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Travel-time residuals of PP waves reflected under the central Atlantic Ocean

C. Dorbath and L. Dorbath

O.R.S.T.O.M. BP 893, Bangui (Central African Republic)

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Travel-time residuals of PP, for waves generated in South and Central America and recorded in Africa and Europe after reflection under the central Atlantic Ocean, are studied to show possible structural variations in the reflection area. The ridge appears as a zone of positive residuals ($\simeq +1.0$ s) indicative of lower velocities, and is clearly differentiated from the surrounding oceanic basins where residuals are negative ($\simeq -3.2$ s); the ridge influence extends up to about 400 km from the crest.

1. Introduction

Several seismological works have studied crust and upper mantle structures underneath and around mid-oceanic ridges, particularly since they have been shown to be of major importance to our understanding of plate tectonics. Various methods have been used: reflection and refraction profiles (e.g. Le Pichon et al., 1965); surface wave dispersion along ridges (Knopoff et al., 1970; Weidner, 1974; Girardin and Jacoby, 1979); and travel-time delays of S waves generated by earthquakes located on ridges (Girardin and Poupinet, 1975; Duschesnes and Solomon, 1977).

Most of these studies give two-dimensional interpretations of the data, either perpendicular or parallel to the rift. Our goal is to look for variations in crust and upper-mantle structures both within the ridge and the surrounding oceanic basins utilising the travel-time residuals of teleseismic P waves reflected under the region. Reflected earthquake waves have been used to show lateral variation in Earth structures; for example PP waves for Atlantic Canada (Stewart, 1976; Girardin, 1980), P′P′ waves (Adams, 1968), multiple ScS (Okal and Anderson, 1975). For PP waves, travel-time anomalies can arise only from effects near the surface, whereas for waves reflected at the core, there can also be an effect from the core–mantle boundary.

The region of study is delimited by the Equator and the 35°N parallel, and the 30 and 60°W meridians. Reflection points of PP waves generated by numerous earthquakes occurring in Central America, the Caribbean loop and South America and recorded in Western Africa and Europe are located in this region. We could therefore rely on numerous records sufficient for an accurate statistical study.

2. Data

The travel-time residual for a P wave recorded at a distance Δ with respect to a particular travel-time table is:

$$R_{\mathrm{P}}(\Delta) = R_{\mathrm{s}}(\Delta) + R_{\mathrm{P,T}}(\Delta) + R_{\mathrm{st}}(\Delta) \qquad (1)$$

where $R_{\mathrm{s}}(\Delta)$ is the source anomaly for a distance Δ, $R_{\mathrm{P,T}}(\Delta)$ is the travel-time table error for P wave at distance Δ, and $R_{\mathrm{st}}(\Delta)$ is the station anomaly for a distance Δ.

The PP residual $R_{\mathrm{PP}}(\Delta)$, under the assumption that the travel-time anomaly R_{R} at the reflection point is identical for incident and reflected rays, is given by

$$R_{PP}(\Delta) = R_s(\Delta/2) + R_{PP,T}(\Delta) + R_{st}(\Delta/2) + 2R_R \quad (2)$$

where $R_{PP,T}(\Delta)$ is the travel-time table error for a PP wave at distance Δ. This formula is valid for a surface focus earthquake.

Neglecting for the moment time residuals peculiar to the focus and the recording station, and travel-time table error, the PP residual equals twice the residual of the P wave as would be recorded at the reflection point. Therefore a map showing PP residuals for a given reflection area will have the same characters as that for P residuals for the same area, though their values will be doubled.

All the earthquakes that we used are located in subduction zones. Jacob (1970) has shown, using a given subduction model, that variations of residuals with distance can appear for certain focal positions and azimuth. Within the frame of our statistical study, where sources and recording stations are spread over large areas, a possible bias due to this effect would not modify relative variations between the ridge and the surrounding basins.

The angle between the P and PP rays over the distance range sampled in this study is about 7.5° for $\Delta = 70°$ and 9.5° for $\Delta = 100°$ at both focus and station. It may therefore be assumed that the difference between PP and P residuals at the stations and at the source is negligible. This implies that

$$R_s(\Delta/2) = R_s(\Delta) \quad \text{and} \quad R_{st}(\Delta/2) = R_{st}(\Delta)$$

The main differences may occur due to variations in the error of the travel-time tables with distance. However, considering the limited distance range covered by our study it is reasonable to assume that $R_{PP,T}(\Delta) - R_{P,T}(\Delta)$ is constant and we shall neglect it. Then eq. 2 minus eq. 1 becomes

$$R_{pp}(\Delta) - R_p(\Delta) = 2R_R \quad (3)$$

The Jeffreys–Bullen Earth model was used to calculate the distance between epicentre and reflection point and its coordinates. For example, for a 85° epicentral distance, the distance between epicentre and reflection point decreases from 42.5° for a surface focus event to 39.9° for a 600-km deep focus. Although the coordinates of the PP point of reflection depend strongly upon depth of focus, the relationships (2) and (3) remain valid since the angles of incidence show only very slight variations.

This approach is valid when reflection points are located in a homogeneous medium, but it is to be considered a first order approximation for an area with lateral heterogeneities, such as near the central rifts.

Finally the PP residuals have been corrected for ocean bottom topography, by reducing the observations to the mean depth of the basins (5000 m). The corrective term is

$$-2h\,\frac{\cos i}{\alpha}$$

with h = level as compared to 5000 m, i = incidence angle, and α = local P velocity.

The depth of 5000 m (instead of the Earth's surface as commonly chosen) has been selected in order to reduce the influence of the corrective term, as fewer points need correction. The local P velocity chosen is 4.5 km s^{-1}, an average of velocities in layers 2A and 2B (Houtz and Ewing, 1976). 623 residuals coming from 30 stations located in Africa and Europe have been computed. The records from Mbour and the WWSSN stations of Porto, Toledo and Malaga were read from the seismograms (20% of the observations), and the remainder from ISC bulletins, selecting only phases identified as PP by ISC.

The frequency distributions of R_P, R_{PP} and $R_{PP} - R_P$ are shown in Fig. 1. R_{PP} and $R_{PP} - R_P$ distributions are much wider than R_p's. Possible explanations for these differences are as follows.

(1) PP can be misidentified.

(2) PP wave arrival is not impulsive, making its accurate reading more difficult. We estimate that the reading error in PP arrival time can reach 2 s or even more. Errors on $R_{PP} - R_p$ are of the same magnitude.

(3) Reflections may occur on a deeper reflector than the sea bottom including an early arrival of PP.

(4) Strong lateral heterogeneities in the reflection area may contribute to a wider residual distribution.

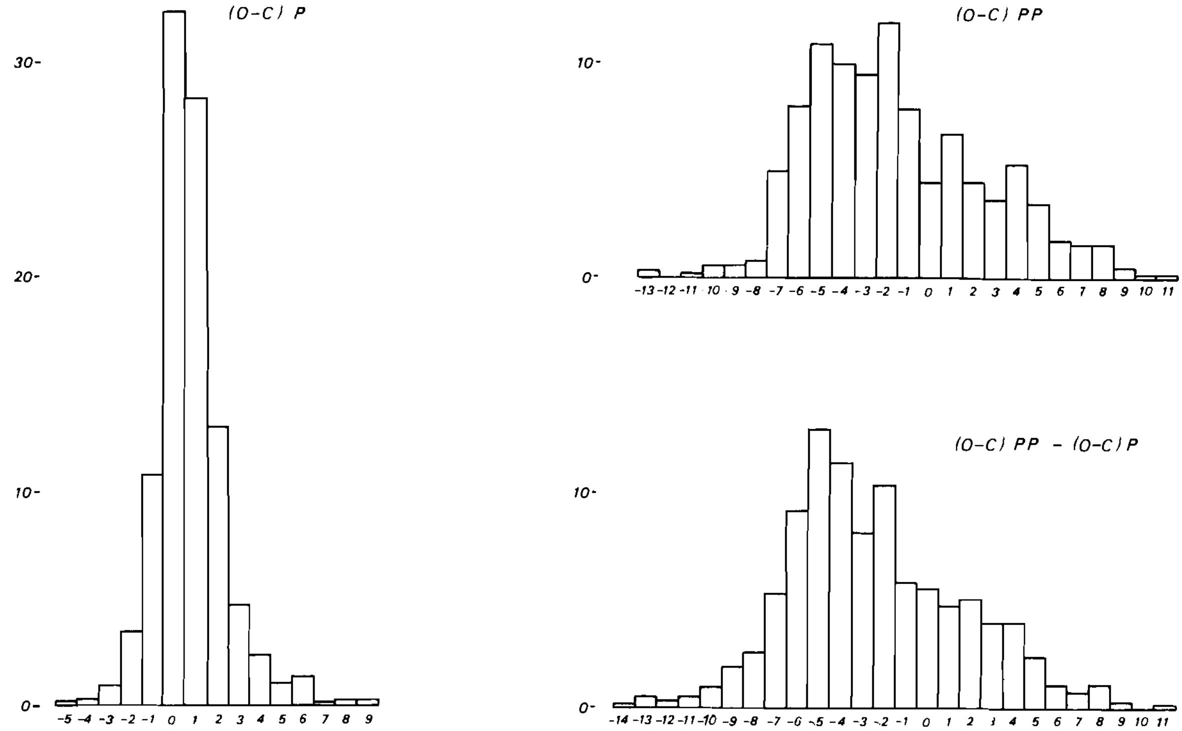

Fig. 1. Frequency distributions of R_P, R_{PP} and $R_{PP} - R_P$.

More detailed information on this wide distribution of PP residuals can be found in Stewart (1976). Furthermore, the R_{PP} and $R_{PP} - R_P$ distributions are not symmetric. This is due to the method used by ISC of accepting only those phases arriving within ± 7 s around the theoretical arrival time as PP. However we found from our readings a distribution centered around a value 2.5 s before the theoretical time.

3. Results

As the R_{PP} and $R_{PP} - R_P$ distributions are assymetric the calculated average values for a region are affected by the truncation at 7 s. Hence they must be considered as maximum estimates when they are very negative. Figure 2 shows $R_{PP} - R_P$ at their respectively reflection points together with the location of the rift as depicted by the epicentres of the major earthquakes recorded between 1969 and 1974 (NOAA data). The geographical locations of sources and receivers produce a very irregular geographic distribution of reflection points; there are more points on the eastern side (Cape Verde basin) while the northeastern part is poorly sampled. This, together with the uncertainty on each residual value, makes automatic mapping of iso-residual curves very difficult. Figure 3 has been obtained using Krigeage's method with the help of Stoclet of the "Centre de Calcul de l'INAG"; it shows iso-residual curves corresponding to 0 and -3.0 s. The following comments can be made.

(1) The surrounding basins, particularly the eastern basin for which we have more data, feature very low residuals, with an average value of -2.7 s.

(2) Zones with positive residuals, i.e. with lower velocities, are spread alongside the rift with a tendency to drift westward north of 20°N. This can be explained by the general topography of the area where the sea bottom rises steadily from basins to crest, with an elevation difference of 2 km over 1000 km distance. A realistic dip of

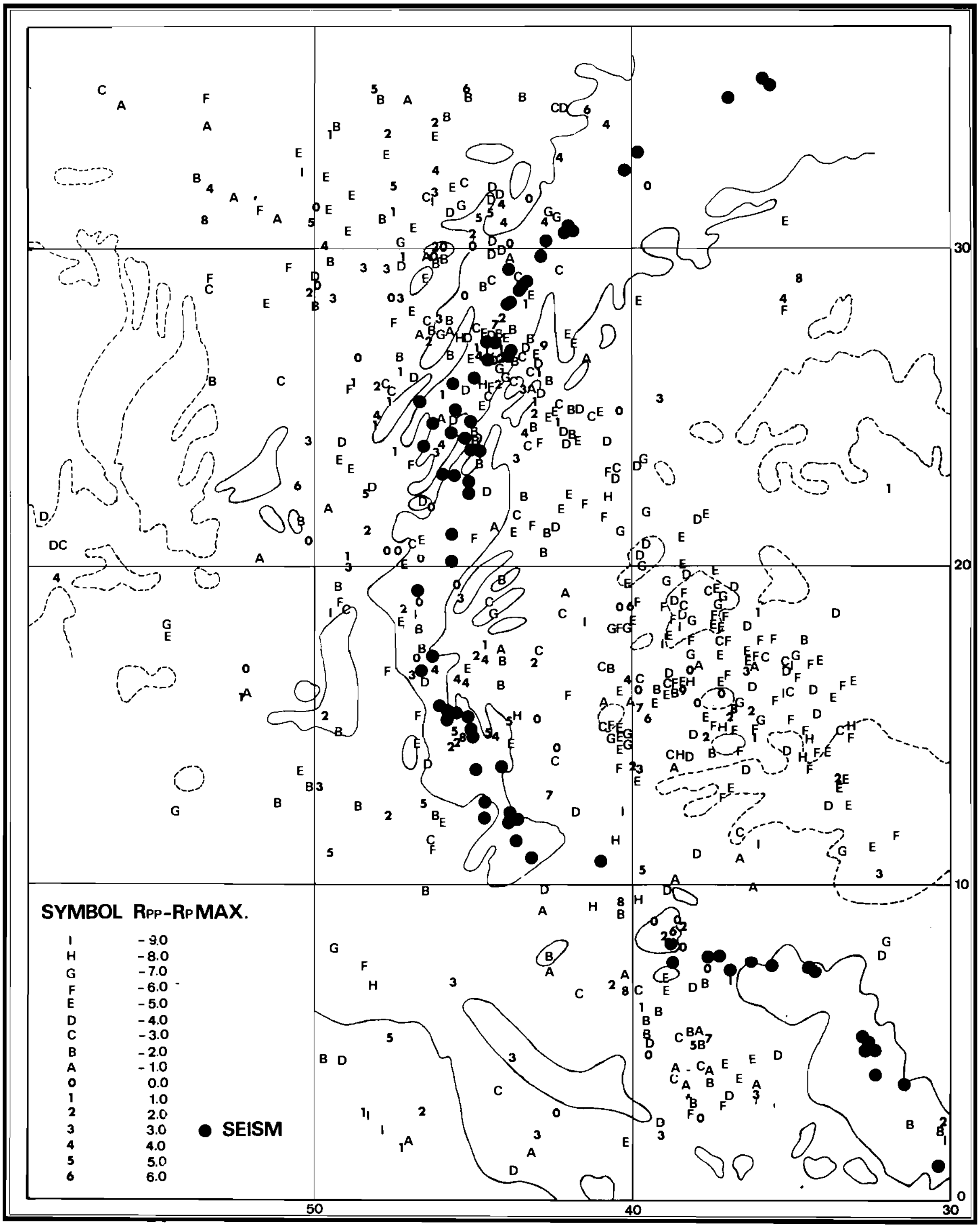

Fig. 2. $R_{PP} - R_P$ at their respective reflection points. The Mid-Atlantic Ridge is depicted by the epicenters of major earthquakes.

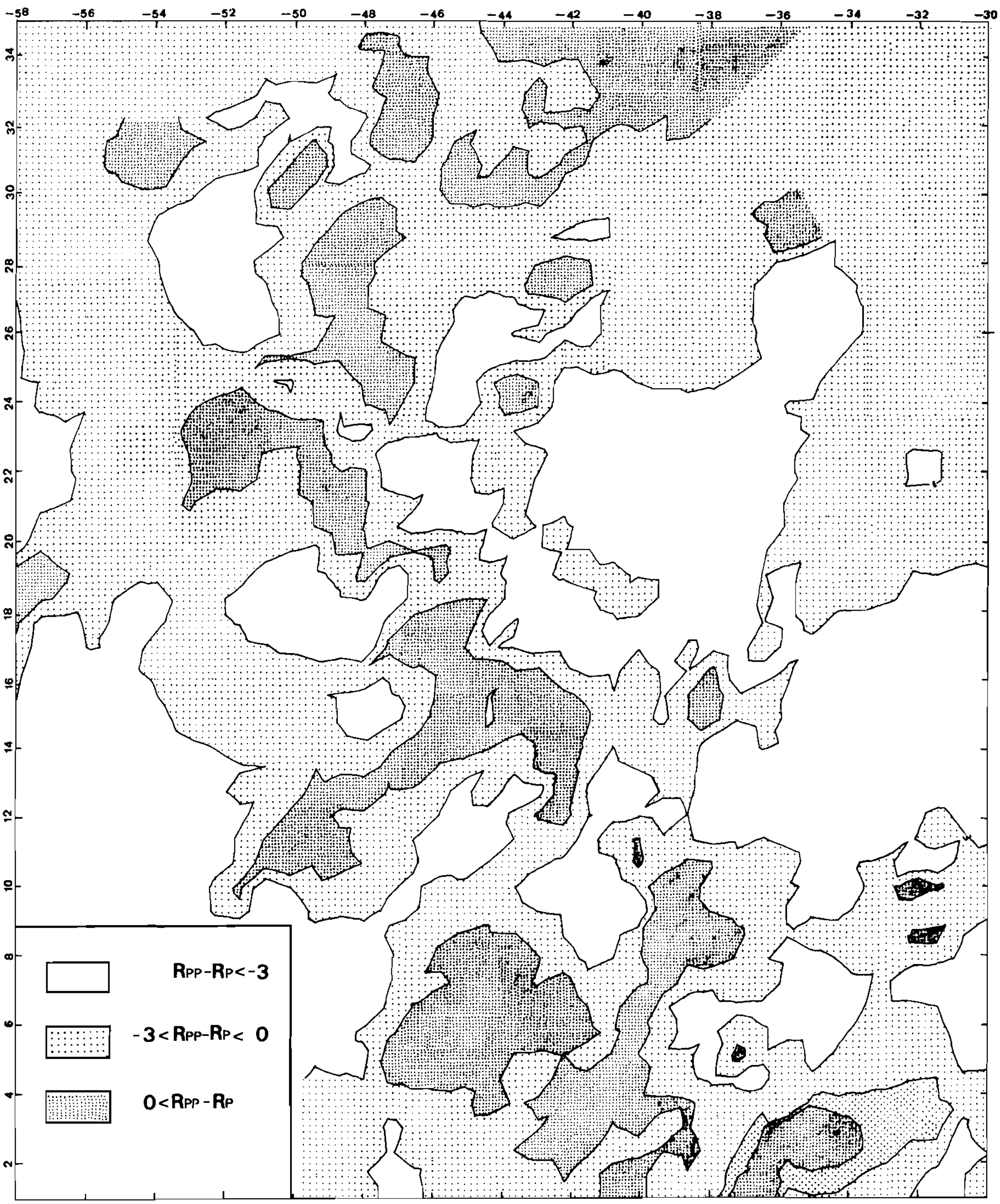

Fig. 3. Results of automatic mapping of iso-residual curves of $R_{PP} - R_P$ using Krigeage's method.

0.11° westward leads to a 0.5° westward displacement of the reflection point and vice-versa. Positive zones west of the rift on Fig. 3 should therefore be shifted towards it, improving correlation. The effect of the ridge on the residuals between 0° and 8°N is not evident due to a lack of data points.

(3) Zones of positive residuals exist outside the ridge, smaller and secluded on the eastern side, larger and linked with the ridge on the western side.

4. Discussion

Most of the area studied is covered by less than 0.1 km of sedimentary cover, according to Ewing et al. (1973). Thick sedimentary deposits (> 1 km) appear only off the South American coast. The two positive residual areas centred on 11°N, 49°W and 6°N, 45°W, on the axis of the Amazon delta, may be explained by the increased seimentary thickness. Smaller positive zones east of the ridge are thought to be a facet of the mapping method and to be of no geological significance.

A previous work on Rayleigh-wave dispersion across the Cape Verde and Canaries basins (Dorbath, 1976) has shown their structure to be similar to Saito and Takeuchi's (1966) model 8.1.0. This model's computed PP residual value of -3.4 s is in good agreement with the observed mean value of -2.7 s which is, as previously stated, a maximum estimate. A comparison of these values implies that reflection actually takes place at the sea bottom, rather than at a deeper interface, this

TABLE I
Average computed P residuals

Distance from the crest (km)	Equivalent P residuals (s)	Age (My)
0	$+1.04 \pm 0.70$	0
100	$+0.88 \pm 0.76$	5
200	-0.12 ± 0.62	12
300	-0.97 ± 0.74	19
400	-1.60 ± 0.49	25
500	-1.60 ± 0.39	30
600	-1.42 ± 0.42	35
700	-1.55 ± 0.38	43
800	-1.70 ± 0.38	53
900	-1.62 ± 0.46	58
1000	-1.70 ± 0.34	63

* After Pitman et al. (1974).

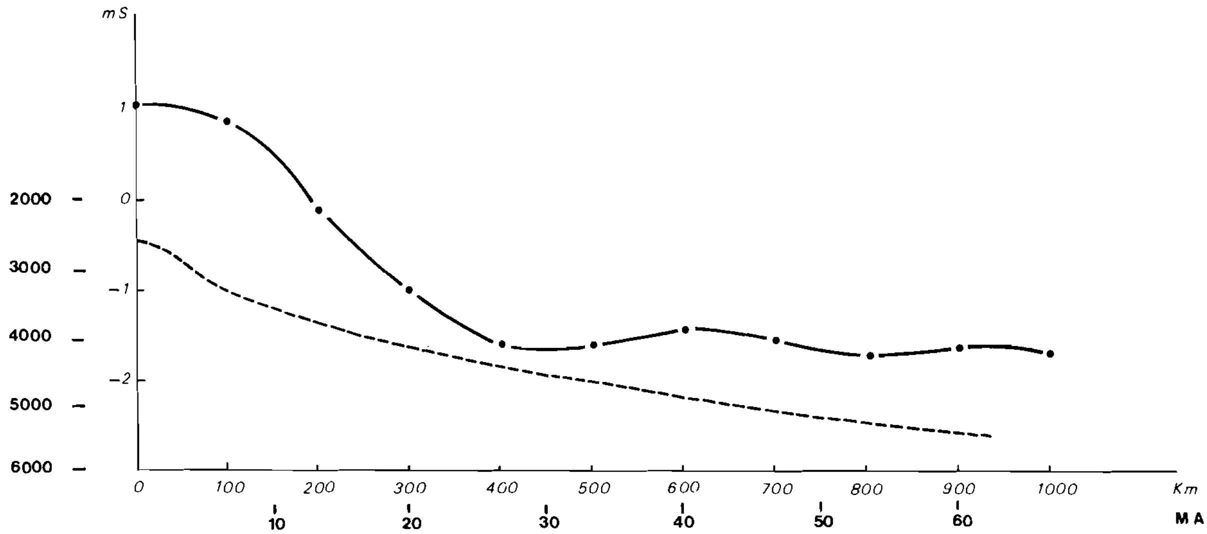

Fig. 4. Average equivalent P residuals versus the distance from the crest and the age of the crust (after Pitman, et al., 1974). Dashed line: generalised water depth.

being true "a fortiori" near the rift where residuals are higher.

No published ridge model fits exactly with our observations. The upper 600 km of the rift model of Knopoff et al. (1970), for example, determined from Rayleigh wave dispersion, gives faster values than those actually found.

There are a few stations actually located on the ridges, e.g. AKU, REY, SID, PDA and JMI which have an average residual for P of 1.3 s, i.e. 2.6 s for PP (Poupinet, 1977), which, corrected for 5000 m below sea level gives a positive residual of at least 0.8 s, in agreement with our observation. Furthermore, Rowlett and Forsyth (1979) observe P residuals of 1.3 to 1.8 s on records made by 4 ocean bottom seismographs located on the Atlantic rift at 11°N. This corresponds to 0.6–1.6 s PP residuals after correction for depth.

The values have been taken from the area to the east of the ridge between 12°N and 20°N, where we have a high density of good observations. Furthermore, to complete the picture, there is no major east–west fracture zone in this area. We have computed average P residuals (deduced from the above PP's) at 100 km intervals from the crest by averaging a 200 km band centered on each point. The results are shown in Table I and Fig. 4. The residual on the ridge is positive 1.0 ± 0.70 s and decreases quickly from the crest up to a distance of about 400 km, then stabilizes to a value of around -1.6 s.

Duschesnes and Solomon (1977) have observed a 6 s variation of S-wave residuals between the rift and a 100 My old oceanic lithosphere. If we take a value of 1.81 for the V_P/V_S ratio (Poupinet, 1977) in the oldest parts of the basins we obtain a value of 3.10 under the rift, using the relative residual variations of Duschesnes and Solomon for S waves and our values for P waves. This value seems to be a realistic one for these regions of partial melting. Our results are therefore compatible with oceanic thermal models such as "wet gabbro–eclogite" and especially "wet periodite" (Duschesnes and Solomon, 1977). The model of Bottinga and Steinmetz (1979) explains about half of the P residual variation we observe, but it must be noticed that this model extends only to a depth of 140 km.

distinct zones: between 0 and 400 km from the crest where the observations can be explained by an abnormally slow mantle and after 400 km where normal oceanic basin values are found.

5. Conclusion

Travel-time residuals of PP waves reflected under the central Atlantic Ocean, generated by Central and South American earthquakes, are more widely scattered than the corresponding P waves. We have found a good correlation in the oceanic basins between measured average values and those computed with a basin model. This demonstrates that most of the waves are reflected at the sea bottom. Large negative values can be explained by reflection on deep discontinuities, and large positive values by misidentification (pPP for example). A statistical processing of PP residuals may therefore be used to show lateral variation of structures.

The mid-Atlantic ridge is clearly outlined on the map by the highest residuals, i.e. the lowest velocities. Other high residual zones can be noticed, particularly in the southwestern part of the area of interest. They can be explained by thick sedimentary deposits.

A P average residual of the order of -1.6 s for the basins fits correctly with existing models. The influence of an abnormal mantle under the rift is effective up to a distance of 400 km from the crest. To explain the delay time beneath the rift relative to the basins, the upper mantle needs a velocity as low as 7.3 km s^{-1} extending from the crust down to 200 km (Rowlett and Forsyth, 1979); this very low velocity and the value of about 3.1 that we find for the V_P/V_S ratio indicate that a zone of partial melting extends beneath the mid-Atlantic ridge.

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