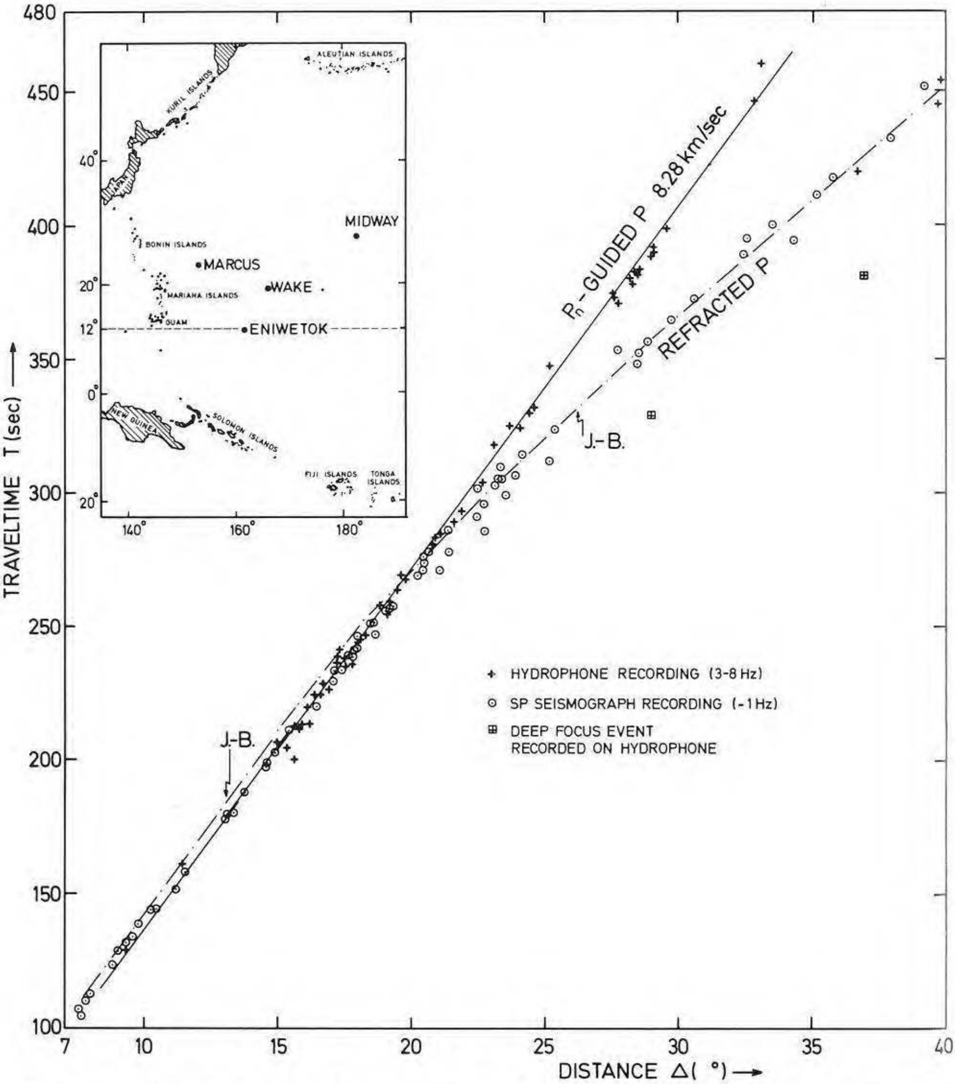

Fig. 11. Time-distance curve for P-waves in the Pacific as reported by Sutton and Walker (1972) and Walker and Sutton (1971). Inlet position map with the recording sites at Marcus, Eniwetok, Wake and Midway

very much the same manner as a low Q zone. Therefore, the presence of thin high-velocity layers in the lower lithosphere is an alternative explanation for the frequency behaviour of P phases refracted through the mantle.

5. A New Model of the Lower Lithosphere

The proposed mechanism of tunneling through high velocity strata in the lower lithosphere does not require that these layers have an infinite extent. A series of

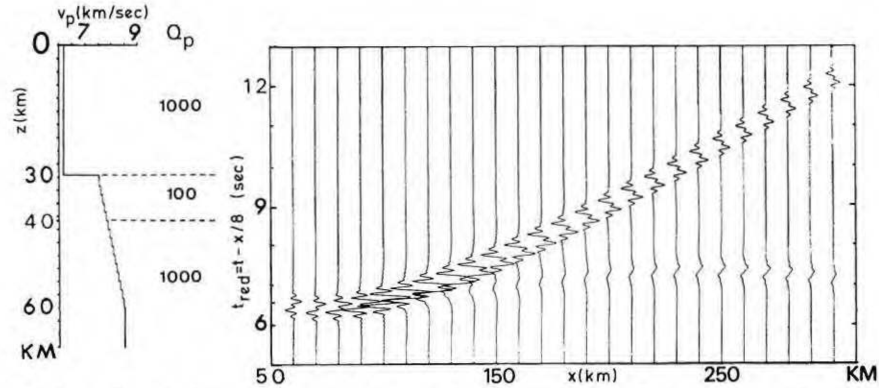

Fig. 12. Synthetic seismogram section for wave field reflected and refracted from a first-order discontinuity over a gradient zone. The first 10 km of the gradient zone posses a low $Q_p = 100$. This low Q zone attenuates high frequencies at short distances on the refracted P-branch. At larger distances high frequencies start to be visible again

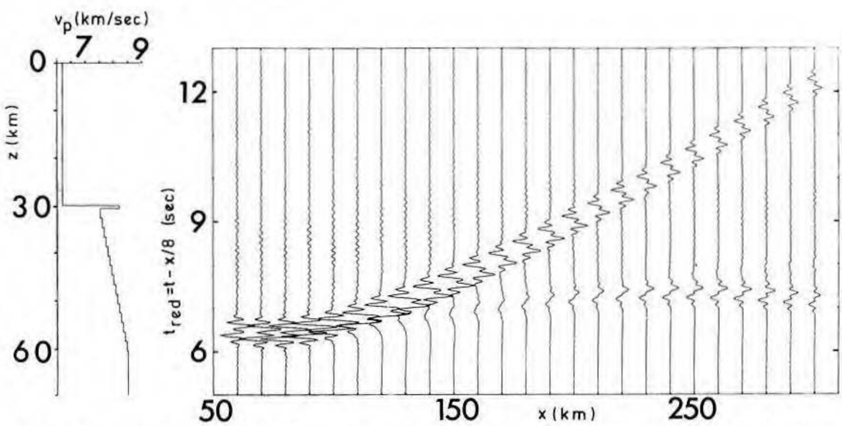

Fig. 13. The same model as in Figure 12, but with $Q_p = \infty$ and a thin high-velocity layer on top of the gradient zone ($h = 625$ m, $v_p = 8.2$ km/s). The low-frequency tunnel wave is now on the prograde branch of refracted P. High frequencies start to arrive on this branch beyond 200 km

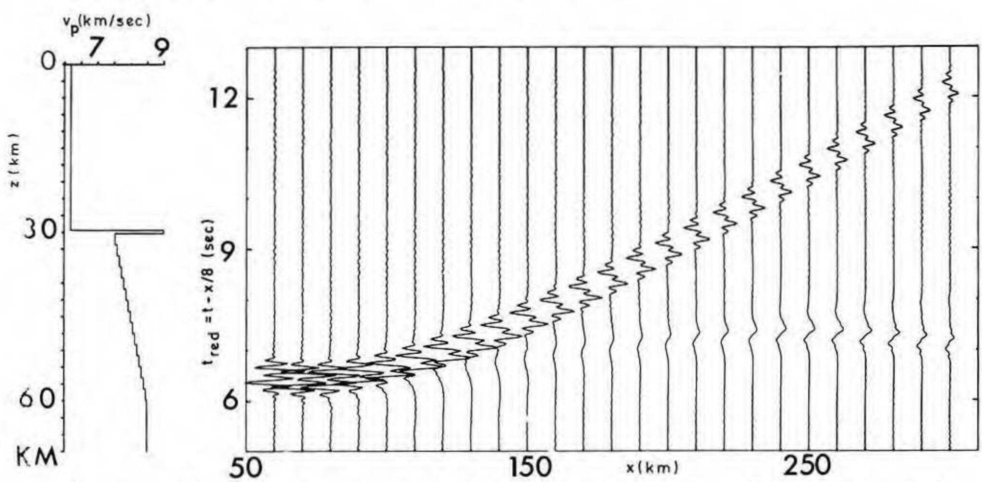

Fig. 14. The same model as in Figure 13, but with $v_p = 9.0$ km/s in the thin layer. High frequencies on the refracted P-branch are strongly suppressed. The wave field is very similar to that in Figure 12

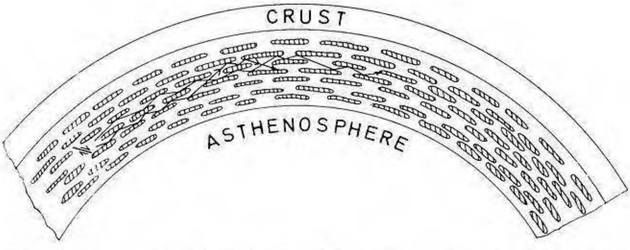

Fig. 15. Schematic seismic model of the lower lithosphere. The model is not to scale. Thin high velocity lenses are imbedded in "normal" upper mantle material. Their lateral extension is estimated to a few tens of kilometres, their thickness to a few hundred metres to a few kilometres. This model does not only explain the generation of tunnel waves it also serves as a possible waveguide for teleseismic transmission of high-frequency P_n-waves

thin parallel lenses of finite horizontal extent would produce a similar effect. A schematic model of the lower lithosphere including such lenses is depicted in Figure 15. The figure is not to scale. The fine structure of the crust and the deeper interior is not indicated. In the lower lithosphere thin high-velocity strata (dashed lenses) are imbedded in material with "normal" velocities, the latter is seen by long-period body and surface waves. Such a structure would also reproduce the high-velocity low-velocity layer alternation observed by Hirn et al. (1973) and Kind (1974). The high-velocity strata should be visualized as extending over only a few tens of kilometres, possibly less. Their thickness is estimated to about 0.5 to 1 km. However, an aggregate of thinner layers concentrated unevenly throughout the lower lithosphere could produce the same result. The high-velocity lenses should be oriented predominantly parallel to each other and more or less horizontally.

Such a model of the lower lithosphere could also explain the teleseismic transmission of high-frequency P_n and S_n waves, to distances of 30° or more (see Fig. 11) (Båth, 1966, 1967; Molnar and Oliver, 1969; Walker and Sutton, 1971; Sutton and Walker, 1972). Teleseismic P_n and S_n transmission is especially effective if the hypocenters are located in the lower lithosphere. In this case P and S waves are overcritically reflected between the high-velocity layers. While the low frequencies tunnel through the thin layers, the high frequencies in the range 2–8 Hz are trapped between the high-velocity layers. Higher frequencies are scattered by inhomogeneities smaller than the high-velocity lenses and leave this part of the upper mantle. Thus the lower lithosphere becomes a wave guide where a high-frequency pass band of P and S waves is propagated by overcritical reflection. S transmission is especially effective since no converted waves can penetrate through the thin layers to leave the wave guide. The frequencies in the pass band are determined by the thickness of the thin layers, their velocity and by the dimensions of scatterers in the wave guide.

Phase velocities of P-waves of up to 8.8–9.0 km/s in thin layers correspond to even higher velocities in the infinite medium (Press et al., 1954). Such high velocities in a depth range of 30–150 km are very likely caused by strongly anisotropic minerals such as olivine with a preferred orientation. Strong seismic anisotropy of the upper mantle with an azimuthal dependence of the seismic velocities is known from the oceans (Raitt et al., 1969) and was recently discovered also in the continental upper mantle (Bamford, 1973). It is probable that the same mechanism which is generating the preferred orientation of anisotropic minerals is also responsible for the horizontal orientation of the thin high-velocity layers. Creep in regional stress fields can induce such a preferred orientation.

6. Conclusions

Tunnel waves provide new evidence for the presence of high velocities in thin layers within the lower lithosphere. The tunneling of low-frequency waves through thin layers is studied by numerical experiments. Tunnel waves are observed in explosion seismic data and may also be present in teleseismic P observations. The model of a lower lithosphere with thin high-velocity layers serves also as explanation of teleseismic P_n and S_n observations and is related to recent observations of upper mantle anisotropy.

This model requires further verification. It may be tested with a number of experiments. 1. Spectra of teleseismic signals should regularly show a strong decrease in high frequency energy as soon as the rays pass from undercritical to overcritical incidence from the asthenosphere to the base of the lithosphere. 2. Presently only group velocities of P_n and S_n waves are reported. Phase velocities of P_n and S_n waves as measured across large arrays should show similar velocities as measured by phase correlation on closely spaced stations on long-range profiles of explosion seismic experiments. 3. Earthquakes should be observed also with mobile stations to obtain better control on S-wave velocities.

This seismic model of the lower lithosphere provides also new bounds to petrological models. The velocities are definitely larger than admitted by some authors (e.g. Green and Liebermann, 1975) who restrict the velocities in this depth range to not more than 8.25 km/s. The presence of low-velocity zones throughout the lower lithosphere (see e.g. Kind, 1974) raises the question whether the transition from the lithosphere to the asthenosphere is sharp or extends over a considerable part of the lower lithosphere.

It must be concluded that by the explosion seismic investigation of the lower lithosphere new mechanical properties of this part of the upper mantle have been detected which are of considerable importance to geodynamical problems.

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