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Crustal velocity structure of southern Guatemala using refracted and Sp converted waves

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RESUMEN

La distribución de velocidades en la estructura cortical del Sur de Guatemala es estimada utilizando los métodos de Velocidad Aparente Mínima y Ondas Convertidas, aplicados a datos sísmicos locales. Los resultados son evaluados reloca-lizando una muestra aleatoria de eventos y por medio de modelado sintético de sismogramas. El modelo de velocidad obtenido posee un gradiente aproximadamente constante en la corteza, que sobreyace una capa intermedia de ~ 15 km de espesor. Las velocidades del Manto son aproximadamente 8.0 km/sec. Este es un modelo promedio para Guatemala con las limitantes impuestas por la compleja tectónica regional.

PALABRAS CLAVE: Guatemala, velocidad aparente, ondas convertidas, estructura cortical.

ABSTRACT

The crustal velocity structure of southern Guatemala is estimated using the methods of Minimum Apparent Velocity and Converted Waves, applied to local seismic data. The results are tested by relocating a random sample of events and by the generation of synthetic seismograms. The velocity model obtained has a roughly constant gradient throughout the crust, underlain by an intermediate layer ~ 15 km in thickness. Mantle velocities are approximately 8.0 km/sec. This is an average model for southern Guatemala with the limitations of the complex tectonics.

KEY WORDS: Guatemala, apparent velocity, converted waves, crustalstructure.

INTRODUCTION

Guatemala is located in a complex tectonic setting, dominated by the interaction of the Caribbean, Cocos and North American plates (Figure 1). Despite the relatively large amount of seismological research done in this region (e.g. Kanamori and Stewart, 1978; Dewey and Suárez, 1991; Ligorría et al., 1995), the crustal structure is not well known. Local crustal models were proposed by Shor and Fisher (1961), White and Harlow (1979), and Chávez (1980). Current hypocentral locations in Guatemala are estimated using a velocity structure derived from the model of White and Harlow (1979).

INSIVUMEH operates the national seismographic network, hereafter called "the network", which covers most of the southern part of the country (Figure 2). Since 1992, the network began digital recording. We use these digital local observations, as well as older analog data, to obtain a new seismic velocity model for southern Guatemala.

For estimating an average crustal structure of Guatemala, we applied the Minimum Apparent Velocity method, which uses travel times from shallow events refracted within the crust. We also applied the Converted Waves method, which estimates the position of the interfaces in the crustal structure from arrival time differences between direct and crustal interface converted phases. The seismic records used in this study were recorded on short-period, vertical component stations of the network. The seismic information allows the application of different seismic techniques and permits internal consistency checks of the models with the observations. A relocation of seismicity was conducted to verify the quality of the results achieved, and synthetic seismograms were compared with actual observations.

MINIMUM APPARENT VELOCITY METHOD

The Minimum Apparent Velocity Method (MAV) was first used by Matumoto et al. (1977) in southern Central America. It has also been used successfully in other complex tectonic environments (e.g. Pardo and Acevedo, 1985; Suárez et al., 1992). F or the description of the method, we follow Suárez et al. (1992).

Consider a layered medium with seismic wave velocity Vi and layer thickness Hi (Figure 3). The raypaths produced by a shallow point source are critically refracted at the different boundaries:

$$\sin \theta_{i,n} = V_i / V_n \quad , \qquad (1)$$

where Øi,n is the angle of incidence in the ith layer for the refracted path reaching the nth layer. The apparent velocity AV of a seismic wave front is defined as

$$\mathrm{AV} = \Delta d / \Delta t \qquad (2)$$

where Dd is the difference in epicentral distance between a pair of seismic stations and Dt is the respective difference in arrival times of the head waves.

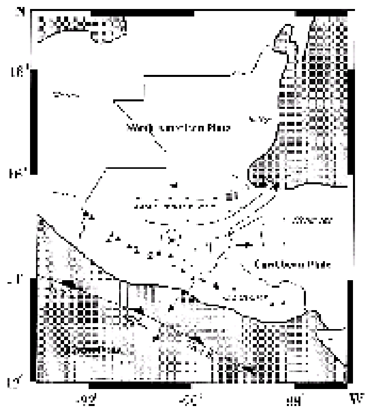

Fig. 1. Main tectonic features of Guatemala (modified after Ligorría et al., 1995). G.C., Guatemala city Graben. HD, Honduras Depression. I.G., Ipala Graben. J.F., Jalpatagua Fault. Arrows show the direction of plate motion. Small dar k triangles indicate approximate position of volcanoes.

As shown in Figure 3, the critical refraction predicted by (1) produces a constant apparent velocity value that shows a sudden increment when a critical epicentral distance is reached. For longer epicentral distances, higher apparent velocities represent refractions from deeper discontinuities. From this diagram, the apparent velocities and critical epicentral distances are read. The thickness of the different n layers of the model is estimated by (Matumoto et al., 1977):

$$H_n = \frac{V_n}{\cos\theta_{n,n+1}}\left[\frac{d_n(V_{n+1}-V_n)}{2V_{n+1}V_n} + \sum_{i=1}^{n-1}\frac{H_i}{V_i}(\cos\theta_{i,n} - \cos\theta_{i,n+1})\right] \quad (3)$$

where Øi,n is the angle of incidence at layer i of a wave critically refracted from layer n, dn is the critical epicentral distance, and Vn is the corresponding layer velocity.

Fig. 2. National Seismographic Network (INSIVUMEH). All stations shown are short-period vertical instruments. Solid triangles identify stations for which Sp converted phases were analyzed.

The MAV method is more reliable when the raypaths are recorded by pairs of stations aligned with the source. The resolution of the method is dependent on the spacing between seismic stations (Suárez et al., 1992). This method can only be applied for shallow events.

Data Selection and Apparent Velocity Analysis

We assembled our dataset based on the following criteria:

- The events were recorded by at least 6 stations of the network,
- The hypocentral location errors (Lienert and Havskov, 1995) were less than 5 km for latitude and longitude, and less than 10 km for depth,
- The event depth was µ 25 km, and
- The ratio of the epicentral distances, for each pair of stations, was µ 0.66

We focused our study on different groups of data. The first was a set of events in 1992-93 (Figure 4). These data include 443 apparent velocities obtained from 175 events split into profiles covering the whole country and conveniently oriented SW-NE, SE-NW, E-W or N-S, in order to test for directional bias. No such bias was found and all profiles were merged into a single group. The second group of data was composed of earthquake relocations from analog records of 1984-1991 including 325 apparent velocities from 126 events (Figure 4).

We were unable to obtain consistent data of clear refractions from the Mohorovicic discontinuity, from local events. Mantle refractions from events with longer epicentral distances (> 250 km), recorded by the network and by stations in El Salvador and Nicaragua (Arriola et al., 1993) suggested a higher apparent velocity layer (~8.0 km/sec), underlying the upper layers initially determined (CAM data in Figure 5).

The final MAV diagram obtained after integrating the different groups of data is presented in Figure 5. The correspondent crustal model is summarized in Table 1. The different groups of data give matching results. A visual check of the scatter of the data in Figure 5 suggests that the accuracy of the velocity determination is roughly ±0.2 km/sec, and the critical epicentral distances are accurate to about ±15 km. This results in a mean error of about ±2.1 km in the depth determination of the interfaces (Table 1).

CONVERTED WAVES METHOD

The Converted Waves (CW) method is based on the conversion of elastic waves at an interface between two layers with different elastic properties. The arrival times and the geometry of the wave paths, can be predicted from

$$V_i / \sin\theta_i = V_r / \sin\theta_r \qquad (4)$$

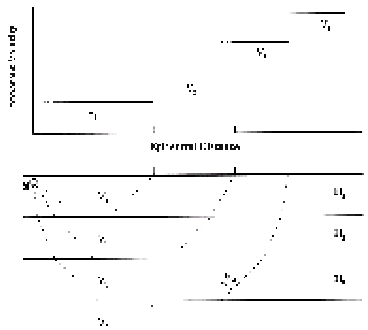

Fig. 3. The Minimum Apparent Velocity method. The ray paths (dashed lines) from a point source S are critically refracted along the nterfaces of the velocity model. The observations on the surface (thick line) are translated into the MAV diagram (Epicentral Distance vs. Apparent Velocity) to infer the velocities V and the thicknesses H for each layer of the model.

where V is the velocity of propagation and q is the angle of the incident (i) or refracted (r) waves.
The ray paths depend only on the velocity contrasts, but the conversion itself is mostly ruled by the type of incident wave and the transmission coefficients (Lay and Wallace, 1995). We limited our analysis to incoming S waves that are transmitted as S and Sp (S converted to P) (Figure 6). The incident wave has to be of the SV type as otherwise an S to P conversion is not possible. Since it is out of the aim of our study to explain the transmission phenomena, we hereafter call our incident wave simply as S wave.
Using the CW method, we calculate the position of the interface by matching theoretical calculations with observed values of the difference in arrival times between incident (S) and converted (Sp) waves. The travel times can be calculated by (Ligorría and Ponce, 1993):

$$T = \frac{D_n}{V_n \cos\theta_n} + \sum_{i=1}^{n-1} \frac{H_i}{V_i \cos\theta_i} \qquad (5)$$

where qi is the incidence angle in the i-th interface, Dn is the depth of the source with respect to the n-th interface, Vi is the velocity of S or P (Sp) waves and Hi is the thickness of each layer of the proposed velocity model. A constant Poisson's ratio of 0.25 is assumed.

Events located below the Mohorovicic discontinuity at a relatively short epicentral distance are selected. The comparison between calculated and observed values is expressed in terms of time residuals, assuming that the optimal model minimizes the residuals. Several comparisons from different stations are made simultaneously and the sum of residuals is expressed in terms of RMS values (Table 2).

The CW method has been applied by several authors, e.g. Båth and

Steffansson (1966); Jordan and Frazer (1975); Sacks and Snoke (1977); Chiu et al. (1986). Our processing sequence was as used by Ligorría and Ponce (1993).

Differential time measurements and modeling

The CW method was applied to confirm or correct the model formerly estimated by the MAV method, and to

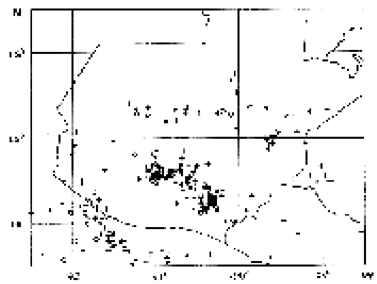

Fig. 4. Seismic events used with the Minimum Apparent Velocity method. Circles show the epicenters of the events chosen from the 1992-93 digital recordings of the network. Crosses are relocations made using analog records of the 1984-91 recording period of the network.

Table 1

Initial crustal model (from MAV application)

Layer	Velocity † (km/sec)	Depth to interface (km)	Critical epicentral distance (km)††
1	5.0	0.0	55
2	5.7	7.0±2	130
3	6.0	15.6±3	160
4	6.6	26.9±2	180
5	7.0	35.1±2	210
6‡	~8.0	49.8±2	-

† Velocity of P-waves, ± 0.2 km/sec. (see text).
†† ± 15 km (see text)
‡ Layer 6 is considered as a half-space underlying the crustal model and with a velocity around 8.0 km/sec.

estimate the depth of the Mohorovicic discontinuity. Events recorded by the network were chosen for the 1992-93 period, because of the better resolution of digital data. The selection criteria was that the epicentral distance should be µ 1.0° (~111 km) and the hypocentral depth > 50 km. We used 39 conversions for 20 events recorded at 9 different stations (Figure 2, Table 2). The S-Sp time differences observed at each station were grouped for each interface (Table 2), taking the MAV model as a starting point.

The procedure was as follows.

- We shifted the deepest interface and matched the observed TS-TSp values to the calculated ones, until we reached an optimum RMS value of the residuals.
- The next shallow boundary was shifted, keeping the deepest interface fixed, until the lowest RMS value was achieved.
- Before proceeding to the next shallow discontinuity, we shift again the previous interface, in order to check if the solution is still optimal.
- · We continue with all interfaces in a similar manner.

The residuals from all stations and interfaces are presented in terms of RMS values in Table 2. The accuracy g of the results was calculated as a linear function of the RMS; i.e. g = (V/0.73)*RMS, where V = 6.2 km/sec is the mean velocity of compressional waves in the crust, and Vp/Vs ~ 1.73. The velocity model obtained is shown in Figure 7 and summarized in Table 3.

DISCUSSION

The seismicity used in both experiments had been located earlier with a different velocity model (Figure 7). Thus, we relocated events with the new model for a ran-

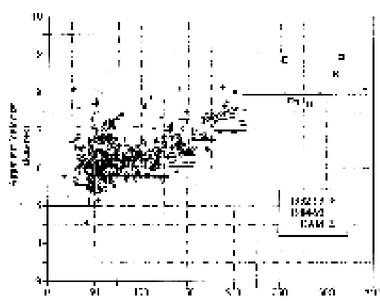

Fig. 5. MAV diagram showing the datasets used (1992-93, 1984-91 and CAM events). The CAM data were obtained from events located at epicentral distances larger than 250 km, recorded by the network and by stations in El Salvador and Nicaragua. These events were included in the analysis as an attempt to define the position of the Crust-Mantle boundary.

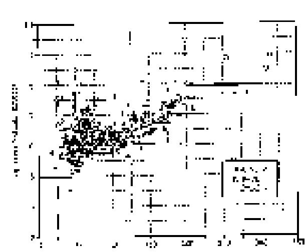

Fig. 6. Geometry of the Sp converted wave. The S-wave front from the source is

converted at the interface between two layers with different velocities (Vi). The phases from such conversion are recorded at the Station, and identified as S and Sp (S converted to P) waves (see Fig 8).

dom sample of 51 events from 1992-93. The results of this relocation are summarized in Table 4.

The location differences in latitude and longitude (Table 4) show the influence of the MAV model. The mean epicentral difference is about 1.5 km which represents 10% of the critical epicentral distance accuracy (±15 km) (Table 1). This is a small effect. The CW estimation yields an average depth difference of 7.5 km which does not produce drastic changes in the basic geometry of our model, since it represents only 5% of the mean depth of the events used (~141 km) (Table 2).
The second test aimed to confirm that the phases read in the CW estimation were correctly identified under the idealized conditions of the method. Synthetic seismograms of 4 selected events of the database (Table 2) were calculated, to check S and Sp phases of both observed and synthetic signals. The focal mechanisms used to calculate the synthetic seismograms were taken from s imilar events (Redondo et al., 1993) adjusted by trial-and-error to the real seismograms, until a match of the amplitudes of the signals (P and S-wave phases) was achieved. Figure 8 shows two examples. The results show that, under the conditions of our experiment, the direct S waves produce converted Sp phases with amplitudes likely to be read. The synthetics were obtained by using the Computer Programs in Seismology package (Herrmann, 1990a,b).

Table 2

[illegible]

[illegible]	Station	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]
1	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]
2	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]
3	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]
4	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]
5	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]

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Table 3

Final Crustal Model (after CW application)

Depth to interface ± γ† (km)	Velocity†† (km/sec)
0	5.0
7±3	5.7
14±2	6.0
24±3	6.6
31±4	7.0
46±3	8.0

† Accuracy (γ) calculated from RMS values in Table 2 (see text).
†† P-wave velocity.

Depth to interface ± g _ (km)Velocity__ (km/sec)05.07±3 5.714±2 6.024□.6 31±4 7.0 46,8.0 Accuracy (g) calculated from RMS values in Table 2 (see text).
P-wave velocity.

CONCLUSIONS

We integrated results from two techniques, applied to different sets of data: shallow and deep seismicity. Both methods were found to be complementary. The MAV method yields reliable estimates of the velocities of the model, and the CW method improves the depth estimates of the interfaces of the model. When the discontinuities were not clearly defined by the MAV method, the CW method helped to obtain a suitable definition.

Our velocity model definition shows a typical continental crustal structure with thickness and velocities similar to those obtained in Chiapas, Mexico (Castro, 1980) and northern Costa Rica (Matumoto et al., 1977). When compared with other models (Figure 7), we propose lower

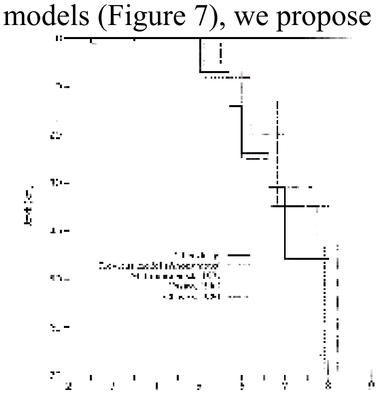

Fig. 7. Final velocity model obtained in this study (solid line), and comparison with other models. The previous model (Anonymous) is the velocity structure routinely used at INSIVUMEH as derived from the model of White and Harlow (1979). The differences between our model and the Anonymous model (heavy dashed lines) were tested by hypocentral relocation (see Table 4).

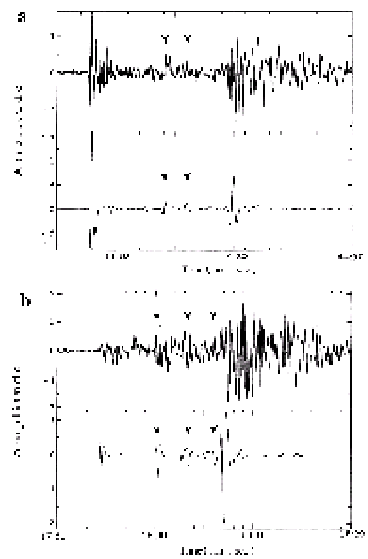

Fig. 8. Two synthetic seismograms, plotted with the respective observed seismograms. a) Station: YUP, Event: 921103 08:43, Depth: 103 km, Epicentral Distance: 35 km, Focal Mechanism: Strike: 110°, Dip: 50°, Rake: 314°. b) Station: IXG, Event: 930123 06:17, Depth: 104 km, Epicentral Distance: 48 km, Focal Mechanism: Strike: 95°, Dip: 35°, Rake: 55°. The arrows indicate prominent Sp converted phases produced at different interfaces. Observed and Synthetic signals were high-pass fil tered with a corner frequency at 1 Hz. Amplitude values are normalized.

velocity values for shallower layers between 10 and 30 km depth, as in the model obtained from surface wave dispersion by Chávez (1980). The Mohorovicic depth is consistent with the results of Matumoto et al. (1977) and Castro

Table [illegible]

[illegible]

[illegible]	Mean	RMS
T0	[illegible]	0.2
Longitude	[illegible]	2.0
Latitude	[illegible]	[illegible]
Depth	-0.5 km	[illegible]

b) Average Station Residuals

Station	P-wave residuals			S-wave residuals		
	Mean	Std. deviation	No. of readings	Residual	Std. deviation	No. of readings
BDC	0.17	0.24	32	0.04	0.12	19
QZC	[illegible]	[illegible]	[illegible]	[illegible]	0.22	20
JUC	0.05	[illegible]	[illegible]	[illegible]	0.22	19
TAK	[illegible]	[illegible]	[illegible]	[illegible]	0.15	16
YUP	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	5
[illegible]	[illegible]	[illegible]	[illegible]	-0.04	[illegible]	[illegible]
[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]
STP	[illegible]	0.27	[illegible]	[illegible]	[illegible]	[illegible]
[illegible]	[illegible]	[illegible]	12	[illegible]	[illegible]	[illegible]
[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	25
[illegible]	[illegible]	[illegible]	[illegible]	-	-	-

[illegible]

(1980) (43-46 km). We conclude that this model is a good first-order approximation of the velocity structure of Guatemala.

A major difference between the velocity model obtained here and the previous model used in Guatemala (Anonymous, Figure 7), is the location of the Mohorovicic discontinuity. As seen in our relocation test, this difference may influence the hypocentral depth determinations. We believe that more reliable results will be obtained as this model is introduced in routine analysis work. The results obtained by this study are limited to the southern part of Guatemala, located in the Caribbean plate. The st ation coverage of the network does not yet provide reliable data for the crustal structure of the northern part of the country, which is located in the North American Plate. This limits the accuracy of both methods applied here, since both are based on si mple geometric ray theory.

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