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Crustal structure south of the Mexican volcanic belt, based on group velocity dispersion

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RESUMEN

Se estimó la estructura de velocidades de la corteza con trayectorias entre las costas de Guerrero y Michoacán, a lo largo de la zona de subducción y la ciudad de México. Esta estructura se obtuvo de la inversión de datos de dispersión de la velocidad de grupo para registros individuales. La estructura de la corteza media (de 5 a 30 km de profundidad) está bien definida, mientras que la estructura superficial y profunda no puede resolverse bien debido al rango limitado de períodos en los datos de dispersión. Una frontera entre la alta y la baja corteza se encuentra bien definida entre 15 y 20 km. Hay una buena comparación entre los sismogramas sintéticos generados con el modelo y los observados, dando confianza a los resultados y al posible uso de este modelo en la determinación rutinaria de las soluciones del tensor de momento sísmico. La gruesa capa superficial de baja velocidad encontrada en la región de Oaxaca no está presente en la región que aquí se estudia.

PALABRAS CLAVE: Estructura cortical, corteza Mexicana, dispersión de ondas superficiales.

ABSTRACT

An average shear-wave velocity structure of the crust has been estimated for paths between the Guerrero-Michoacán coast, along the Mexican subduction zone, and Mexico City. This structure is obtained from inversion of group velocity dispersion data, estimated from broadband seismograms of regional events recorded in the city. For a better definition of the dispersion curve we have used a new stacking technique of the period-group velocity amplitude distributions of individual records. The mid-crustal structure (5 to 30 km depth) is well defined but both the shallow and the deep structures are not well resolved due to limited period range of the dispersion data. A clear boundary between the upper and lower crust is found between 15 and 20 km. Synthetic seismograms computed for the model compare well with the observed ones, giving us confidence in the results and in the possible use of this model in routine determination of moment tensor solutions. The thick low-velocity superficial layer found in the Oaxaca region is not present beneath the region studied here.

KEY WORDS: Crustal structure, Mexican crust, surface-wave dispersion.

INTRODUCTION

A detailed knowledge of the velocity structure of the crust and the upper mantle in Mexico is essential (a) in understanding the tectonic evolution, (b) in estimating ground motion from future earthquakes, and (c) in locating earthquakes and estimating their focal parameters. Since the pioneering study of Meyer *et al.*(1961) on crustal structure in the Central Mexican Plateau, seismic waves from both controlled sources (e.g., Valdés *et al.*, 1986; Nava *et al.*, 1988; Gomberg *et al.*, 1988; GEOLIMEX Working Group, 1993) and earthquakes (e.g., Fix, 1975; Lomnitz, 1982; Gomberg *et al.*, 1988; Gomberg and Masters, 1988; Suárez *et al.*, 1992) have been used to infer crustal and upper mantle structure in different regions of Mexico. Although there is an urgent need to know both the P- and the S-wave crustal structure between the southern Pacific coast of Mexico and Mexico City, it is poorly known at present. The need arises not only from scientific interest in the geologic structure between the provinces north and south of the Mexican Volcanic Belt (MVB) but also because the earthquakes which cause damage in Mexico City originate along this part of the coast. A detailed crustal model is a basic requirement to understand the nature of the incoming wavefield to Mexico City, which is essential for the prediction of future ground motion, and for moment tensor inversion. For these reasons, in this paper we develop a flat-layer crustal model between the coast of Guerrero-Michoacán and Mexico City based on the inversion of group velocity dispersion measurements. This dispersion has been measured on broadband recordings at UNM, a station in Mexico City, from nine well located moderate coastal events (Figure 1). We compare this model with others for Mexico and discuss its implication in the interpretation of the waveforms recorded in Mexico City during the great 1985 Michoacán earthquake ($M_w = 8$).

DATA

Since April, 1991 a broadband GEOSCOPE station has been operating on the Universidad Nacional Autónoma de Mexico (UNAM) campus. The characteristics of GEOSCOPE stations are given by Romanowicz *et al.* (1991). This station records almost all moderate and large earthquakes which occur in Mexico. Unfortunately, locations and origin times are poorly known for most of these events. We selected those events which could be well located, having been recorded locally on the accelerograph and/or the seismic array, presently in operation along the coast of Guerrero and Michoacán (Anderson *et al.*, 1994; Suárez *et al.*, 1986). Figure 1 shows the epicentral locations of these events (listed in Table 1). As an example of the data, Figure 2 presents seismograms of event 9 (Table 1) recorded at UNM.

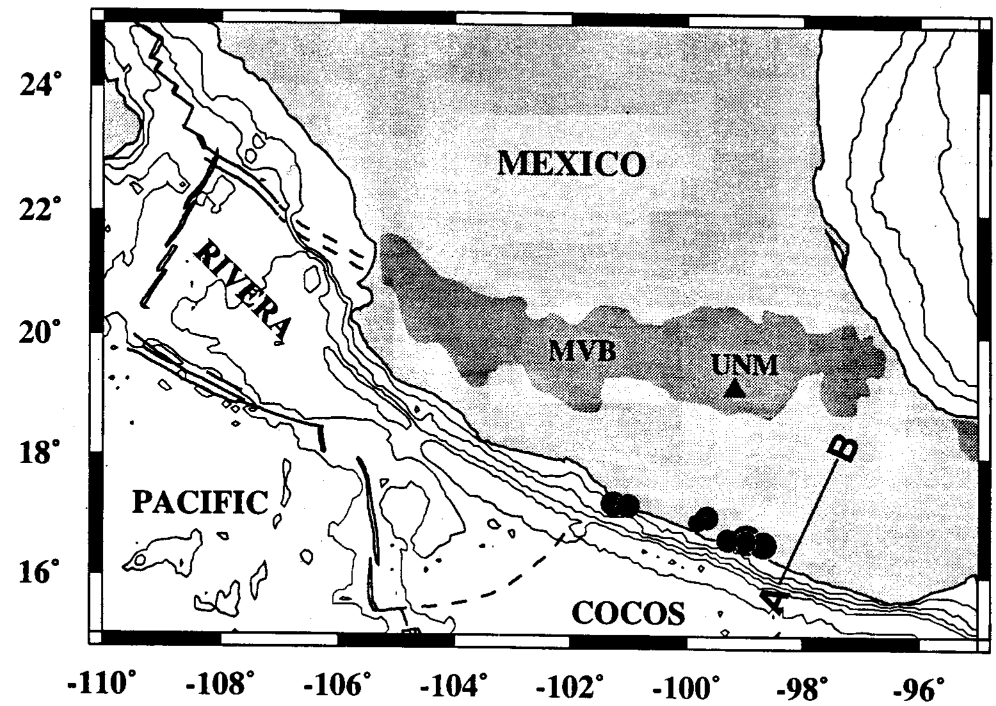

Fig. 1. Location map of the region of interest. The earthquakes are shown by solid circles. Seismograms analyzed are from UNM, a broadband station located in the Mexican Volcanic Belt (MVB). Line AB denotes Oaxaca refraction profile. Velocity structure of Central Mexican Plateau which lies to north of MVB (not shown in the figure), is known from refraction and phase velocity measurements.

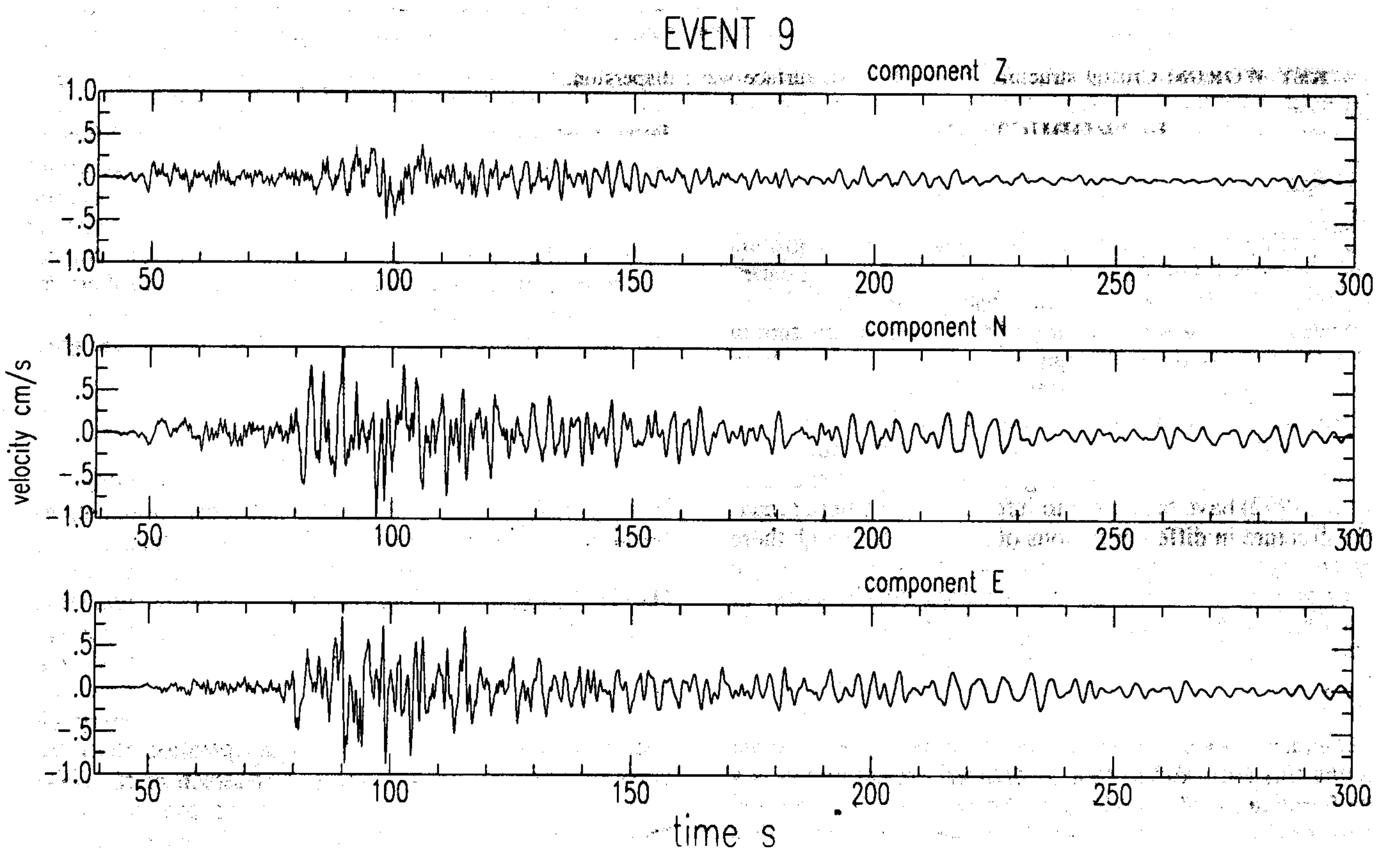

Fig. 2. Broadband seismograms of event 9 (Table 1).

Table 1

Earthquakes used in this study

Event No	Date YMD	Distance to UNM	Latitude (°N)	Longitude (°W)	Depth (km)	M
1	910421	301.4	16.61	98.98	16.0	4.2
2	910528	275.1	16.92	99.82	27.2	3.6
3	920109	262.3	17.00	99.65	30.2	4.7
4	920331	321.6	17.22	101.27	11.0	5.1
5	921224	299.8	16.62	99.29	18.4	4.8
6	930331	254.6	17.19	101.01	6.0	4.8
7	930515	312.7	16.55	98.68	15.6	5.6
8	930515	312.7	16.55	98.68	15.6	5.9
9	931024	299.3	16.63	98.97	34.6	6.5

ANALYSIS

The group velocity dispersion of the fundamental mode of the Rayleigh wave was computed using a frequency-time analysis (see e.g., Dziewonski *et al*, 1969; Herrmann, 1973; Keilis-Borok, 1989). The frequency-time analysis of one individual record consists of the following steps: (a) computation of the Fourier transform of the input signal, (b) multiplication of the complex spectra by a Gaussian filter:

$$H(\omega)=e^{-((\omega-\omega_0)/a\omega_0)^2} \tag{1}$$

where ω_0 is a central frequency of the filter and a is the relative bandwidth ($a = 0.5$ was used), and (c) computation of the inverse Fourier transform of the filtered spectra. This results in a frequency-time dependent function $S(\omega_0,t)$. For a single mode the amplitude of this function at a fixed frequency, $A_s(\omega_0,t)$, is, approximately, a Gaussian function of time with the maximum at group time $\tau(\omega_0)$. It is more convenient to use the period-group velocity representation which is obtained through a simple coordinate transformation:

$$T = 1/\omega_0 \tag{2}$$

$$u = r/t \tag{3}$$

where T is a period, u is the group velocity at that period, and r is the event-station distance. The dispersion of group time $\tau(T)$ is related to the dispersion of the group velocity $U(T)$ through the relation:

$$U(T) = r/\tau(T)\ . \tag{4}$$

The isoline map of the function $A_s(T,U)$ in the period-group velocity plane gives a convenient graphical representation of the signal. The location of the maximum of amplitude at each period helps to define the dispersion curve. In principle, it is possible to separate several modes if the ridges associated with each mode in the frequency-time representation of amplitude are sufficiently separated. For events occurring along the Mexican subduction zone and recorded at UNM, the small epicentral distance may result in large variations of the measured velocities for the different events. These variations are caused by the local changes of the geological structure, and the uncertainties in location and origin times. The heterogeneities of the crust result in diffraction effects (multipathing, reflection, etc.) that make the identification of modes difficult. This deterioration in the measurement is especially important at short periods where higher modes also contribute significantly in the same group velocity window. Figure 3 shows an example of the amplitude distribution in the period-group velocity plane of the vertical component for event 9. For the reasons mentioned above and seen in this figure, the dispersion curve is difficult to define with confidence. In order to improve the definition of the dispersion curve, we experimented with a stacking procedure which accumulates the information provided by all the available events and to provide a mean dispersion curve for the region of interest. To accomplish this, we use logarithmic stacking in the period-group velocity domain. As input information for stacking we have n individual period-group velocity dependent amplitude functions corresponding to n different events with different epicental distances r_i. The mean period-group velocity diagram, $A_s(T,u)$, is the product of the individual ones:

$$A_s(T,u)=N_1(T,u)\bullet N_2(T,u)\bullet \ \mathrm{K}\ \bullet N_n(T,u) \tag{5}$$

where $N_i(T,u)$ is the normalized amplitude diagram for event i. Let $U_i(T)$, $\tau_i(T)$ be the dispersions of group velocity and group time for event i, respectively. At a given period T, an individual envelope for a single mode is approximately a Gaussian function of time:

$$N_i(T,t)=e^{-[(a/T)(t-\tau_i(T))]^2}\ . \tag{6}$$

Using equations (2), (3), and (4) in the vicinity of the dispersion curve (U_i~u), we obtain:

$$N_i(T,u)=e^{-[(ar_i/(TU_i^2(T)))(u-U_i(T))]^2}\ . \tag{7}$$

Let us assume that the epicentral distances are the same for all the events (it is approximately true for the events considered). If the dispersion $U(T)$ is the same for all records then the resulting envelope has a unit maximum value and a bandwidth equal to $U^2(T)T/(ar_in)$. For different dispersions $U_i(T)$, the maximum amplitude of the resulting envelope is smaller than 1. Let us consider the dispersions, $U_i(T)$, distributed around a mean value $U_o(T)$:

$$U_i(T)=U_o(T)+\delta_i(T) \tag{8}$$

where δ_i (T) is the group velocity deviation for each record, caused by both local structure heterogeneity and uncertainties in the event location. In this case the mean envelope

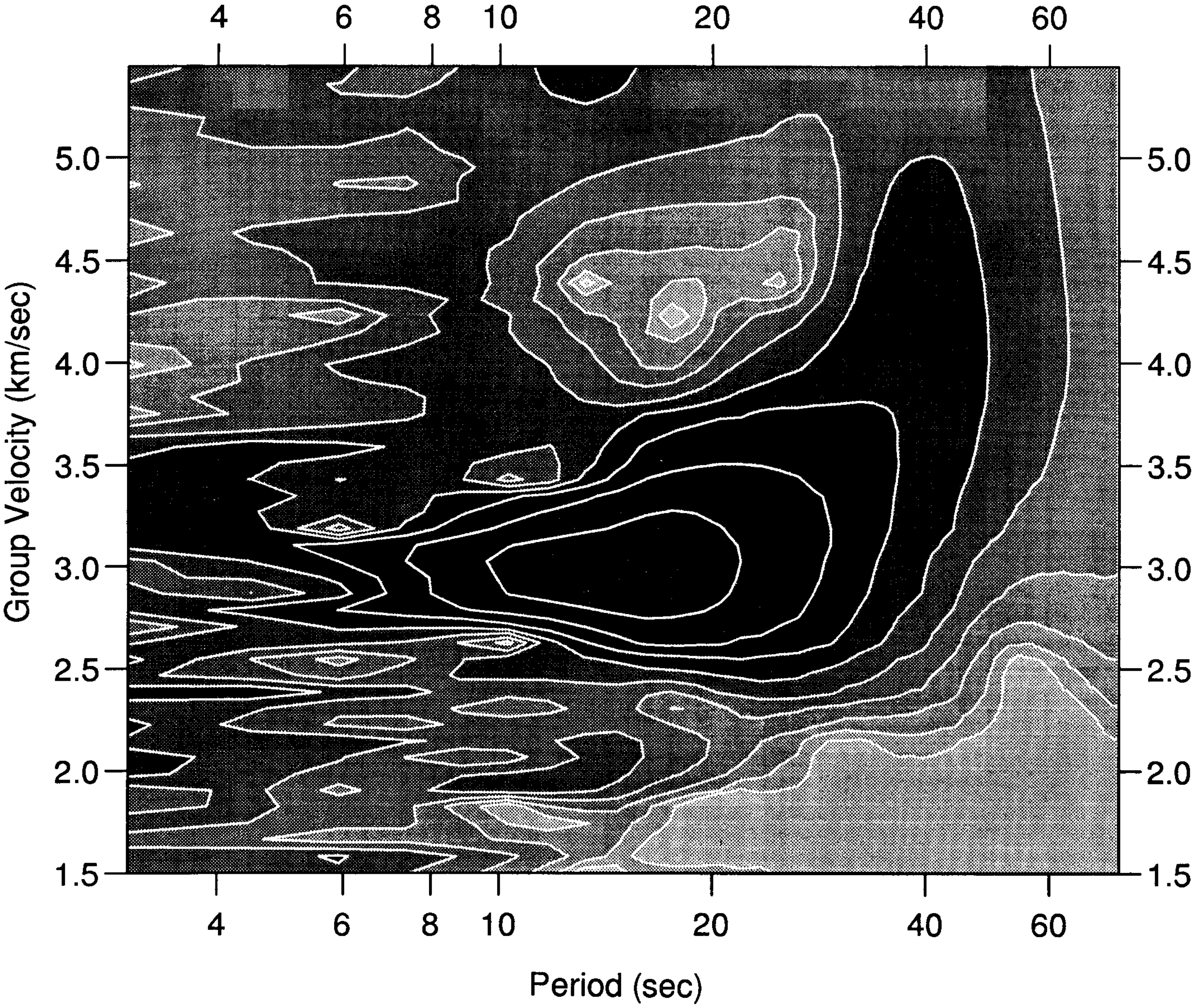

Fig. 3. Period-group velocity diagram for the vertical record of event 9. The isolines correspond to the values: 0.01, 0.02, 0.03, 0.05, 0.09, 0.15, 0.25, 0.42, 0.71.

will have a maximum at velocity $U_o(T)$ with the amplitude equal to

$$Max(A_s(T,u)) = e^{-\left[(ar/(TU_0^2(T))\right]^2 \sum_i \delta_i^2(T)} \tag{9}$$

The amplitude of the mean envelope at a given period depends on the variance of group velocities and its width is proportional to the inverse of the number of records *n*. Therefore, the resulting period-time envelope has a strong amplitude in the narrow region where we have arrivals with similar dispersion. The maximum value of this envelope can be used to evaluate the variance of group velocity using equation (9).

The result of period-group velocity stacking of vertical components of all 9 events is shown in Figure 4. It corresponds to the fundamental mode of the Rayleigh wave and has significant amplitudes in the period range of 6.5 to 50 sec. Figure 5 shows the final estimation of the group velocity dispersion. The shaded area has the half-width of the standard deviation calculated with equation (9).

CRUSTAL STRUCTURE

The measured group velocity curves were inverted to infer the vertical distribution of shear-wave velocity. The inversion consisted of two steps. First, we performed the gradient inversion of the average dispersion curve using a set of programs by Herrmann (1987). The starting model used in this inversion was the one proposed by Campillo *et al.* (1989), which is based on results of a refraction study in Oaxaca (Valdés *et al.*, 1986). In the second step, we estimate the uncertainty of the model using a linearized inversion. Towards this goal, we generated a set of random models. To be consistent with the results of Valdés *et al.*

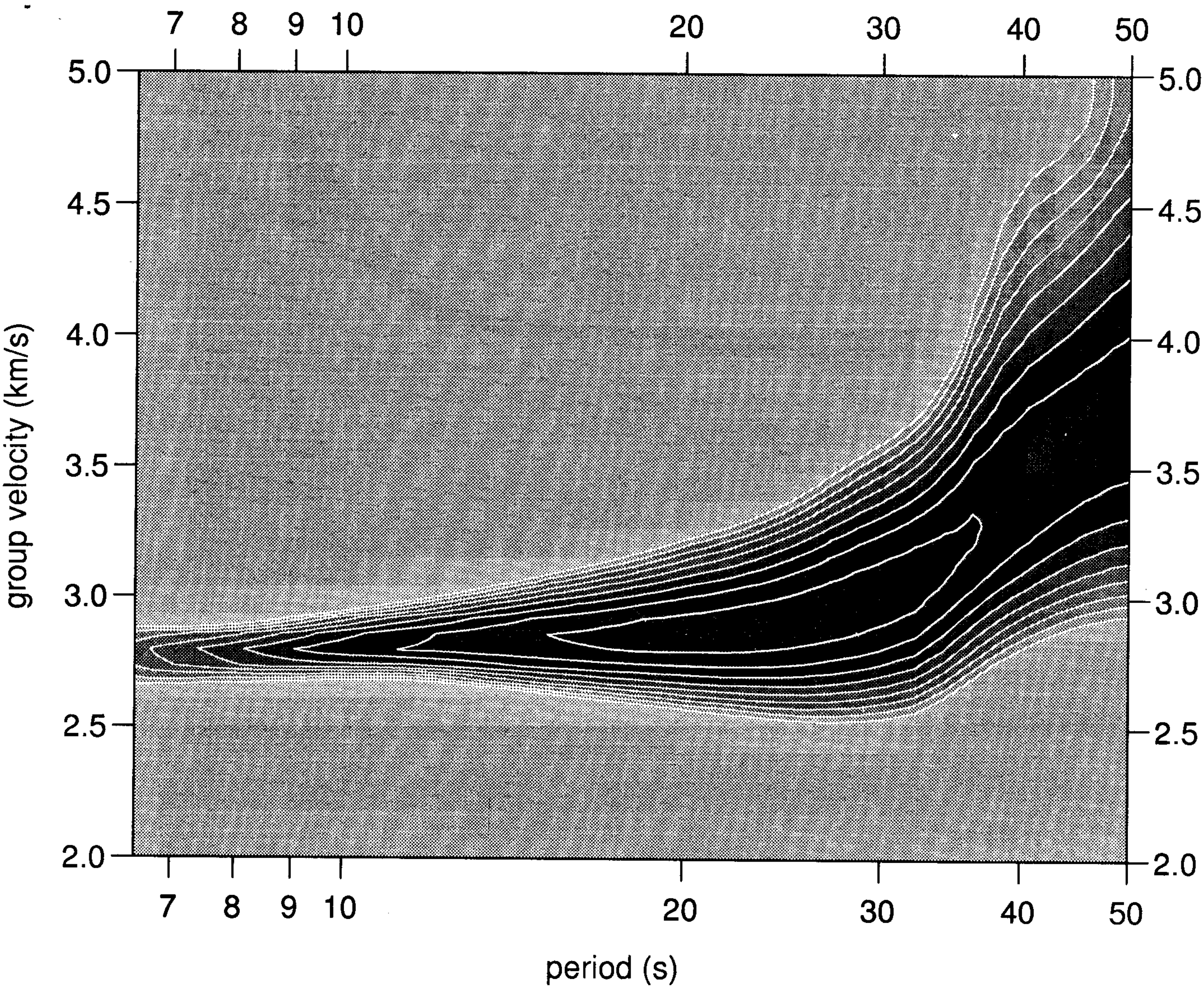

Fig. 4. Period-group velocity diagram obtained after logarithmic stacking of the 9 events for the vertical and radial components (Rayleigh wave). The isolines correspond to the values: 0.01, 0.02, 0.03, 0.05, 0.09, 0.15, 0.25, 0.42, 0.71.

(1986) we assumed that each model consists of five layers. The velocity in each layer was taken as constant. We began with the model obtained from the gradient inversion. For each layer we allowed random changes in the shear-wave velocity and the depth of ± 0.5 km/sec and ± 1.0 km, respectively, while keeping the Poisson ratio and the density constant. For each model we calculated the dispersion curve of the group velocity of the fundamental mode of the Rayleigh wave using the subroutines of Herrmann (1987). If the curve fell in the shaded area determined from the stacking procedure in Figure 5, the corresponding model was kept. We tested 60,000 models and found a set of about 1000 "acceptable" models (those that approximate the observed dispersion curve within the standard deviation). Using this set we computed the mean value and the standard deviation of the shear-wave velocity distribution with depth. The result, illustrated by the shaded area in Figure 6, shows that the velocity is relatively well defined in the middle crust (5 to 30 km) but both shallow and deep structures are not well resolved. This is due to limited period range of the available dispersion data. The short-period information is absent because of the shallow heterogeneities and interference of different modes. On the other hand, the event-station distances are too short to observe well-dispersed long-period surface waves. The inversion shows the existence of a low-velocity upper crust with average shear-wave velocity of about 3.3 km/sec and a lower crust with an average velocity of 3.75 km/sec. The boundary between the upper and lower crust is well defined and lies between 15 and 20 km depth. Finally, based on the inverted velocity distribution, we propose an average four-layer model of the crust between the Guerrero subduction zone and Mexico City. Figure 6 shows this model, which is one of many that approximate the observed dispersion curve, within one standard deviation. Between the depth range of 5 to 30 km, the best resolved range, this model is close to the mean value for all the acceptable models.

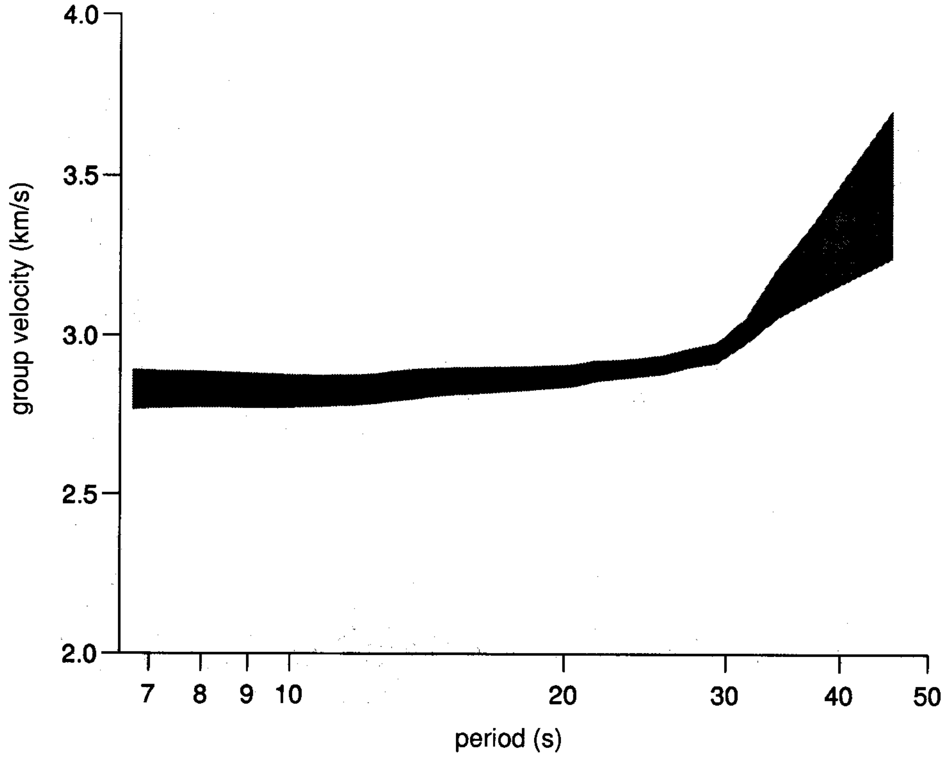

Fig. 5. Rayleigh wave group velocity dispersion curve obtained from the stacking. The shaded area at a given period represents the standard deviation.

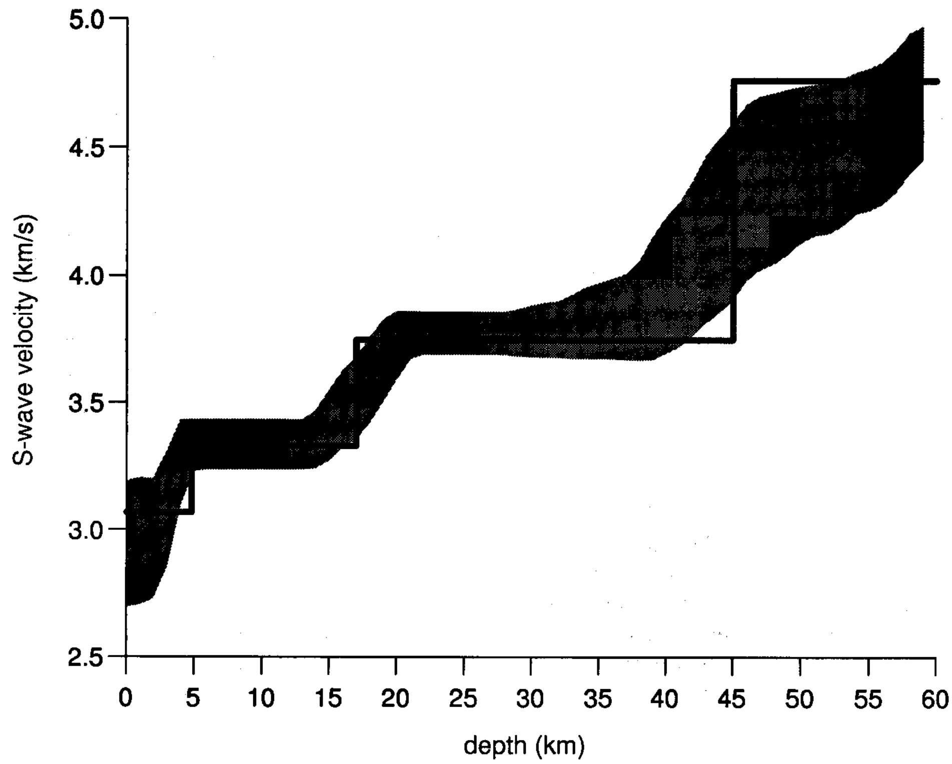

Fig. 6. Shear-wave velocity structure corresponding to the dispersion curve in Figure 5. The shaded area indicates the uncertainty in the result of our inversion. The solid line shows an average four-layer crustal model between Guerrero-Michoacán coast and Mexico City.

In Figure 7, we compare the observed vertical seismogram of event 9 (Table 1) with the synthetic seismograms computed for the inferred crustal structure shown in Figure 6. The focal mechanism used in the computation (strike = 276°, dip = 17°, rake = 67°) is taken from Harvard Centroid Moment Tensor (CMT) solution catalog. The value of the moment is 1.0×10^{19} N-m. Both the observed and synthetic seismograms have been low-pass filtered at 0.1 Hz. This cutoff frequency removes the higher modes (Herrmann and Kijko, 1983) whose amplitudes are strongly dependent on the spectral characteristics of the source. The depths of the source used in the computations are 10, 20, 30, and 40 km. The best fit between the synthetic and observed seismograms is obtained for a depth of 30 km which is close to the depth 35 km determined by the local data (Table 1). The overall agreement between the observed and synthetic depth seismograms gives confidence in the average crustal model and suggests that it could be used in rapid inversion of moment tensor of coastal earthquakes.

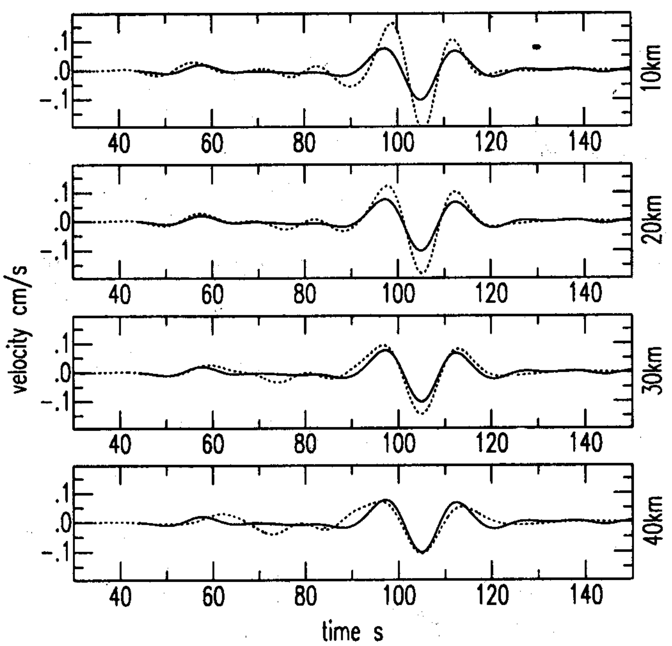

Fig. 7. Low-pass filtered (at 0.1 Hz) observed (continuous line) and synthetic (dashed line) vertical seismograms for event 9. Synthetics are shown for depths of 10, 20, 30, and 40 km.

COMPARISON WITH OTHER CRUSTAL MODELS

In Figure 8, the crustal model inferred in this paper is compared with some others reported for Mexico. These include: (1) Central Mexican Plateau (CMP) model of Gomberg *et al.* (1988), (2) Oaxaca crustal model of Valdés *et al.* (1986), and (3) the model used by Campillo *et al.* (1989) to explain the waveforms recorded in Mexico City during the 1985 Michoacán earthquake. The crustal shear-wave velocity of the CMP model is slightly slower at the top of the lower crust than the one inferred in this paper. The depths of the Moho and the top of the lower crust, deduced by Valdés *et al.* (1986) from the seismic refraction experiment in Oaxaca (Figure 1) are in good agreement with the dispersion curve and the crustal model found in our study. The basic difference between the model we propose, for the region between the coast of Guerrero-Michoacán and Mexico City, and that reported for the Oaxaca region lies in the upper 5 km; in this layer the shear-wave velocity in the former region is much faster than the latter region. Figure 9 shows the dispersion curves associated with the crustal models given in Figure 8. The model obtained in Oaxaca with a thick sedimentary layer, produces a group velocity curve (denoted by (d) in the figure) which is clearly below our measurements given their probable uncertainties for periods lower than 20 s.

IMPLICATIONS FOR GROUND MOTIONS IN THE VALLEY OF MEXICO DURING THE MICHOACAN EARTHQUAKE OF 1985

Campillo *et al.* (1989) were the first to recognize the importance of regional crustal structure on strong-ground motions observed in the Valley of Mexico, produced by earthquakes along the subduction zone. They tried to interpret the large displacements recorded in the valley during the great Michoacán earthquake assuming the crustal model for Oaxaca reported by Valdés *et al.* (1986). However, the displacement records at hill-zone sites in the valley showed arrival times of 2 to 3 s-period Lg and 10 s-period Rayleigh waves which were incompatible with this crustal structure, given the characteristics of the earthquake source-time history revealed by teleseismic records. These records show a strong emission in the period range of 2 to 4 s which starts about 8 s after the beginning of the rupture process (Singh *et al.*, 1990). As the propagation velocity of Lg and Rayleigh waves for the Oaxaca model do not agree with the waves observed in the Valley of Mexico, Campillo *et al.* (1989) modified this model by removing the low velocity layer of about 5 km thickness. This modified crustal model along with the source function deduced from the teleseismic records gave rise to synthetics which were very consistent with observations in the hill-zone of the Valley of Mexico. As shown above, this modified crustal model is also very close to the one determined in this study using independent data. Consequently the present study confirms that the large 3 s ripples observed on hill-zone records of the valley during the 1985 earthquake are Lg waves. Since Lg is a guided wave with geometrical spreading weaker than body waves (and consequently, a slower decay with distance, see Campillo, 1990, for a review of the characteristics of Lg waves), Lg is a very efficient mode of short-period wave propagation. It is worth noting that the natural period of the lake-bed sites in the Valley of Mexico, at which the ground motion suffers great amplification during earthquakes, coincides with these 3 s incident Lg waves. This demostrates the importance of the knowledge of the crustal structure in proper assessment of seismic hazard.

CONCLUSIONS

Using a new stacking technique for the period-group velocity amplitude distribution of individual records of regional events recorded on a broadband seismograph in

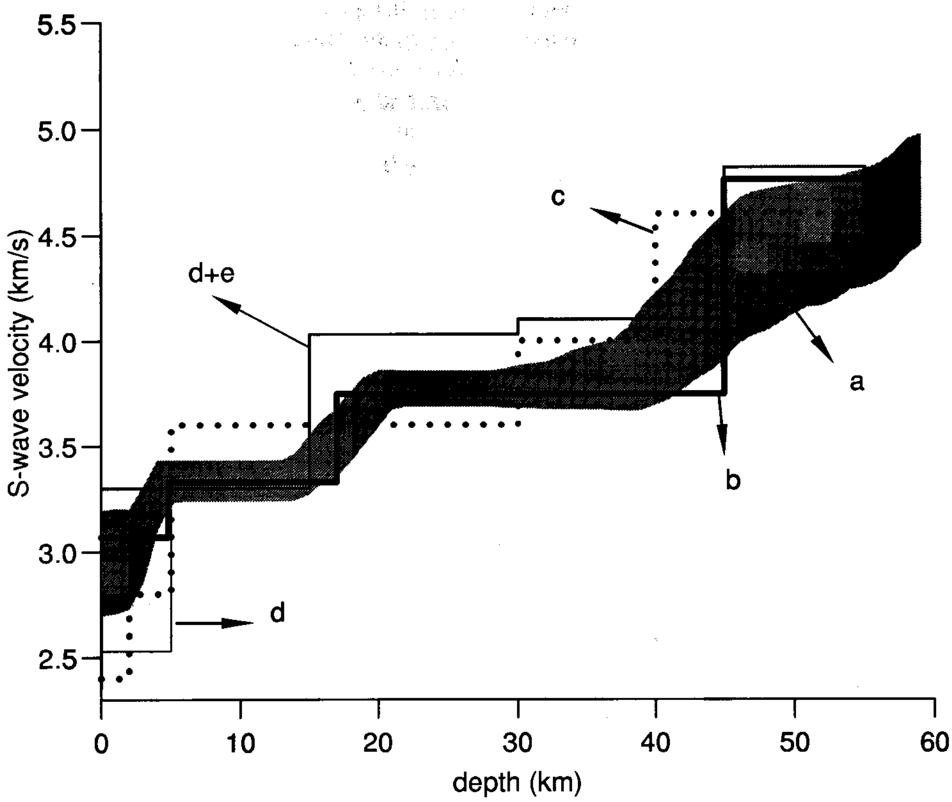

Fig. 8. Comparison of some shear-wave crustal structures in Mexico. a: From inversion of stacked data (this paper). b: The average four-layer model proposed in this paper. c: Inversion of phase velocities, Central Mexican Plateau (Gomberg *et al.*, 1988). d: Shear-wave velocity structure from a seismic refraction experiment in Oaxaca (Valdés *et al.*, 1986), assuming a Poisson solid. e: Same structure as d but without low-velocity shallow layer (Campillo *et al.*, 1989).

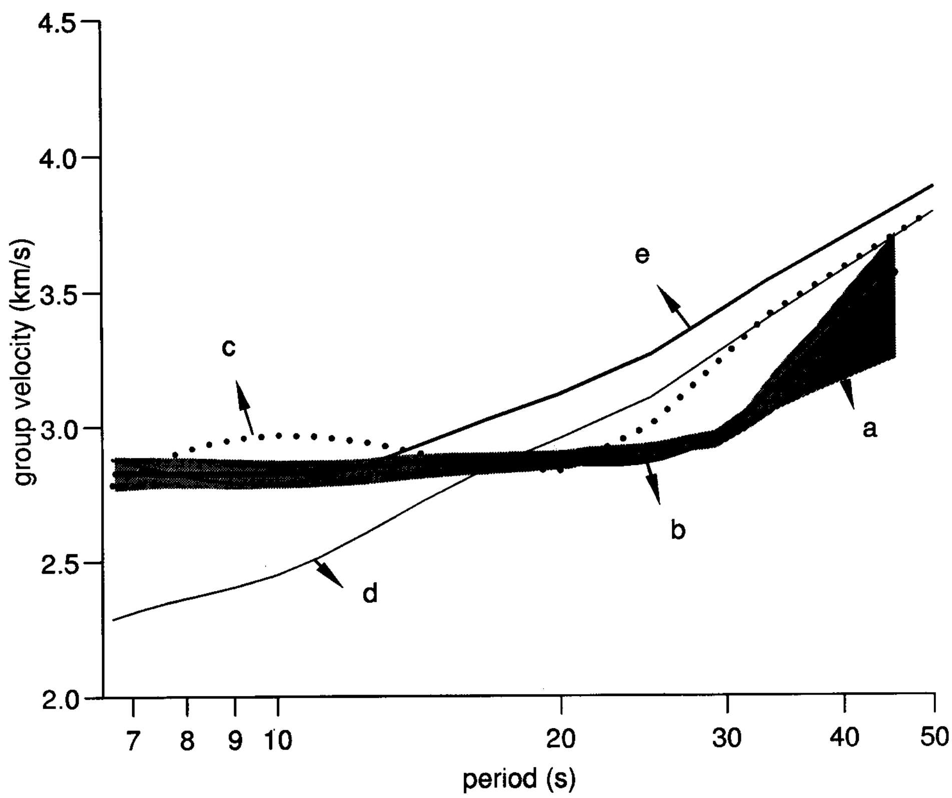

Fig. 9. Group velocity dispersion curves corresponding to the crustal models (a) to (e) of Figure 8.

Mexico City, a mean dispersion curve for the fundamental Rayleigh mode has been constructed for the region between the coast of Guerrero-Michoacán and Mexico City. The average shear-wave velocity of the crust, deduced from this dispersion curve, has the following salient features:

(1) It lacks the thick low-velocity superficial layer reported by Valdés *et al.* (1986) for the Oaxaca region. Our result is in accordance with (a) previous inference based on the analysis of strong motion data recorded in Mexico City during 1985 Michoacán earthquake, and (b) the new Oaxaca model obtained from the seismic refraction traverse reported by the Geolimex Working Group (1993).

(2) The crustal structure is appropiate for modelling of longer-period (from about 8 to 40 sec) seismic waves from earthquakes occurring along the Guerrero-Michoacán coast and recorded at inland stations towards and up to Mexico City and, hence, for the moment tensor inversion of such events.

The dispersion curve and inverted crustal model obtained in this paper provide a basis for comparison with those that might be obtained across the Mexican Volcanic Belt (MVB). This may help to explain the cause of reported regional seismic-wave amplification in and around the Valley of Mexico (e.g., Ordaz and Singh, 1992; Singh *et al.*, 1995).

Clearly we need to map the crustal structure of many more tectonic regions of Mexico. The continuing installation of broadband seismographs in south-central Mexico should provide the needed data for such mapping in the near future.

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