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The upper mantle under Europe: an interpretation of some preliminary results from the NARS project

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Abstract

Using teleseismic recordings of 5 strong Japanese earthquakes in 1983/84, Dost (in press) has recently determined average higher mode phase velocities with the broad-band digital NARS array. These are the first such results ever to be obtained for a single geophysical province, in this case the west European platform. Moreover, the frequency range has been doubled with respect to earlier higher mode phase velocity determinations.

It is to be expected that we will refine and extend our measurements in the coming years.No strong earthquake at intermediate depth has so far occured since the installation of the NARS network, which leaves important gaps in velocity measurements at several frequencies. Nevertheless we have made a preliminary interpretation of the data now available, giving us guidelines for future observational efforts.

Inversion of the phase velocities for an average velocity/density structure for the west European platform results in an upper mantle with a low velocity channel between 120 and 250 km that is not very pronounced (average Vs is 4.35 km/s) and with remarkable layer of high velocity and density at roughly 100 km depth. Comparison with older higher mode data enables us to determine the difference in shear velocity between the Scandinavian shield and the west European platform: shield Vs is 4.49 km/s averaged over 120–250 km.

The velocity difference between shield and platform can be explained by a temperature difference of at least 250 degrees, or effects of partial melting under the platform.

We have extrapolated laboratory values of shear velocity and density of single crystal specimens of 4 different rock types (olivine, orthopyroxene, clinopyroxene and garnet) to ambient upper mantle pressure and temperature. Despite the uncertainty inherent to such a procedure, comparison of these curves with the inversion results suggests strongly that a layer of eclogitic composition is present between 80 and 120 km depth.

Introduction

One may judge the relevance of seismological data by their ability to resolve small scale detail deep into the Earth. Classical seismology was mainly concerned with the interpretation of the travel times of well defined P- and S phases on the seismogram. Observations come from a global network that reports to international agencies, and that ultimately grew to more than 1000 contributing stations today. There are many shortcomings to this network: stations have often widely differing instrument characteristics and recording is in analogue form, but it has yielded vital information on the radial structure of the Earth. The data base from the International Seismological Centre (ISC) now spans more than 20 years. However, classical seismology has never given us the spatial resolving power to bridge the gap between structures of planetary interest at a scale of 10^3 km, to the structures of geodynamical interest with a length scale of, say, 50 km. Some recent developments are changing this picture, however.

Increased computing power, and advanced numerical techniques (Nolet 1985), enable us now to use this database for imaging purposes, a technique which has become known as 'seismic tomography'. Tomography is able to resolve detail of a few hundred km in the mantle, although in areas of high seismicity and station density, such as the Mediterranean, this can be brought down to 100 km (Spakman 1986, this issue). With dedicated arrays even higher resolving power is possible (PASSCAL, 1984). The result of the imaging is a 3-dimensional model of the compressional velocity variations in the Earth.

Furthermore, seismologists have greatly increased their understanding of the more complicated parts of seismograms. Reverbatory sequences arriving after the S wave can be identified as a collection of waves channelled between the Earth's surface and the large velocity gradients near 400 and 650 km depth (e.g. Nolet 1982). These are higher harmonics of surface waves commonly called 'Love' and 'Rayleigh' waves, depending on the type of their particle motion. Records from classical seismological observatories are usually not sensitive enough at the low frequencies where these waves can be analysed. Nolet (1975, 1977) was, however, the first to show that the phase velocities of higher modes could be obtained from the records of the World Wide Standard Seismograph Network (WWSSN) by using a number of stations collectively as an antenna. These observations yield models for the S-velocity in the upper mantle, which are averaged over the horizontal length of the station array, but resolve the horizontal layering within 50 km depth intervals or better. Sufficiently accurate measurements of the dispersion of higher modes are able to give information on the magnitude of density gradients in the Earth (Nolet 1978). Experience so far has shown that many regions on the Earth are sufficiently homogeneous in the lateral direction to let the higher modes propagate in a coherent fashion. Thus, measurements are not hampered by structures of geological scale (say 100 km); on the other hand, local geological structures will not be resolvable with higher modes either.

The NARS network (Fig. 1) is a digital seismograph network that was specifically set up to overcome the defects of WWSSN type of equipment. The station density of NARS enables us to measure higher modes over smaller distances than anywhere else available: the results are representative for the west European platform, over which the stations are distributed. The broad-band instrument response and the digital recording were designed for observation of higher modes to relatively high frequencies, in order to increase the vertical resolution. The network is operational since 1983. The scientific aims have been described by Nolet & Vlaar (1982), the more technical details of the array can be found in Dost et al. (1984). In a recent paper, Dost (in press) has published preliminary results from Rayleigh mode phase velocity measurements over the array. This paper is a first attempt to interprete these preliminary results, primarily to give us guidelines for future observational efforts and partly to reformulate the scientific aim of the project in view of these findings and recent developments in theories concerning the constitution of the Upper Mantle.

Recent upper mantle models

Among geophysicists and petrologists there is still lack of consensus about the exact nature of the upper mantle under continents. Various models have been proposed. Some of these models are characterized mainly by their petrological and mineralogical properties (the pyrolite and eclogite models), others (the tectosphere model and the model of lithospheric doubling) are also more or less defined by mechanical properties. As a consequence, different models can sometimes be combined into one. In this section we give a brief

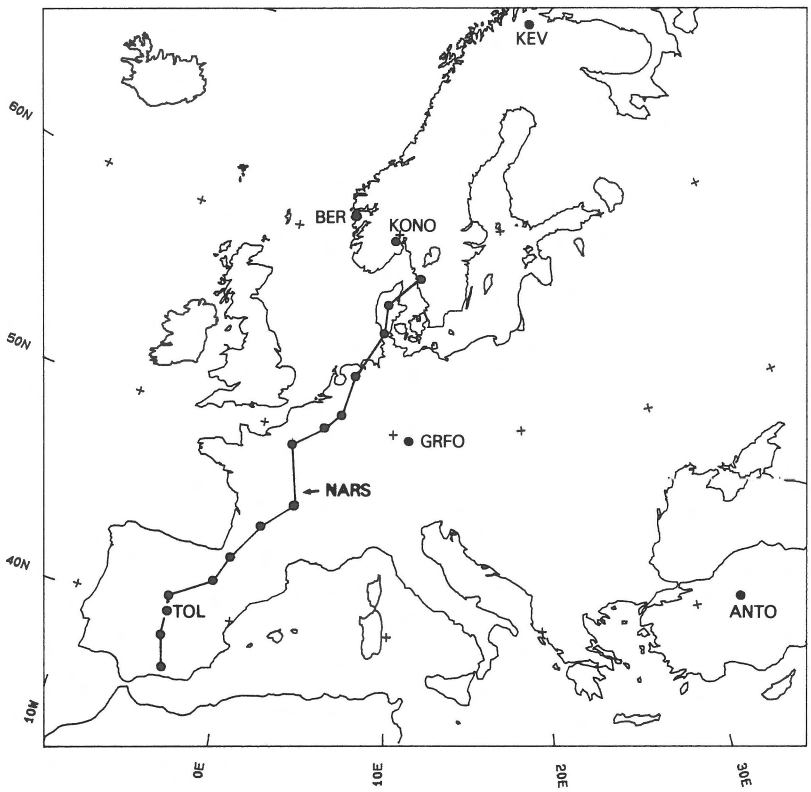

Fig. 1. Stations belonging to the network of Autonomously Recording Seismographs (NARS), and the digital seismograph stations belonging to the Global Digital Seismograph Network (GDSN) in Europe

overview of upper mantle models that are currently under discussion.

Pyrolite models

Ringwood (1975, 1982), Weidner (1985) and others proposed an upper mantle based on a theoretical undepleted mineral assemblage called 'pyrolite', which would yield a basaltic magma upon partial melting, leaving a residue of an Alpine type pyridotite. Pyrolite consists of about 57% olivine [o1, $(Mg,Fe)_2SiO_4$], 17% orthopyroxene [opx, $(Mg,Fe)SiO_3$], 12% clinopyroxene (cpx, containing Ca, Al and Na) and 14% garnet [ga, $CaMg_2Al_3Si_3O_{12}$]. The '400 km discontinuity' is explained by the phase transitions from opx to ga at a depth of about 375 km, involving a density increase of 10%, and the breakdown of forsterite to the β-phase near 400 km. Akaogi & Akimoto (1977) claimed that the pyroxene to garnet transformation is partly responsible for the 400 km discontinuity, although the solid solubility of enstatite [$MgSiO_3$] in pyrope [$Mg_3Al_2Si_3O_{12}$] is a gradual process that starts already at 150 km depth and increases suddenly at pressures between 140–175 kbar (1000° C), resulting in the formation of a homogeneous garnet. Near 525 km, the β-phase transforms to a spinel structure, which itself breaks down between 650 and 700 km to perovskite [$(Mg, Fe)SiO_3$] and periclase [$(Mg, Fe)O$]; at this depth, pyrope garnet transforms to the ilmenite structure. There is some doubt, however, whether the 650 discontinuity can be explained as a multivariant phase change, in view of the extreme sharpness of this boundary, which has been established with NARS data (Paulssen 1985).

Ringwood (1982) defined the differentiation of a pyrolite oceanic mantle model in a more precise manner. During partial melting, a lithosphere builds up which is more depleted as it is closer to the surface. Thus, the basaltic crust tops a layer of residual harzburgite, which consists of 78% olivine and 22% orthopyroxene. Between 30 and 40 km depth some clinopyroxene is already present, and the assumed proportions are 65% olivine, 35% pyroxene. Below 40 km the pyrolite has been subjected to a very small degree of partial melting (<2%) and is therefore depleted in the most incompatible elements (Rb, Ba, La, U) only. The zero pressure densities of harzburgite and pyrolite are 3.32 and 3.38 gr/cm^3 respectively.

Upper mantle models of this type exist in several slightly different versions. All of these are based on an upper mantle which consists chiefly of a peridotite, in which pyroxenes are the second most abundant group of minerals next to olivines. More particularly, the uppermost part of the mantle consists of a spinel lherzolite and the deeper part of the upper mantle has a garnet lherzolite composition. This view is widely accepted, but several models exist that differ from this in fundamental aspects:

Eclogitic models

Press (1969) proposed a mechanism based on eclogite fractionation to explain a layer of low densities found beneath ocean ridges. His mechanism made use of the solid-liquid phase transition of eclogite; the eclogite that is present in the asthenosphere is the source of a basaltic melt. This melt, which is lighter and more mobile than the peridotitic residue, migrates upwards, where it cools and solidifies into eclogite. Eclogite has a high density, probably in excess of 3.5 g/cm^3. Nolet (1977) invoked this mechanism to explain a negative density gradient in his upper mantle model for western Europe near 200 km depth. Cara et al. (1980) showed, however, that a high density 'cap' in the upper mantle was not needed to explain newer higher mode data then available.

Anderson (1979, 1984) also argued that eclogite is a major constituent of the upper mantle: he pointed to the fact that the volume of the lunar crust, which represents 10% of the moon, contrasts sharply with the figure of 0.5% found for the Earth (it may actually be close to 1% of the volume of the Earth's mantle). In his view, the heavy eclogite sinks and accumulates at the 650 km discontinuity, which it cannot penetrate because of the presence of the light Al_2O_3. In this view the Lehmann discontinuity near 220 km should have a positive density gradient and represent the top of the eclogite layer. The 650 km discontinuity implies the chemical transition to a lower mantle with (Mg, Fe)SiO_3 in the perovskite structure.

The tectosphere model

Jordan (1978, 1981) pointed to a fundamental difficulty in the efforts to extend the cooling lithosphere model, so successful in explaining the evolution of the oceans, to the old cratonic parts of the continents. In oceans, increased water depth maintains isostatic balance when the lithosphere cools and its density increases. Continued cooling of the lithosphere of continental cratons would result in a thickness of about 300 km. There is ample seismic evidence, notably from higher modes and ScS waves, that the oldest cratons exhibit a distinct, presumably cool, structure to this depth. However, in order to keep the mantle in isostatic balance by mere crustal thickening, the depth to the Moho should increase by 10–15 km, which is contradicted by seismic observations from refraction experiments. Jordan (1978, 1981) proposed therefore that the old shields possess a continental 'root', which consists primarily of peridotite, depleted in major-element basaltic constituents. Such a root could be formed by 10–20% partial melting. If the melt migrates upward, the residue would be 1 or 1.5% lighter, compensating for the temperature effect. Moreover, the lower density of the root would stabilize it against disruptive convective motions, explaining the strong correlations between upper mantle seismic velocities and surface geology. This model is also supported by evidence from South African xenoliths, which on average are 1.3% less dense than rock of standard pyrolite composition.

Lithospheric doubling

Vlaar (1982, 1983), Vlaar & Cloetingh (1984) considered the implications of quasi-horizontal subduction of young oceanic lithosphere, which they proposed to be a quite general phenomenon in plate tectonic processes of the recent past. Horizontal subduction will result in a doubling of lithospheric structure. Taking Ringwood's differentiation model for the lithosphere, this would result in a heavily layered upper mantle. The density would increase in the top lithosphere, but there might be a density drop at the depth of the second harzburgite layer. The velocity structure is difficult to predict, as laboratory data on this point are rather inconclusive.

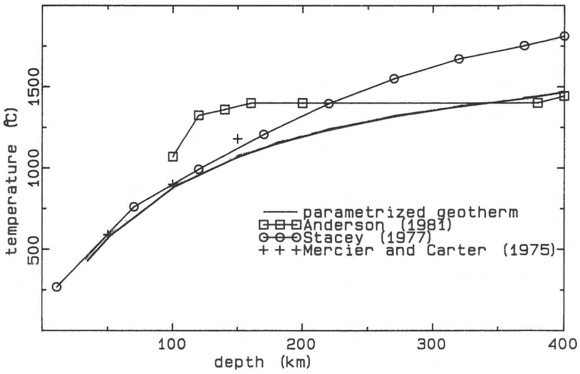

Fig. 2. Temperatures in the continental upper mantle as proposed by Stacey (1977), Clark & Ringwood (1964) and Mercier & Carter (1975) compared to the more general Earth geotherm of Anderson 1981). Also shown is the parametrized geotherm described in the text.

Elastic constants

Recently, there have been new attempts at comparing seismic velocity profiles with laboratory elasticity data, with a most remarkable lack of consensus: Bass & Anderson (1984) and Anderson & Bass (1984) concluded that the preliminary reference Earth model (PREM, Dziewonski & Anderson, 1981) is incompatible with seismic velocities and density predicted for a pyrolite model, and advocated an eclogitic composition beneath 400 km (and possibly shallower) depth. Weidner (1985), on the other hand, found that a pyrolite model with a 1400° C foot to the adiabat can adequately model the seismic velocities throughout the upper mantle. This discrepancy is – at least partly – due to lack of adequate laboratory data, and illustrates the difficulties that seismologists still face when trying to interpret their results in terms of petrological models.

Attemps to model the region above 200 km depth are even more dangerous, since here the temperature gradient exceeds by far the adiabatic gradient which is given by

$$\tau = \frac{\gamma T g}{\varphi}$$

With the Gruneissen parameter $\gamma = 0.8$, gravity $g = 10\,m/s^2$ and bulk velocity $\varphi = 3.86 \times 10^7 m^2/s^2$ at the top of the mantle, τ will be close to 2.1×10^{-7} T K/m, where T is the temperature. In view of uncertainties in γ and T, we estimate that τ in the upper mantle lies between 0.2 and 0.3 K/km. Some proposed upper mantle temperatures are depicted in Figure 2.

The actual temperature gradient is of the order of 5 K/km and it is obvious from Figure 2 that one cannot apply finite strain theory along an adiabatic geotherm in the upper part of the continental mantle. Instead, we have to resort to laboratory measurements of seismic velocities at room temperature and pressure (RTP), and their isothermal and isobaric derivatives with respect to P and T, and extrapolate these along an assumed geotherm. This is the approach we follow in this paper to construct theoretical velocity curves for mantle minerals.

There are several problems associated with this approach. First of all, the geotherm may be uncertain to 200 or 300 degrees in this part of the mantle. We have adopted a parametrized geotherm of the form:

$$T = \frac{10^4 z}{580 + 5.4z}\ °C$$

which gives a reasonable fit for $50 < z < 400$ km to the average of the models in Figure 2: in the upper 200 km it follows closely the curves and data that are typical for continental geotherms, at the bottom of the interval it coincides with Anderson's more general geotherm.

Secondly, laboratory measurements for different rock specimens may differ widely in value. Anderson et al. (1968) stressed the point that laboratory data on polycrystalline compounds are compatible with single crystal measurements only when the porosity is low. For the interpretations in this paper we therefore rely fully on single crystal data. The most recent compilation of these data has been published by Sumino & Anderson (1984). Their values for the four most important upper mantle minerals are summarized in Table 1. The data in Table 1 enable us to determine the density along the parametric geotherm:

$$\frac{d\varrho}{dr} = -\frac{g\varrho}{\varphi} + \alpha\varrho\tau$$

where φ is the bulk sound velocity, ϱ the density, α the termal expansivity and τ the superadiabatic temperature gradient.

In order to estimate the velocities along the profile, we must integrate:

$$V(T,P) = V_O + \left(\frac{\partial V}{\partial P}\right)_P dT + \left(\frac{\partial V}{\partial P}\right)_T dP$$

along the parametrized geotherm. Experimental evidence indicates that the partial derivatives themselves are not constant, but vary with temperatures. Accurate laboratory values on all of the rocks in Table 1 are not available to our knowledge. Soga (1967) determined a theoretical curve for garnet, which is well parametrized by a relationship of the form:

$$\left(\frac{\partial V}{\partial T}\right)_T = \left(\frac{\partial V}{\partial T}\right)_0 [1 + aT^{1/2}]$$

with a = 0.0175. In view of the large uncertainties inherent to these temperature corrections, we do not strive for higher precision, and have adopted the same correction for all rock types. We note that Soga's results fall below the experimental values determined by Fielitz (1976) for eclogites and peridotites (Fig. 3). For lack of better data, we have to accept an uncertainty of a factor of 2 in the temperature corrections. Neither Soga nor Fielitz's paper give any information on the temperature correction at near-solidus conditions, where a severe deterioration of the shear modulus may be expected (Goetze 1977). We have, therefore, not tried to model any of these effects, nor the even more drastic effects to be expected with partial melting. We further assumed $\partial V/\partial P$ to be independent of temperature.

The computational results for olivine are given in Figures 4 and 5. In Figure 4 we see that the density increase caused by pressure effects alone (along an adiabatic temperature curve) is approximately offset by the effects of temperature. The largest uncertainty in this curve comes from the uncertainty in the geotherm. In Figure 5 we see that temperature has an effect on the shear velocity which is about twice as large as the effect of pressure. Here the largest uncertainty comes from the uncertain extrapolation of $\partial V/\partial T$ from room conditions to the geotherm.

However, in spite of the uncertainties in the absolute level of the density and the velocity, Figures 4 and 5 show that the changes to be expected as a result of temperature and density increase

Table 1. Mineral properties used in the calculations.

	Olivine Ol (7.2% Fa)	Orthopyroxene Opx (20% Fe)	Clinopyroxene Cpx (0% Fe)	Garnet Ga (16% Fe)
Vp (km/s)	8.422	7.781	8.078	8.921
Vs (km/s)	4.887	4.719	4.864	5.002
Density (gr/cm^3)	3.311	3.354	3.198	3.705
$\partial Vp/\partial T$ (10^{-3} km/s/K)	−.487	−.633	−.633	−.433
$\partial Vs/\partial T$ (10^{-3} km/s/K)	−.341	−.265	−.265	−.188
$\partial Vp/\partial P$ (km/s/Mbar)	10.2	20.4	20.4	75.2
$\partial Vs/\partial P$ (km/s/Mbar)	3.63	5.13	5.13	2.50
α (10^{-6}/K)	24.65	47.7	22.5	19.2
$\partial\alpha/\partial T$ (10^{-8}/K^2)	4.7	4.5	4.5	2.6

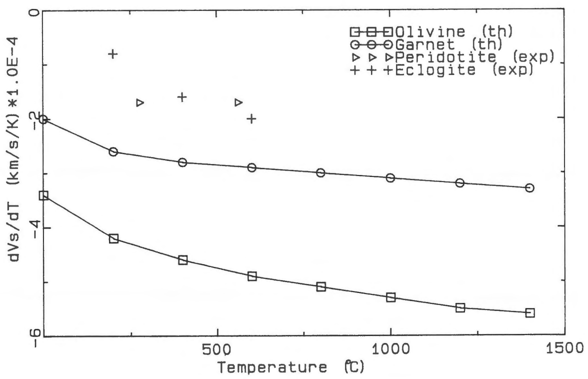

Fig. 3. Assumed temperature dependence of dVs/dT for garnet and olivine, as extrapolated theoretically (th) from laboratory values at T = 0, and comparison with experimental (exp) data from natural rock samples from Fielitz (1976) on peridotite and eclogite.

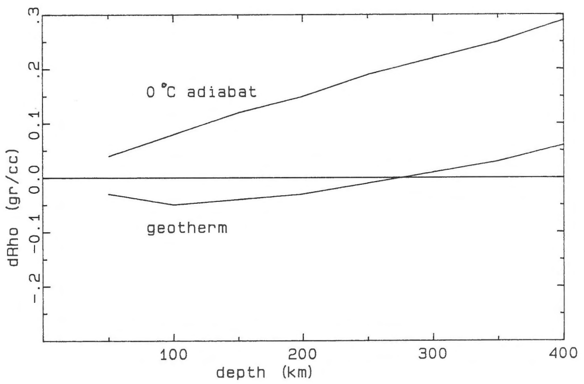

Fig. 4. Extrapolation of the density of olivine along the 0 degree adiabat, and along the geotherm. These curves are much more accurate than the velocity curves in Figure 5.

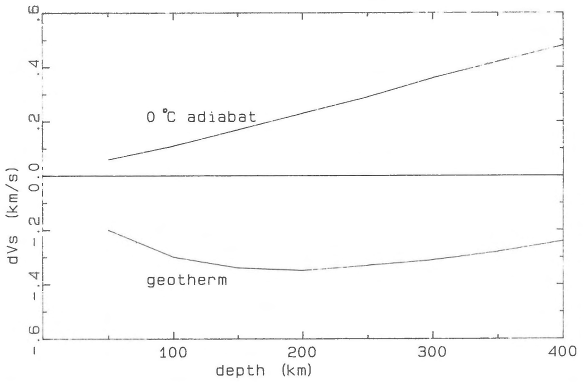

Fig. 5. Extrapolation of the S-velocity of olivine, along the 0-degree adiabat, and along the geotherm (taking both effects of pressure and temperature into account). The bottom curve is very uncertain because of the uncertain value of dVs/dT, but the fact that the curve is very smooth is significant.

alone are not very drastic. This is an important observation, since it indicates that large density or velocity gradients must have been induced by a change in mineralogy, chemical composition or indicate partial melding.

A petrological interpretation

Dost (in press) has attempted a preliminary interpretation of the first higher mode dispersion measurements resulting from the NARS experiment. Most of the effort in obtaining these data was spent in extending the frequency interval of observation toward much higher frequencies than hitherto available, and little attention has so far been paid to the problem of resolving the very low frequencies. Also, no strong Japanese earthquake at intermediate depth (70–200 km) has occurred in the first two years of NARS, which has hampered the observation of some modes (especially the first higher mode). Thus, there are still important gaps in the higher mode phase velocity measurements.

Nevertheless, the results now available constitute already a significant step forward with respect to the earlier results obtained by Nolet (1975, 1977) and Cara et al. (1980). The frequency band of observation has been doubled, and now extends to more than 60 mHz. More important, all earlier investigations were forced to consider the Scandinavian Shield and the west European Platform as one unit, but the NARS data are representative for the platform only; the age differences in the upper mantle are expected to be much smaller for this more limited 'geophysical' province. Moreover, by comparison of the new results for the platform with the shield-platform model M7 from Nolet (1977), Dost (in press) was able to derive a model, termed SCSH1, for the Scandinavian Shield mantle. His results are reproduced in Figure 6.

When comparing the velocity model with the Vs predicted for various minerals, the platform velocity model (WEPL1) falls below any of the predicted velocities except for clinopyroxene (Fig. 7). A geotherm which is some 300 degrees above the adopted geotherm could bring the predicted velocities in line with the WEPL1 values, but this does

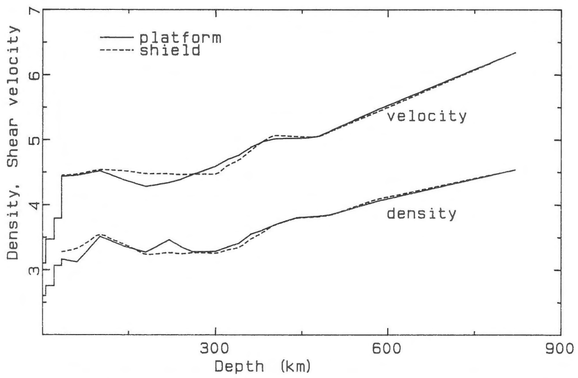

Fig. 6. Models WEPL1 (platform) and SCSH1 (shield) from Dost (in press). Density is in gr/cm³, shear velocity Vs is in km/s.

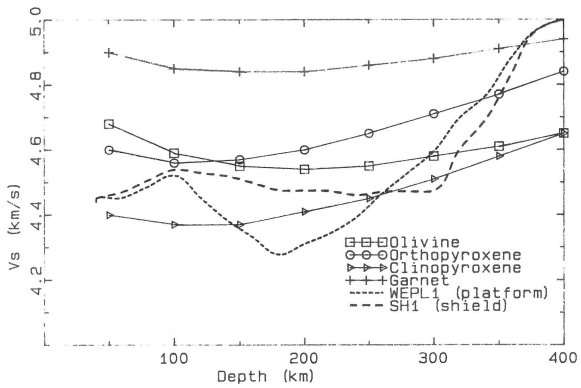

Fig. 7. Comparison of the velocity models with laboratory extrapolations.

not seem very likely since the discrepancy is already present below the Moho, where temperatures are somewhat better constrained. With our adopted $\partial Vs/\partial T$ on the high side already, the extrapolation itself cannot be the direct cause of the discrepancy. We suspect that secondary effects, such as partial melting or perhaps other fluid inclusions not accounted for in the extrapolation, play a role. This is corroborated by the fact that the discrepancy is largest at the depth of the low velocity channel, which is widely believed to be caused by partial melting.

Although the absolute level of the velocity is difficult to compare with laboratory extrapolations, we have seen that large velocity gradients cannot be explained by simple temperature/pressure effects. The most pronounced gradient in model WEPL1 occurs just above 100 km depth. It coincides with a strong gradient in Vp found in refraction work in France (Hirn et al. 1973). Some of the regionalized models for France, deduced by Souriau (1981) from local fundamental Rayleigh modes, also show indications of a gradient in Vs in the uppermost mantle. Resolution analysis shows that the gradient in WEPL1 is well resolved, and in view of the coherence of the phase velocities over the full length of the platform, we must assume that it is a general, continuous phenomenon over the platform.

What are the likely explanations for such a velocity gradient? Since there is no widespread volcanism under western Europe, nor a pronounced thermal anomaly, we rule out that there should be partially molten rock just below the Moho. One possible explanation is to assume a largely peridotitic mantle, with an eclogitic layer near 100 km depth. Another possibility might be the presence of water-bearing minerals such as hornblende, in the first 30 or 40 km below the Moho. It is well known that even very small amounts of fluid may have a drastic effect on the shear modulus.

The velocities below 120 km are consistent with a peridotitic composition with some partial melting to produce the low velocity channel. The sharp bend near 300 km depth could be less pronounced in view of the resolving power of the data, in which case the velocity gradient below 250 km is well explained by a decrease in the amount of partial melting, and, especially at a somewhat deeper level, by the solid solution of pyroxenes into garnet.

Resolution analysis shows that the density model is not very well resolved, with the possible exception of the sub-Moho gradient. Thus, inferences drawn from this density model should not be considered conclusive. Figure 8 shows the density profile for model WEPL1 together with the extrapolations from laboratory data. The gradient near 100 km, which coincides with the velocity gradient, indicates that a substantial amount of higher density material is needed to explain the impedance of this layer, as detected by the higher mode phase velocities. The average density between 70 and 150 km is 3.41 g/cm³, which can only explained by a garnet content of at least 25%. There are no other

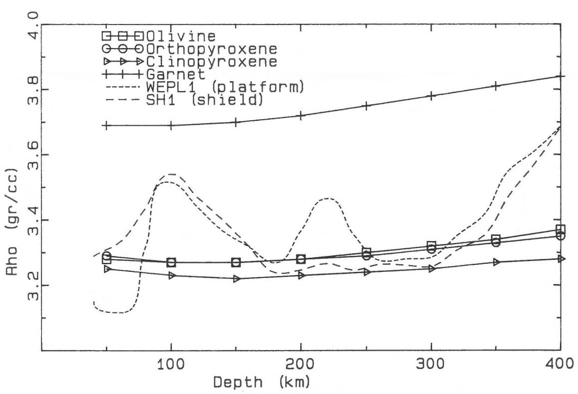

Fig. 8. Comparison of the density models with laboratory extrapolations.

likely minerals with a sufficiently high density (see Table 1). An eclogitic layer near 100 km depth would also explain the velocity gradient, and makes an explanation in terms of fluid inclusions beneath the Moho unnecessary.

The interpretation of the shield model SCSH1 must be done with some care, since a 'difference'model is by its very nature less well constrained than the original models (M7 and WEPL1). The most pronounced difference between the models is the shear velocity, which is everywhere higher under the shield. Between 120 and 250 km depth, the platform has a clear, although not very pronounced, low velocity channel with an average velocity of 4.35 km/s. By contrast, the velocity under the shield remains almost constant, at an average 4.49 km/s. If this is due to the effects of temperature alone, the difference must be at least 250 degrees if we assume $\partial Vs/\partial T = -5.0 \times 10^{-4}$ km/s/K (Fig. 3). For lower values of the derivative this temperature difference must be even higher. However, it is likely that other effects also play a role. Partial melting has already been invoked to bring the velocities below the laboratory estimations, and the possible absence of partial melting would of course explain much of the velocity difference. The fact that the density in the shield model SCSH1 falls slightly below that of the platform model would indicate basalt depletion, and this would make it even more difficult to explain the velocity difference by the effect of temperature, since at the same temperature the shield velocities are expected to fall below the platform values.

Discussion

The most surprising result of this investigation is the indication of an eclogitic layer near 100 km depth. From the theories on the upper mantle, which we reviewed in section 2, only Vlaar's concept of lithospheric doubling (Vlaar 1982, 1983) could provide the necessary mechanism to produce such a layer: it would be the high-pressure form of a basaltic layer that tops obducted oceanic lithosphere. Although such a layer would be unstable, the density difference by itself might not be enough to disrupt it, and the layer could remain in place for a very long time, or even 'freeze in' (Christensen, pers. comm. 1985).

The absence of a low velocity channel under the shield confirms the presumption made by some seismologists, that the upper mantle under shields is cooler to sufficiently large depth to preclude the possibility of a true low velocity channel. Recent evidence from waveformfitting methods to higher modes travelling over the northern edge of the Eurasian continent seem to point in the same direction (Lerner-Lam, pers. comm. 1985).

Much of the uncertainty in the above interpretation is due to the paradoxical situation that velocities, which can be determined with a reasonable resolution in the upper mantle, are difficult to extrapolate from RTP laboratory data to ambient mantle conditions; whereas densities, for which the laboratory extrapolations can be done with much more confidence, are difficult to determine in the Earth. It is not probable that we can soon improve on our extrapolation of laboratory velocity measurements (even if we could, there would be too many variables to deal with). There are, however, several promising possibilities to improve on our estimation of density in the upper mantle. First of all, we can extend the data set toward lower frequencies. If events at intermediate depth occur, these will certainly help to constrain the phase velocities of the first two higher modes better than the present data set. We can make an effort to extend the data set with group velocities and local S-velocities, which are both known to improve on the density resolution (Nolet 1978). And finally, we can use new techniques of non-linear waveform-fitting (Nolet et al. 1986) to take the effects of lateral heterogeneity into account. All this will be subject of future research associated with the NARS project.

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