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EVIDENCES OF THE EXISTENCE OF AN ABNORMAL SEISMIC SIGNAL ATTENUATION IN SOUTHERN MEXICO

J. YAMAMOTO*

RESUMEN

El análisis de temblores ocurridos en la costa de Chiapas, México, indica que existe una fuerte atenuación de las señales sísmicas (ondas P) en un cierto rango de distancias y azimuts, esto se observa especialmente en las estaciones sismológicas localizadas en el este de Norteamérica. En el presente trabajo se discuten tres posibles explicaciones del origen de este fenómeno, basadas en nuestras observaciones y otros estudios.

ABSTRACT

The analysis of earthquakes occurred off the coast of Chiapas, Mexico, indicates the existence of an abnormal high seismic signal (P-waves) attenuation in a certain range of distances and azimuths. This feature is observed primarily in seismic stations located east of North America. In the present paper three possible explanations of this phenomenon origin are discussed based on our own observations and other studies.

INTRODUCTION

As it was pointed out by Yamamoto (1978) one interesting feature common to earthquakes occurred off the coast of Chiapas, Mexico (Figure 1) is that the P-wave amplitudes observed at a certain range of distances and azimuths are appreciably smaller than amplitudes predicted by theory. This fact is specially noticeable at seismic stations located eastern United States. As a result of this situation, calculations of some seismic parameters like magnitude or seismic moment made with records of eastern Unites States stations may not be directly compatible with other stations estimates.

* *Instituto de Geofísica, UNAM, MEXICO.*

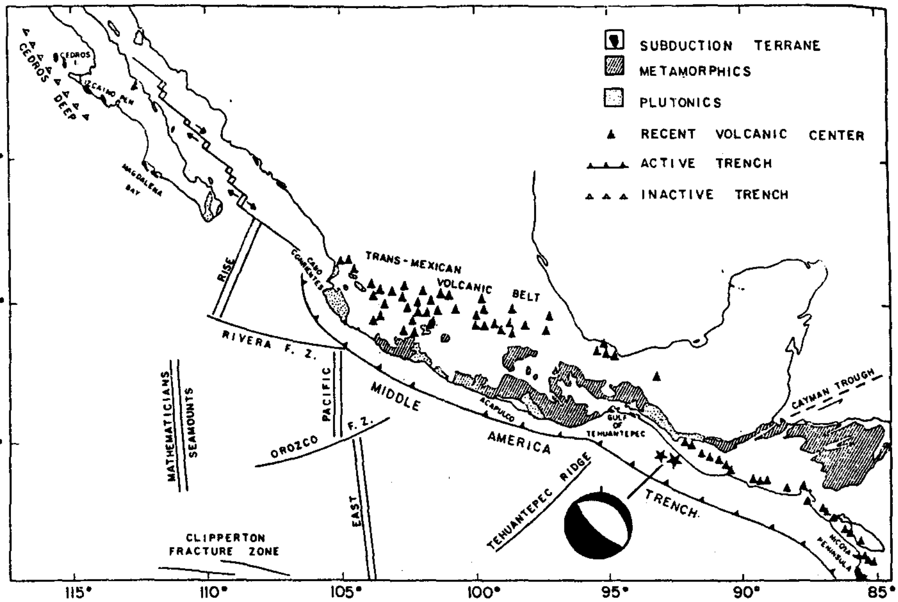


Fig. 1. Map of Mexico and Central America showing plate boundaries and other geological features. Earthquakes selected for study denoted by (★). Epicentral coordinates are given in Table 1. Focal mechanism diagram is also shown. The darkened quadrants are the quadrants of dilatation, first motion. (Base map taken from Karig *et al.* (1978).

The observed low P-wave amplitudes could be produced by a variety of mechanisms such as some sort of assymetry in the radiation of the source or by some anomaly in the anelastic attenuating characteristics of the medium through which waves travel. Thus, in the present paper a detailed analysis of the data using waveform modelling in the time domain is made in an attempt to reach a plausible explanation to this phenomenon.

DATA

The largest foreshock (OT = 11:22, M_s = 6.3), main shock (OT = 14:01, M_s = 7.3) and the most prominent aftershock (OT = 08:33, M_s = 6.4) associated with an earthquake sequence started on April 29, 1970 off the coast of Chiapas, Mexico (14.5^{o}N, 92.8^{o}W) are used in the discussion. Complementary information regarding these earthquakes is given in Table 1.

Table 1

Events studied. P,T and B are the focal mechanism parameters

Date	Origin Time	Lat N	Long W	Depth (km)	M_s	P		T		B	
						Trend	Plunge	Trend	Plunge	Trend	Plunge
29 Apr. 70	11:22:35.9	14.47°	92.72°	45	6.3	213°	27°	53°	62°	308	9
29 Apr. 70	14:01:33.9	14.45°	92.71°	44	7.3	213	27	55	61	308	10
30 Apr. 70	08:32:57.5	14.58°	93.16°	16pP	6.4	215	27	49	63	309	6

Events studied P, T and B are the focal mechanism parameters and represent the maximum Pressure, Tension and Null vectors respectively.

A set of WWSSN long- and short-period seismograms from about 70 stations were collected. Figures 2 and 3 show some typical examples of long-period P-wave records of the foreshock, and aftershock. The long-period P waveform of the aftershock displays a very simple character. The foreshock on the other hand, appears to be far more complicated. It shows among other things, an extra secondary arrival superimposed on the first down swing. This secondary onset (suppossedly due to a multiple source; see Yamamoto, 1978) disappears for a narrow interval of distances and azimuths. The main shock shows even more complicated P-waveforms due to precursory activity.

INTERPRETATION OF P-WAVEFORMS

Most of the results of this paper were obtained by computing time domain synthetic seismograms and comparing these directly to observed records. The fitting process was done manually, by trial and error. Theoretical P waveforms were computed in the 30 to 80 degree distance range using the Haskell-Thomson method as described in Yamamoto (1978). The computer program includes source and receiver crust response, anelastic attenuation, and instrument response (Herrmann, 1978).

Figure 2 shows the observed long-period vertical component P-wave and the corresponding synthesized signals for the aftershock. The synthetics were computed assuming a point source at 15 km depth and as a source function a trapezoid defined with three time segments (2.5, 1.0, 2.5 seconds). The seismic moment (M_0) was assumed to be 10^{26} dyne-cm. The low-angle dipping nodal plane (strike = 292°, dip = 19°, and slip = 72°) obtained from P-wave first motion data was taken as the fault plane. A ratio T/Q = 1 was assumed.

From the same figure, it can be seen that the match between the shape of the observed and computed signals is remarkably good from the first arrival to approximately 23 seconds into the record. Stations located in the western United States show the best fit. P-wave signals observed at stations in the eastern United States show longer periods and smaller amplitudes than those predicted by the synthetics. European stations show a similar effect but to a lesser degree. It seems that a ratio T/Q of the order of 3 or 4 should be a better choice to match the synthetics to the observed seismograms at eastern United States stations.

The long-period P-wave signals associated with the foreshock exhibit special features. Stations located in the western United States and some in Europe show a

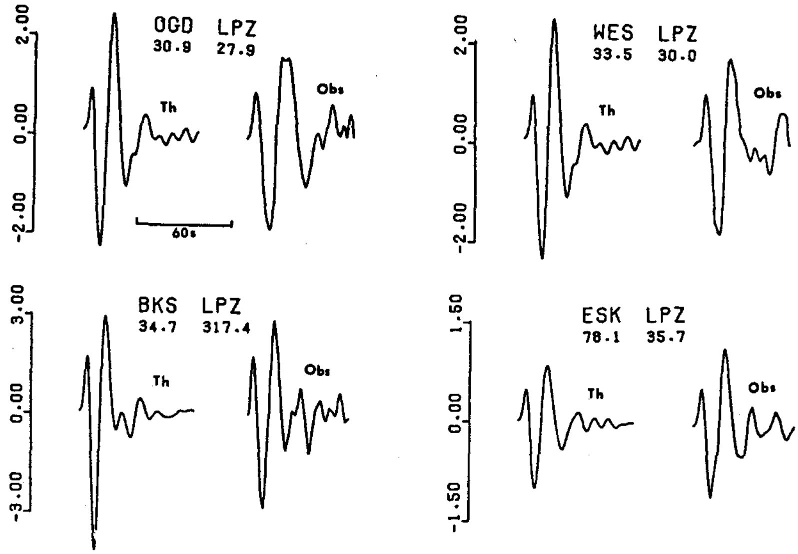


Fig. 2. Synthetic and observed seismograms of the aftershock. The numbers below station and component identifier are the distance and azimuth in degrees, respectively. Vertical scale is in cm. A ratio T/Q (= T*) equal to 1 was assumed for all stations.

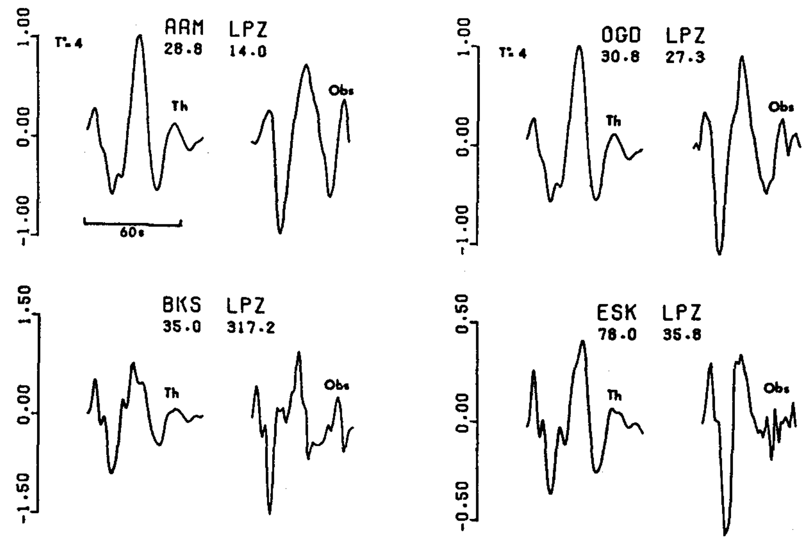


Fig. 3. Synthetic and observed seismograms of the foreshock. The numbers below station and component identifier are the distance and azimuth, respectively. Vertical scale is in cm. For same stations a ratio T/Q (= T*) equal to 4 was used. Two sources at a depth of 50 km were used.

clear secondary onset 7 seconds after the first arrival (see BKS in Fig. 3). Thus, to model this earthquake a source time function composed of two trapezoids was used to simulate two consecutive events. The sources were set at a depth of 50 km.

In Figure 3 the observed and synthesized signals are shown. As a direct consequence of the form assigned to the source time function, an extra second arrival approximately 7 seconds after the first P is clearly shown by the synthetics. As mentioned before, however, stations in the eastern United States do not show the predicted secondary arrival (see AAM). As before, one way to reconcile the observations with the theory is by assuming a ratio T/Q of the order of 3 or 4 in the computation of the synthetics at eastern United States stations. There are also other ways to reduce eastern United States amplitudes, but this manner seems to be the easiest. It can be shown that an increase in the value of T/Q produces an increment in the apparent period of the signal at the expense of its amplitude, a similar result can be obtained by increasing the source time duration. Observations, for example, at BKS and WES (Fig. 7), which are practically at the same distance from the source seem to show precisely that behavior.

The main shock is even more complicated. Unfortunately it is not very suitable for P-waveform modelling since most stations went off scale.

DISCUSSION

Figure 4 is a plot of the observed P-wave amplitudes on the long-period vertical component versus the amplitude measuread on the corresponding synthetic seismogram. Amplitudes were measured from the zero line to the peak of the first up-swing, since this is the best modelled part of the record. From this figure it is clear that the P-wave amplitudes observed at eastern United States stations (AAM, GEO, SCP, WES, and OGD) depart considerably from the expected amplitudes computed assuming a double couple point source and a ratio T/Q = 1. This value of the ratio T/Q is an average value that adequately simulates the non-elastic attenuation along most ray-paths (Langston and Helmberger, 1975). In particular WES and OGD are the most anomalous followed by AAM, GEO and SCP. Some european stations (PTO and ESK) and GDH in Greenland also show slightly lower amplitudes.

The same pattern is seen in the case of the foreshock, as shown in Figure 5. The figure shows that there is a tendency for the observations at eastern United States stations (in particular WES and OGD) to deviate from the general trend of the data. Some European stations show the same tendency, but it is less pronounced.

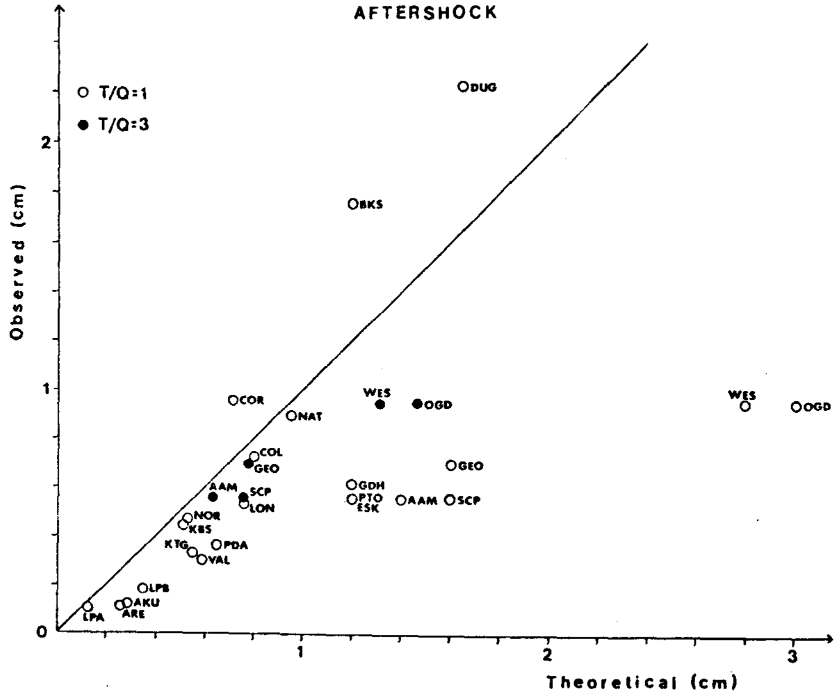


Fig. 4. Comparison of observed and theoretical amplitudes for the aftershock.

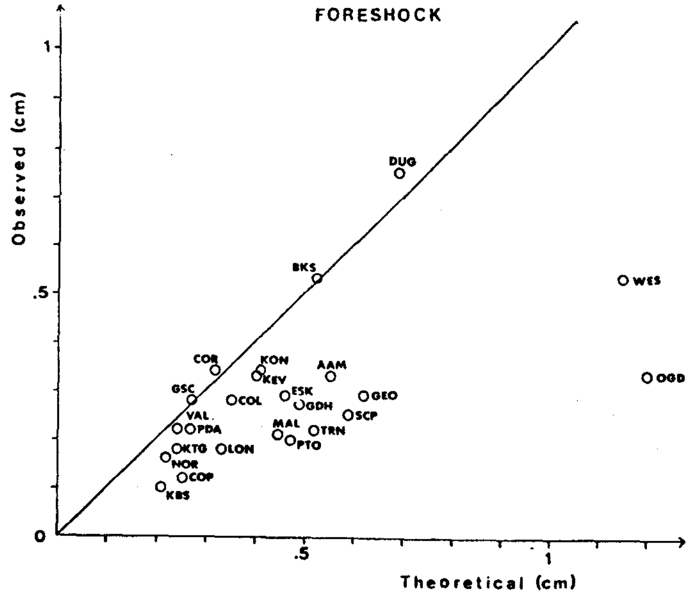


Fig. 5. Comparison of observed and theoretical amplitudes for the foreshock.

In addition to the attenuated P-wave amplitudes, the P waveforms themselves have been altered. It was noted above that most of the modelled stations are more or less consistent with the value T/Q equal to one except for the eastern United States stations (see WES and OGD in Figs. 2 and 4) for which T/Q equal to 3 or 4 is necessary to match synthetic and observed signals. In a similar way the anomalous observations shown in Figures 3 and 5 can be reconciled with the theory by assuming T/Q values equal to 3 or 4 in the calculations of the synthetics. This is shown in Figure 4 with solid circles.

Eastern United States stations recording anomalously attenuated signals are located in a very narrow range of azimuth (~14° - 27°) and epicentral distance (~28° - 33°) as shown in Figure 6. Some other stations in eastern United States but at

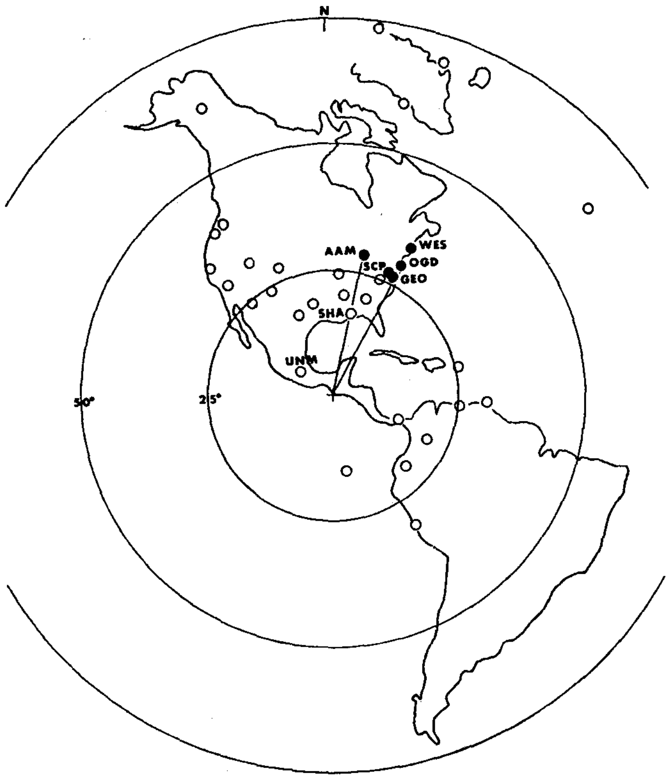


Fig. 6. Map showing the stations at which very attenuated P-wave signals are observed. Polar projection centered at epicentral region.

shorter distances (e.g. SHA, $\Delta \sim 17^{\circ}$) also show signs of abnormal attenuation, judging from the fact that no secondary arrival is seen on the records of the foreshock. Local stations (e.g. UNM, $\Delta \sim 7^{\circ}$), on the other hand, show the secondary arrival. This restricted range suggests that the cause of the fluctuations in amplitude is very limited in horizontal dimensions and depth.

Thus, the analysis of the P-wave amplitudes and waveforms indicates that the signals reaching eastern United States stations have been considerably attenuated during their journey. This phenomenon is responsible for the disappearance for a certain range and distances and azimuths of the secondary arrival shown by the foreshock. The attenuation, regardless of its nature, seems to affect the entire spectrum of the signal, but it is more severe for high frequencies, causing the secondary onset to be completely wiped out. The resultant signal, therefore, has smaller amplitude, longer period and a smoother appearance.

Of course, it is also possible that the observed low P-wave amplitudes and the disappearance of the secondary arrival are not a product of the attenuation in the med-

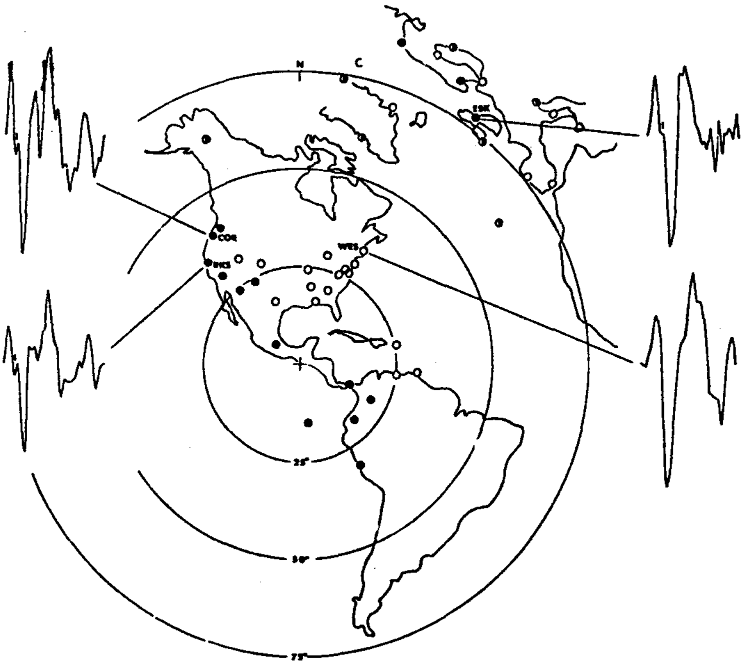


Fig. 7. Illustration of attenuated multiple at station WES.

ium, but instead, due to some kind of asymmetry in the radiation of the source itself. However, this possibility is not very likely, since some European Stations (e.g. ESK) show the secondary arrival and in addition their P-wave amplitudes agree reasonably well with the theory. Note that WES and ESK lie practically along the same great circle path (Fig. 7).

Kanamori and Stewart (1976) observed that the amplitudes of the surface waves generated by the Guatemala earthquake of 1976 (located 400 km northeast of our epicentral region) are clearly too small in the azimuth range of 0^{o} to 90^{o}. They concluded that this is produced because the rupture propagated toward the west. In our case, this type of explanation is not applicable either, since the three earthquakes analysed show a similar amplitude distribution pattern, and according to Yamamoto (1978) the rupture associated with the main earthquake propagated toward the northeast and then toward the south. In addition, the aftershock was found to be too simple an event to expect that the observed amplitude abnormalities are produced by a rupture propagation mechanism.

It therefore seems that the observed low amplitudes at eastern United States stations are not caused by source energy directionality neither complexities in the rupture process. The question is then, where along the ray-path the attenuation is taking place and what is its mechanism. The attenuating region, if it exists, might be located beneath the recording stations, at the source region, or in between. The attenuation may also be averaged along the entire raypath. With our present data we cannot give a final answer to these questions. We can however, explore some of the more likely possibilities.

Seismologists generally agree, based mainly upon studies by Evernden and Clark (1970), Solomon and Toksöz (1970), Mitchell (1973), Nuttli (1973) and Der *et al.* (1975), that the eastern United States is characterized by a lower attenuation of seismic waves than is the western United States. Among this general pattern however, a localized region of high attenuation in northeastern United States has been reported by Solomon and Toksöz (1970) result that is consistent with our present observations. In contrast, Buttler and Ruff (1980) did not found a systematic pattern of amplitude variation across United States stations. Thus, receiver effects cannot be ruled out without additional studies as an explanation for the low amplitudes observed at eastern United States stations in the present study.

Molnar and Oliver (1969) in a comprehensive study of Sn wave transmission in Mexico concluded that the upper mantle beneath the Gulf of Mexico transmits Sn waves very efficiently. Paths crossing volcanic chains show inefficient transmission of Sn waves and conversely, efficient transmission is associated with paths where volcanoes are absent, although some clear exceptions were found. The attenuation of Sn waves along the studied paths may be produced by the Mexican volcanic belt or by a highly attenuating structure beneath the plateau. The last possibility is supported by the results of Steinhart and Meyer (1961) who reported that the crustal thickness (~45 km) is not great enough to explain the observed large Bouguer gravity anomaly.

Near the source, the P-wave signals recorded at eastern United States and European stations travel along similar but not identical paths. At a distance of 200 km from the source (measured along the ray-path), the separation between rays roughly speaking, is of about 45 km and at 300 km about 70 km. Therefore, the possibility of a small attenuating zone at the source region (Fig. 8) is not at all unlikely.

Theoretically, the average Q along the entire ray-path to station WES is about 400 (assuming T/Q = 1). If the observed small amplitudes are produced by anelastic attenuation then a value T/Q = 4 is more realistic for that path. Dividing the total ray-

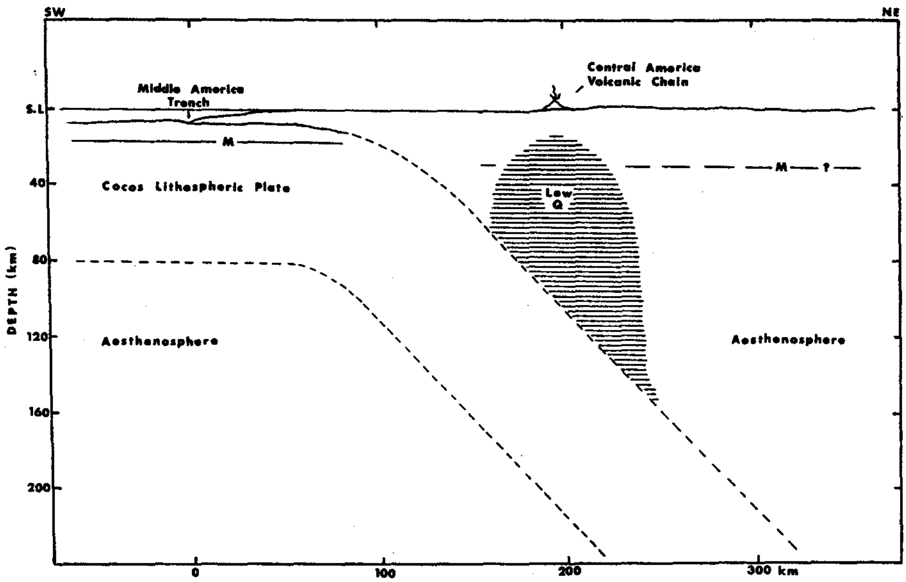
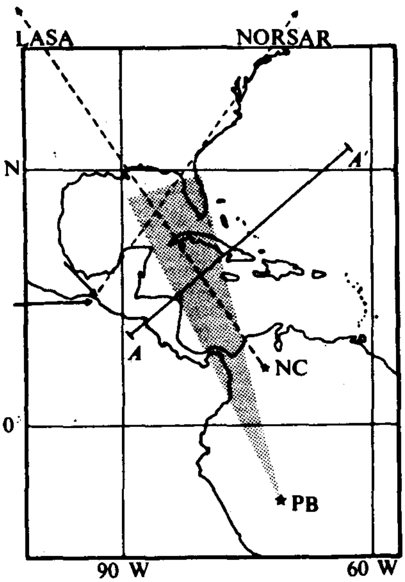


Fig. 8. Schematic diagram of the structural model used as an alternative to interpret our data. Boundaries of the hypothetical anomalously low-Q zone are highly speculative. Similar models widely documented have been proposed for other subduction zones (see Grow and Qamar, 1973).

path in two segments (T_1, Q_1 and T_2, Q_2) and assuming that most of the anomalous attenuation takes place along the segment T_1, then the difference in Q's between the anomalous (Q_1*) and the "normal" (Q_1) is $1/Q_1^* - 1/Q_1 = 3/ET_1$, where E is the fraction of the total ray-path along which the anomalous attenuation takes place. Thus, the translation from the difference in T/Q values to absolute difference in Q's is not straightforward, since we have to know the size of the anomalous region and the effective Q of the "normal" medium. For example, if T_1 = 40 seconds (10% of the total ray-path) and $Q_1 = 150$, then $Q_1^* \sim 2$. That is, the anomalous region would have an effective Q unreasonably low. It is doubtful that such a low Q value can occur for P waves. Although this result is very suggestive it cannot be taken as definitive to reject the possibility of the existence of a localized high attenuating region along the ray-path. Since, due to the trade off between anelastic attenuation and source time function, time-domain modeling alone may lead to unrealistic Q-estimates (Der and McElfresh, 1980).



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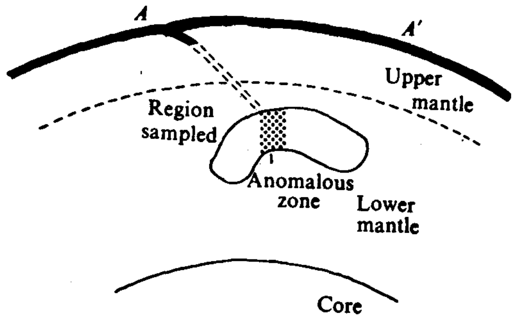

9b

Fig. 9a. Surface projection (shaded area) of the mantle traversed by rays with large positive residuals. Stars labeled Nc and PB represent epicenters of northern Columbian and Peru-Brazil events respectively used by Jordan (1975).

Fig. 9b. Vertical section along the line AA' (see Fig. 9a) showing the anomalous zone which lies along the projection of the Benioff zone beneath the Middle America trench as proposed by Jordan (1975).

It is also possible that the variations of amplitude have their origin at a certain distance from the source. Jordan and Lynn (1974), in a study of ScS-S and PcP-P differential travel times and absolute S-wave travel time from two deep South American earthquakes, concluded that the scatter observed in the travel-time data is produced by a region of anomalously high velocity in the lower mantle beneath the Caribbean (Fig. 9b). According to them, the velocity anomaly extends from a depth of 600 m to 1 400 km. Jordan and Lynn (1974) speculate that the observed velocity anomaly beneath the Caribbean may be caused by the presence of a piece of slab detached from the lithosphere plate subducting at the Middle America trench (Fig. 9b).

Sheppard (1973) studied the differences in epicentral locations made using the NORSAR array (Norway) with the locations made with the global network of seismic stations. He concludes that the large differences in location observed for earthquakes in southern Mexico are due to a velocity anomaly near the source region and that the anomaly probably extends very deep into the mantle. Davies and Capon (1972) reported similar results using information from LASA array.

It is interesting to notice how the distribution of P-wave amplitudes described earlier in this section shows the same anomalous pattern as the data presented by the above authors, even though the observations are of a different nature. Figs. 6 and 9a indicate that we are sampling practically the same part of the mantle. This correspondence is unlikely to be fortuitous.

Defocussing of rays propagating through a downgoing slab is another, and perhaps the most viable explanation to the observations. It is known that slabs of lithosphere which descend into the mantle are characterized by higher seismic velocities (typically 7% higher, Engdahl, 1973) and lower attenuation than the surrounding mantle. Thus, the paths of rays emerging from a source near the sinking lithospheric slab are expected to be strongly distorted (e.g. bent away from the axis of the slab) producing extremely low amplitudes in some regions and enhancing the signals in others.

Davies and Julian (1972), using a ray tracing technique and assuming a realistic velocity model for the descending slab, corroborated the above effect and were able to explain the anomalous short-period P-wave low amplitudes recorded in western Canada and Europe from the Longshot nuclear explosion fired in the Aleutians. They noticed that the effect of a geometrical shadow on the shape of the signal is similar to the effect which would be produced by an anelastic mechanism.

In a more elaborated paper, Ward and Aki (1975) studied the effect of a sinking lithospheric slab on short-period body waves. Their discussion is based on synthetic-seismograms and spectra at teleseismic distances from sources located near or in the slab. They found that energy is focused in a certain direction and defocused in another and that waves propagating down the slab show appreciably larger waveform broadening and smaller amplitudes than waves propagating through the slab. Ward and Aki also showed that the presence of the descending slab and faulting dynamics can have a similar effect on the long-period P-waveforms.

A further problem remains. The above explanation can be easily applied to a region in which the descending slab is very well developed (Aleutians). In southern Mexico however, seismic evidence indicates that the downgoing slab is not clearly defined. Even more, it may have a transitional character.

CONCLUSION

Three possible explanations of the origin of the strong attenuation of P-wave signals observed at stations in the eastern United States were discussed. The first alternative is that the low amplitudes result from anelastic attenuation caused by an anomalous region located near the source. Estimates using conservative values for those parameters suggest that the effective Q for the anomalous region would be many times lower than the Q of the surrounding medium and would probably be unreasonably low. The second alternative is that the low amplitudes result from some sort of defocusing effect produced by refractions in an isolated heterogeneous region beneath the Caribbean. The third, and perhaps the simplest alternative is that the observed pattern of amplitudes is produced by rays propagating in the presence of a downgoing slab (Cocos lithospheric plate).

Additional studies are necessary to decide among the three possibilities. This is important in connection with elucidating the mechanics of lithospheric plate subducting at the Middle America trench. Besides, regardless of the anomaly nature it should be taken into account when computing some seismic parameters. Since, magnitude or seismic moment estimations using eastern United States stations could be underestimated up to a factor of 2.

BIBLIOGRAPHY

BUTLER, R. and L. RUFF, 1980. 'Teleseismic short period amplitudes: source and receiver variations, *Bull. Seism. Soc. Am., 70,* 831-850.

DAVIES, D. and B. R. JULIAN, 1972. A study of short period P-wave signals from longshot, *Geophys. J. R. Astr. Soc., 29,* 185-202.

DAVIES, D. and J. CAPON, 1972. *Semiannual Technical Summary, Seismic Discrimination, 39, MIT, Lincoln Laboratory,* December.

DER, Z. A., R. P. MASSE and J. P. GURSKI, 1975. Regional attenuation of short-period P and S waves in the United States, *Geophys. J. R. Astr. Soc., 40,* 85-106.

DER, Z. A. and T. W. McELFRESH, 1980. Time-domain methods, the values of t_p* in the short-period band and regional variation of the same across the United States, *Bull. Seism. Soc. Am., 70,* 921-924.

ENGDAHL, E. R., 1973. Relocation of intermediate depth earthquakes in the Central Aleutians by seismic ray tracing, *Nature Phys. Sci., 245,* 23-25.

EVERNDEN, J. F. and D. M. CLARK, 1970. Study of teleseismic P. II. Amplitude data, *Phys. Earth. Planet. Int., 4,* 24-31.

GROW, J. A. and A. QAMAR, 1973. Seismic-wave attenuation beneath the Central Aleutian Arc, *Bull. Seism. Soc. Am., 63,* 2155-2166.

HERRMANN, R. B. (Editor), 1978. Computer programs in earthquake seismology, Vol. 1: General Programs, *Saint Louis Univ., Pub. No. 240,* pag. X-1.

JORDAN, T. H. and W. S. LYNN, 1974. A velocity anomaly in the lower mantle, *J. Geophys. Res., 79,* 2679-2685.

JORDAN, T. H., 1975. Lateral heterogeneity and mantle dynamics, *Nature, 257,* 745-750.

KANAMORI, H. and G. S. STEWART, 1976. Seismological aspects of the Guatemala earthquake of February 4, 1976, *Calif. Inst. Technol., Contribution* 2901.

KARIG, D. E., R. K. CARDWELL, G. F. MOORE and D. G. MOORE, 1978. Late Cenozoic subduction and continental margin truncation along the Northern Middle America Trench, *Bull. Geol. Soc. Am., 89,* 265-276.

LANGSTON, C. A. and D. V. HELMBERGER, 1975. A procedure for modelling shallow dislocation sources, *Geophys. J. R. Astr. Soc., 42,* 117-130.

MITCHELL, B. J., 1973. Surface wave attenuation and crustal anelasticity in Central North America, *Bull. Seism. Soc. Am., 63,* 1057-1071.

MOLNAR, P. and J. OLIVER, 1969. Lateral variations of attenuation in the upper mantle and discontinuities in the lithosphere, *J. Geophys. Res., 74,* 2648-2682.

NUTTLI, O. W., 1973. Seismic wave attenuation and magnitude relations for Eastern North America, *J. Geophys. Res., 78,* 876-885.

SHEPPARD, R. M., 1973. *Semiannual Technical Summary, Seismic Discrimination, 12, MIT, Lincoln Laboratory,* June.

SOLOMON, S. C. and M. N. TOKSOZ, 1970. Lateral variation of attenuation of P and S waves beneath the United States, *Bull. Seism. Am. 60*, 819-838.

STEINHART, J. S. and R. R. MEYER, 1961. Explosion studies of continental structure, *Carnegie Inst. Wash. Publ.*, 622.

WARD, R. W. and K. AKI, 1975. Synthesis of Teleseismic P-waves from sources near sinking lithospheric slabs, *Bull. Seism. Soc. Am., 65,* 1667-1680.

YAMAMOTO, J., 1978. Rupture processes of some complex earthquakes in Southern Mexico, *Ph. D. Dissertation, Saint Louis Univ.*, 203 pp.

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