

INVESTIGATION OF THE 600-KM DISCONTINUITY UNDER FRANCE THROUGH TRAVEL-TIME AND AMPLITUDE ANOMALIES

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In this paper, we give a model of the 600-km discontinuity under France, which explains large travel-time and amplitude anomalies on teleseisms as recorded in the French network. It features a large descent from the Ardennes to the Pyrénées, of which tectonic implications are discussed.

1. Introduction

With the advent of large seismic arrays, such as LASA or NORSAR, scientists have been able to improve their knowledge of the structure of the earth under receiving areas.

This knowledge leads to a fine structure which takes the form of a regional lateral inhomogeneity, since the mean structure of the earth has been known for a long time. Continental-size arrays may allow investigation of deeper heterogeneities beneath stations.

In the case of France, large anomalies, in travel-time data as well as in amplitude, leads us to propose a model for the 600-km discontinuity, featuring a large descent from 550 km in the northeast to 710 in the southwest.

2. The Toulx-Ste-Croix effect

The seismic station of Toulx-Ste-Croix, France (TCF), is part of the French seismic network (cf. Fig. 1). It is located on the Limousin plateau at $46^{\circ}17'17''\text{ON}$ by $02^{\circ}12'50''\text{OE}$, and is equipped with a short-period vertical seismometer. This station exhibits a very strange amplitude anomaly for events located in eastern Kazakh (EK) and Sin Kiang (SK): for a given particular event, let r be the ratio of the amplitude recorded in TCF over the mean average amplitude over France. For an event located in EK, $r = 0.42 - 0.45$, whereas for an event from SK,

$r = 3.0$. Thus TCF is seven times more sensitive to SK events than to EK ones.

Emphasis should be given to the following points: first, the ratios r are computed for every event and this behaviour has been checked every time an event occurs in EK or SK. Second, on the basis of an average of worldwide epicenters, station TCF is no more or less sensitive than the other stations of the French network. This anomaly is very puzzling, since the epicentral distances are $\text{TCF} - \text{EK} = 49^{\circ}$, $\text{TCF} - \text{SK} = 60^{\circ}$ and the difference in azimuth is only 4° . Thus the two rays from EK and SK to TCF have very close paths, especially in the region near the receiving station. Mechler and Rocard (1970) proposed to name this effect after station TCF.

A previous study in various regions in France, by Mechler and Rocard (1964), had shown that the amplitude recorded from a local event (up to 20°) could be increased by moving the station within a 10-mile wide area, and since then a proof was given that this was due to curvature of the Mohorovičić discontinuity, acting as a focusing or defocusing mechanism (Mechler, 1967, 1969). However, in the present case of the TCF effect, the Moho cannot account for it, since the two rays from SK and EK to TCF cross it at two points only 2.5 km apart. The laws of diffraction optics give a certain width to the rays and it is impossible, with a "light" of at least 6 km wavelength, to identify details that size. More precisely, diffraction theory shows (Mechler and Rocard, 1970) that these rays are no more really "separated" after they reach a depth

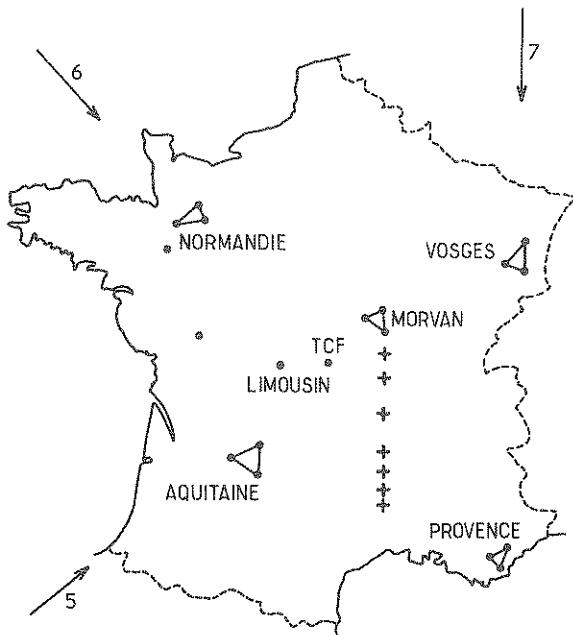


Fig. 1. Map of the French seismic network. The dots indicate permanent stations, the crosses portable stations installed for Cannikin; the arrows show the directions of the profiles drawn on Figs. 5, 6 and 7.

under France of 70λ , which means that any structure shallower than 420 km can in no case account for the TCF effect. On the other hand, the strong density of seismic stations in France (all of which are not reported in Fig. 1 because some of them did not turn out to be satisfactory sites for permanent stations, and some were portable stations) makes it impossible for the TCF effect to be explained by a structure located in the mantle close to the epicenter or even in the deep mantle near the turning point of the ray: the rays linking a given epicenter to TCF and any other French station close to it (about 60 km apart) would indeed not be separated until they reach the mantle under France; especially, at their turning point, they would be only 20 km apart. That way, we restrain the possible feature responsible for the TCF effect to the middle mantle under France. In order to give a theory explaining it, we make the following assumption:

The origin of the TCF effect lies in the uneven depth of the 600-km discontinuity (600 D) whose curvature acts as a focusing or defocusing mechanism. Such an assumption is obviously associated with travel-time anomalies. Hence, in the following sections, among all the possible models of lateral inhomogene-

ities which can be derived from our experimental travel-time data, we choose the one which varies the depth of the 600 D (section 3). It indeed turns out to give satisfactory explanations for the amplitude anomalies (section 5).

3. Mapping the 600 D through travel-time anomalies: procedure

In order to gather information on the possible irregularities of the 600 D, we studied the travel-time anomalies of the records in the French stations of 179 teleseisms of world-wide distribution. The travel-time anomaly can be written as:

$$\Delta T = o - c = A_{\text{epicenter}} + A_{\text{transfer}} + A_{\text{receiver}}$$

We eliminate anomalies $A_{\text{epicenter}}$ due to local structure under the epicenter by considering the differences $\Delta T_a - \Delta T_b$ of travel-time anomalies as recorded in two stations close enough that the rays be identical in the upper mantle under the epicenter.

As far as A_{receiver} is concerned, we may eliminate most of it by applying altitude corrections, limestone corrections for those stations which need it, and Mohorovičić corrections (Plantet, 1972). The latter are yielded by previous studies of local events, which led us to a fairly good knowledge of the Moho structure under our network. The influence of the dipping of the Moho is cancelled by averaging our results over triangles of stations roughly 30 km wide (Fig. 1). At that point, the remaining differences are:

$$\Delta T_1 - \Delta T_2 = (A_{\text{transfer}})_1 - (A_{\text{transfer}})_2 + \epsilon_1 - \epsilon_2$$

1 and 2 being subscripts naming triangles of stations and ϵ the errors in reading the arrival time. Furthermore, whenever possible, we averaged our data over a large number of events coming from the same seismic region; this has the effect of lowering ϵ to less than 0.2 sec. We then remain with an anomaly which we assume is due to the difference in depth of the two points where the mean rays going to triangles 1 and 2 cross the 600 D. The order of magnitude of $\Delta T_1 - \Delta T_2$ may be up to several seconds.

An extensive description of the inversion of these data is given by Mseddi (1972). Emphasis should be given to the following points:

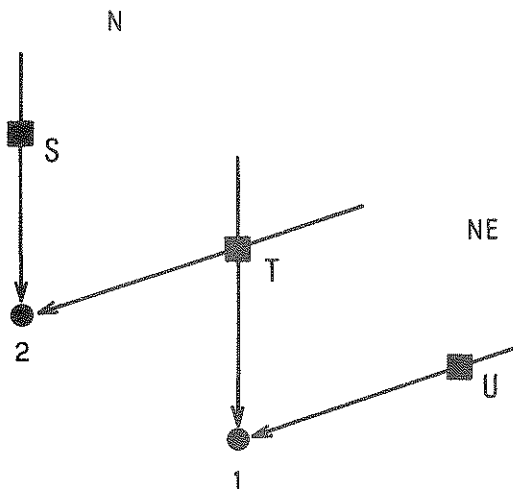


Fig. 2. Scheme of the closure checks: the dots indicate the stations, the squares the points where the rays cross the 600 D.

(1) We used the best available data about the velocity increase at the 600 D under Europe by using the model of Ansorge and Mayer-Rosa (1968). They give a profile featuring a jump from 9.4 km/sec to 10.2 km/sec at 550 km and velocities increasing with depth above the discontinuity as well as below it. When we considered a different depth for the 600 D, for example a deeper one at 650 km, we read directly the velocity at the deep end of the discontinuity from the lower part of their profile and we extrapolated the upper part of the profile down to 650 km for the velocity at the shallow end of the discontinuity.

(2) Our method is a differential method, in which we are able to obtain differences in depth between two points, which principle is illustrated in the scheme of Fig. 2. Given a starting point *S*, the study of the differences $\Delta T_1 - \Delta T_2$ for an epicenter located north gives us the depth of point *T* which in turn yields that of point *U* when an event comes northeast. This method enables us to have a closure check because we may, under happy matching, come back to point *S*. A number of these checks were performed, which yielded an accuracy of ± 20 km.

(3) As we said, the method needs a starting point, which we chose as 550 km under the Rozoy—Vervins regions of Ardennes, which is the point where a ray from Aleutian Islands to Morvan crosses the 600 D. This choice was made in order to link our map with German studies of the upper mantle (Ansorge and

Mayer-Rosa, 1968). Other values were tried, principally 500 and 600 km. However, the smallest standard deviations in the closure checks, as described in item (2), were obtained with the value 550 km which we therefore decided to choose.

4. Mapping the 600 D: results and improvements

Using the method described in section 3, we obtained a map of the 600 D. However, the inversion procedure used to compute the map assumed a horizontal discontinuity, whose depth would vary from point to point acting as a delaying process. As the purpose of the map is to exhibit a variation of depth of the 600 D, we must take into account that it is not horizontal, its dipping modifying the inversion procedure from the ΔT 's to the depth values (Plantet, 1972): the first map gives us the general trends of dipping, which are used in drawing an improved map, corrected for the dipping of the 600 D, which is the one shown on Fig. 3 (Mechler et al., 1971).

The main features of this map are the following:

First, the 600 D exhibits a descent from 550 km in the northeastern region of France (Ardennes) to 710 km under the Pyrénées; next, the curves of equal depth have a general NNW—SSE trend in the western part of France, bending to WNW—ESE in the east; finally, a very strong anomaly occurs in the Alsace—Vosges region of eastern France (cf. Fig. 4) where the 600 D dives from 520 to 650 over 140 km (Mechler and Rocard, 1971).

5. Further results from the map

As we already stated in section 2, our map is the result of an assumption we made when deciding that the 600 D should be the only feature responsible for the observed anomalies. Before indicating how our map may be used to account for the TCF effect, we show that the results to which it leads are consistent with other data: for that purpose we should use an entirely different method to obtain data about the depth of the 600 D at given points. Such a method may be the study of the triplicated hodochrones at distances close to 20° , for which the turning point of the rays is located at approximately 600 km. In

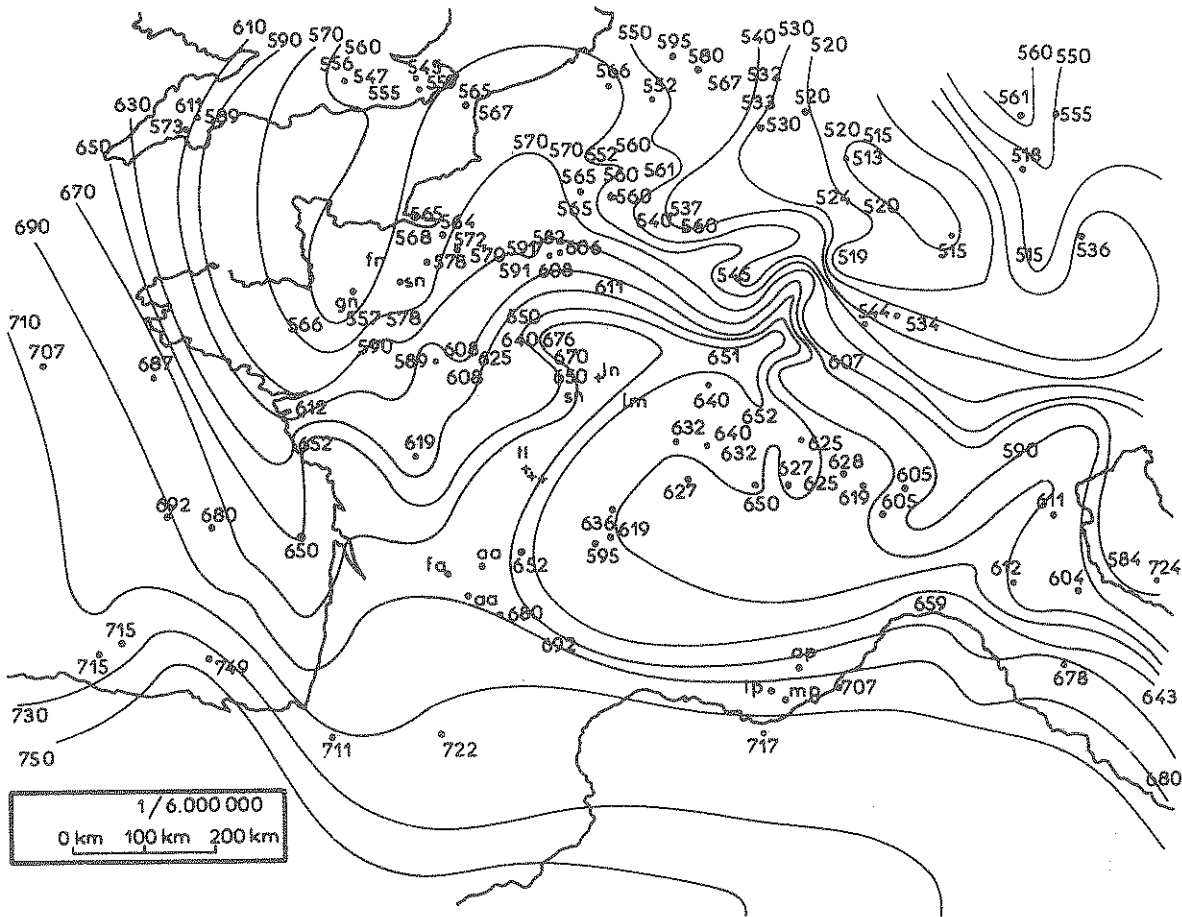


Fig. 3. Improved map of the 600 D under France and western Europe.

that particular case, the hodochrone is 5-fold because of the combined action of the 350- and 600-km discontinuities. Inversion of the hodochrones by the Wiechert-Herglotz method yields the depths of those discontinuities. This work was only possible for the two seismic regions 20° away from France: eastern Mediterranean Sea and western Turkey. A precise study of the hodochrones of 32 events (Plantet, 1972) leads to the following results: in the case of western Turkey, the depths of the 350 D and of the 600 D are found to be 388 km and 608 km; in the case of the eastern Mediterranean, the depths are 388 km and 627 km. The two new points obtained for the 600 D (when reported in the region of the turning points of the rays) agree with the map. However, this is subject to caution, since the precision on them is about

± 10 km. What can be stated is that an entirely different method yields results consistent with the general N-S dipping in that area.

The map on Fig. 3 was achieved during the summer of 1971. A further opportunity to check it came with the Cannikin test, detonated on Amchitka Island, Nov. 6, 1971. In its occasion, we installed an array of portable stations spreading N-S (i.e. in the direction of incidence) over 300 km (see Fig. 1). The results concerning the 600 D as obtained from this profile are less accurate, as we were not able to make a proper Moho correction. Nevertheless, they match the map within 25 km and they do confirm the dipping of the 600 D from N to S with a slope identical to the one previously obtained. Furthermore, a general study of the arrival time of the Cannikin test as a function of Δ over the

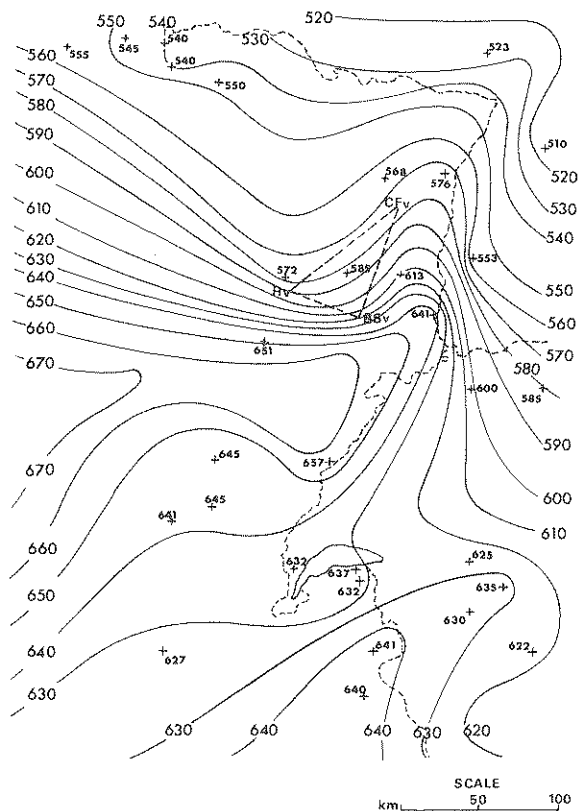


Fig. 4. The great 600 D anomaly under the Vosges.

whole of France yields a best-fit value of $p = dt/d\Delta = 5.22$ sec/degrees, which agrees with the standard tables. If, on the other hand, we restrain ourselves to the special N-S profile, we find a value of 5.46 sec/degrees. A more precise standard error calculation yields $5.18 < p < 5.27$ sec/degrees in the first case and $5.43 < p < 5.48$ in the second, with 90% expectation probability. This increase in p is consistent with the dipping of the 600 D (Burlot, 1973).

A further success of our map is its ability to account for the TCF effect as exposed in section 2: by a simple reading of the map, we are able to draw a scale profile along the 60° azimuth direction shown on Fig. 1. This profile is drawn on Fig. 5A. It is easy to put on this profile the points where the rays coming from EK and SK (and reaching our different seismic stations) cross the 600 D. On Fig. 5B, we plotted the corresponding relative amplitude of our records. It is easy to see the correlation of the two curves: curve B reflecting the curvature of curve A. The TCF effect is then explained

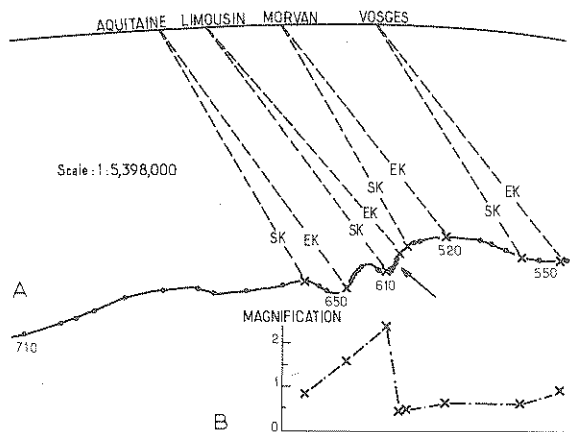


Fig. 5A. Profile along the direction of arrow No. 5 on Fig. 1. The dots indicate experimental points on the 600 D. The arrow shows the Vosges anomaly. B. Relative magnification as recorded for the various rays drawn on part A. The abscissae of the crosses are those of the 600 D crossing points of the rays.

by the focusing mechanism of the 600 D. We should notice that the drastic slope under the Vosges, pointed out in section 4, leads to very satisfactory results, as shown by the arrow on Fig. 5.

Further studies have been made with profiles of different azimuth; where it was not possible to find two regions of high magnitude events close apart in the same azimuth, we used amplitude records of both P- and P_c P-waves, in order to move the crossing point

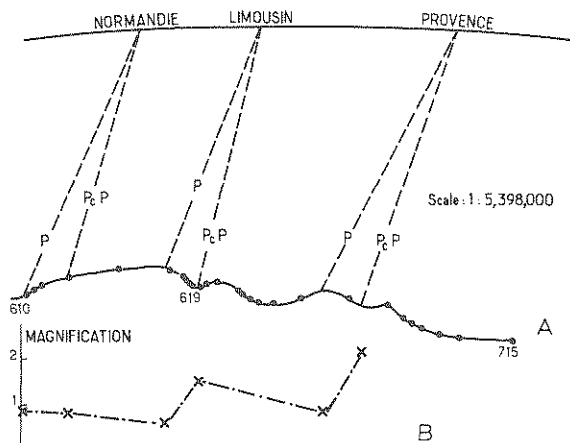


Fig. 6. A. Profile along the direction of arrow No. 6 on Fig. 1. The dots indicate experimental points on the 600 D. B. Relative magnification as recorded for the various rays drawn of part A. The abscissae of the crosses are those of the 600 D crossing points of the rays.

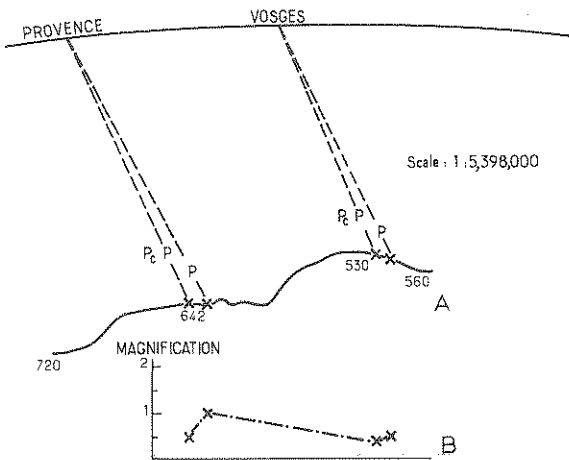


Fig. 7. A. Profile along the direction of arrow No. 7 on Fig. 1. The dots indicate experimental points on the 600 D. B. Relative magnification as recorded for the various rays drawn on part A. The abscissae of the crosses are those of the 600 D crossing points of the rays.

along the 600 D: the results are shown in Figs. 6 and 7, whose principle is the same as in Fig. 5. In this way, we are able to account for the high P_cP amplitude from Nevada test site as recorded in Limousin and Provence, and poor P_cP both in Provence and Vosges from Milrow test (Aleutian Islands). This explanation for P_cP/P ratios should be handled with care, since recent work (e.g. Ibrahim, 1971) has shown the extreme sensitivity of core-mantle reflection coefficients (and consequently of P_cP/P) to local conditions of the core-mantle boundary. However, especially in the case of Nevada records, this last interpretation of our observed anomalous P_cP amplitudes would lead to placing a region with several strong lateral inhomogeneities under Ungava Bay, northern Quebec, that is far away from the main features of the earth; there presently is no evidence for such an anomaly in the lower mantle at this location.

6. Possible alternatives to our model

Before discussing the implications of our model, we shall study whether any alternative explanations may be found. In section 2, we already discussed the depth of the laterally inhomogeneous region responsible for the TCF effect, and showed that it has to

lie in the middle mantle under France or more generally Europe.

As far as travel-time anomalies are concerned, it is well known that the inversion of laterally inhomogeneous data is not unique and that many structures may be responsible for a given set of data. In the case of our data, one such structure would be a laterally varying P-wave velocity, an example of which is derived by Brown (1972) from amplitude studies of P-wave arrivals. However, in Brown's paper, amplitudes are used to compute the direction of emergence of the wave through S, N, and E component ratios, which is an alternative way of computing a slowness. No absolute amplitude anomaly comparable to the TCF effect is studied and such a model would not account for it because it does not provide any focusing or defocusing mechanism.

In a similar way, a structure involving a thicker mantle layer could certainly be fitted to match our travel-time anomalies, but would be unable to explain the TCF effect. Finally, out of the numerous models which can explain our data on time residuals, the one proposed here is one of the very few which are able to give a proper explanation for a number of amplitude anomalies. Any other alternative would have to involve curvature of discontinuities (which indeed need not be the 600 D only), in order to give rise to very strong amplitude variations.

7. Discussion of our results

7.1. The dive under the Vosges

The strangest feature in our model is the strong dip of the 600 D under the Vosges region, as drawn on Fig. 4. The calculated average dip angle is about 43° , and this slope appears exceedingly large. However, we would like to emphasize the high ratio of SK/EK amplitudes at TCF, which, being about 7, needs a very strong anomaly as a system responsible for it. Furthermore, we must recall the good fit of both travel-time and amplitude data from our model in this particular area, as shown on Fig. 5.

Further study of the focusing mechanism is now under development: it might be possible that a combined action from both the 350 D and the 600 D would reduce the size of the necessary features within the mantle.

7.2. The main feature: NE-SW dip of the 600 D

The main feature of our model is undoubtedly the general dipping of the 600 D under France from 550 km in the northeast to 710 km in the southwest. As far as it provides good explanations of both the important travel-time and amplitude anomalies as recorded in France from teleseisms, this model is satisfactory. However, this feature is an important one and some of its implications must be considered.

One may first wonder whether such a feature can be "seen" by gravimetry. The total field created when moving the 600 D down for 100 km over the area of France is of the order of a hundred milligals. This gravimetric anomaly would be spread over France and its gradient would be very low. The present knowledge of the gravimetric anomalies over France cannot confirm such a feature. Furthermore, the local anomaly under the Vosges would be of much smaller amplitude and, as it is located deep in the mantle, would be spread over a great distance at the earth's surface; gravimetry would be no more sensitive to any such local anomaly. In any case, gravimetric anomalies would indeed be lessened by isostasy. The main discussion remains that of the variation of the depth of the 600 D. Variations in depth of the different layers of the crust and the mantle are not unknown: an example is the well-known effect of isostasy on the depth of the Moho. As another example, it is generally agreed that the more one goes east across the North American continent from the Rockies to the Great Plains, the thinner the LVZ and that in the same time, the 400 D dives from 350 km under the Rockies to 450 km under the eastern United States.

In our particular case, the slope of the 600 D is much greater. However, Ansorge and Mayer-Rosa (1968), in their extensive investigation of the mantle under Germany give a value of 550 km under the Ardennes, whereas Payo (1973) in a similar study of the Iberian peninsula, proposes 700 km under Spain. Therefore, if an agreement is to be found between these two authors, it can only be through data similar to ours.

The reason why the 600 D should undergo such a descent has to be discussed. Good agreement is generally found throughout the literature upon the nature of the 600 D, which is believed to be a phase transition of $(\text{Mg, Fe})_2\text{SiO}_4$ from β structure to Sr_2PbO_4 struc-

ture. Such a transition has been observed in materials with β structure but no pressure high enough has ever been developed in laboratory experiments in order to observe it directly in the above silicates (Ringwood, 1970).

When dealing with the upper mantle under Spain, one has to take into account the 1954 deep-focus earthquake which proved the existence of active material at these depths under Spain. Furthermore, there is recent evidence for accumulation of stress between Spain and North Africa (Mc Kenzie, 1970). Therefore, it is appealing to make the assumption that the African plate goes down under Spain, which explains both earthquake and tectonic stress. This assumption is, indeed, consistent with recent work by Papazachos (1973), showing the African plate to dive under the Aegean Sea. Even in this region of complicated tectonics, it seems reasonable to assume a uniform trend of the African plate with respect to the European continent.

The inclusion of a downgoing slab of lithosphere into the mantle under Spain could, by a diffusion process, act as to modify the parameters of the phase equilibrium at the 600 D. The first parameter which could be modified is, of course, temperature. The question whether the 600 D-transition is mostly temperature- or pressure-dependent (that is whether dp/dT associated with the transition should be larger or smaller than dp/dT associated with the geothermal gradient) is a very hard one since the latent heat is mostly unknown and even the geothermal gradient at those depths is not well known. The problem of the thermal regime of a downgoing slab has been extensively studied by Toksöz et al. (Toksöz et al., 1971, 1973; Minear and Toksöz, 1970). From these papers, it appears that the position and the trend of the isotherms are very much dependent upon a great number of parameters (such as radioactivity, rate of descent, thermal conductivities of both lithosphere and asthenosphere, etc.) of which none can be described as well known in the current problem. Even the general effect (cooling or heating) on the mantle can be inverted by a variation of those parameters (Minear and Toksöz, 1970; Toksöz et al., 1973). In those conditions we made a very rough calculation of the equation of heat for a cold slab under Spain, in which the isotherms turned out to be consistent with the general dipping of the 600 D. Of course, this is the result of a large number of best fits of the parameters (mostly age of the plate and thermal conduc-

tivities). It is also based upon the assumption of a temperature-dependent transition and on the approximation of a static problem.

It might, too, be possible that the diffusion process be not of entropy but of material, modifying over a huge number of years, the chemical composition of the asthenosphere in the region under Spain and France.

8. Conclusion

We have shown that our model which was built on travel-time anomalies, gives satisfactory results explaining the TCF effects and similar amplitude anomalies over the French network. It also provides a satisfactory link between apparently contradictory German and Spanish data. The interpretation is still rough. We tried to show that it might be explained by tectonic features in the area. The question is still open and further study of this subject will be directed in the following ways: a more accurate determination of more features of the discontinuity, a modelling of the TCF effect by ray theory, and an investigation of any possible effect created by the 350-km discontinuity.

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