The April 22, 1991, Valle de la Estrella, Costa Rica (Mw = 7.7) Earthquake and Its Tectonic Implications: A Broadband Seismic Study

SASKIA D. B. GOES, AARON A. VELASCO, SUSAN Y. SCHWARTZ AND THORNE LAY

Institute of Tectonics and C. F. Richter Seismological Laboratory, University of California, Santa Cruz

The rupture process of a large back arc thrusting earthquake, the April 22, 1991, Valle de la Estrella, Costa Rica (Mw = 7.7), earthquake, is investigated using broadband body waves and long-period surface waves. We find the source process to be relatively simple, with the source models separately obtained from body and surface waves being very consistent. The event occurred on a shallow, southwest dipping rupture plane on which most energy is released updip of the hypocentral location (10-20 km deep). High-frequency radiation appears to have been released over a relatively small source area. Our preferred model has a focal mechanism with strike 102 ± 10º, dip 17 ± 14º and rake 63 ± 17º, a seismic moment of 3.8 ± 1.5 x 10^20 N m, and a total rupture duration of 40 ± 6 s. The earthquake appears to be associated with the North Panama Deformed Belt (NPDB), a thrust and fold complex that has accommodated the oroclinal deformation of Panama. This event, along with previous large events north of Panama in 1882 and 1916, indicates that there is substantial convergence along the NPDB, marking the NPDB as a probable emerging plate boundary. It remains difficult to gauge the earthquake hazard in the region because of the tectonic complexity.

INTRODUCTION

Frequent large earthquakes occur on the west coast of Costa Rica due to subduction of the Cocos plate beneath the Caribbean plate along the Middle American Trench. Much of the concern about earthquake hazard in this region is thus focussed on a large seismic gap along the Nicoya Peninsula (Figure 1) which last fully ruptured with an Ms = 7.7 event in 1950 [e.g., Nishenko, 1991]. However, on April 22, 1991, a major earthquake (National Earthquake Information Center (NEIC) 2156:51.82 UT, 9.685ºN, 83.073ºW, m_b = 6.3, M_S = 7.6, M_w = 7.7) occurred along the east coast of Costa Rica near the border with Panama (Figure 1), killing 75 people, injuring 560, and leaving close to 10,000 homeless [NEIC, 1991]. Buildings and roads in the area around the coastal city of Limón suffered major damage. The earthquake was felt as far north as El Salvador and triggered a tsunami with local heights of up to 2 m [NEIC, 1991]. The event occurred in a region, known as "Valle de la Estrella", with poorly characterized seismic potential, a back arc thrust environment where Central America is being underthrust by the Caribbean plate. Back arc activity thus constitutes a major seismic hazard in the region in addition to that posed by the subduction zone on the Pacific margin.

The region around Costa Rica and Panama is tectonically very complicated [e.g., Dengø, 1985; Escalante, 1990; Holcombe et al., 1990; Mann et al., 1990]. The Cocos, Nazca, and Caribbean plates, and perhaps some minor plates, interact in a complex fashion to accommodate the evolving plate boundaries in the region (Figure 1). The complex plate interactions result from (1) the collision between Panama and the South American plate which produced oroclinal rotation of Panama, and perhaps formation of a distinct Panama Block [e.g., Silver et al., 1990], and (2) the subduction of the Cocos-Nazca plate boundary (the Panama Fracture Zone) and the young lithosphere near the Cocos Ridge along the Middle American Trench [e.g., Jacob et al., 1991; Protti, 1991]. The Cocos Ridge distorts the trench and subducts seismogenically, with large events in 1904 (M_w = 7.6), 1924 (M_w = 7.0), 1941 (M_w = 7.5), 1952 (M_w = 7.2), and 1983 (M_w = 7.2) [Astiz and Kanamori, 1984]. Highly oblique convergence between the Nazca and Caribbean plates is probably taken up along a diffuse zone, possibly involving plastic deformation of Panama [Silver et al., 1990] or the presence of one or more microplates [e.g., Adamek et al., 1988; Vergara-Muñoz, 1988; Silver et al., 1990]. While there is a clear offshore thrust complex south of Panama [Silver et al., 1990] and sparse seismic lineation, there have not been large thrust events in this region in the past few decades, and the nature of present motions in this region is unclear [DeMet et al., 1990; Jordan, 1975]. If this is the principle plate boundary between the Nazca and Caribbean plates, current motions in this region are expected to be largely strike slip [Jordan, 1975].

Some of the regional convergent and rotational motions appear to be accommodated along the North Panama Deformed Belt (NPDB) (Figure 1). North of Panama this wide belt of thrusts and folds has been well defined by seismic reflection surveys [e.g., Silver et al., 1990]. Towards the coast of Costa Rica however the belt narrows and probably comes onshore near the port of Limón, where onshore thrust faults have been previously associated with the NPDB [Ponce and Case, 1987]. The Valle de la Estrella earthquake, which appears to have ruptured a fault which extends offshore and generated a small tsunami, provides evidence of a somewhat wider thrust zone still encompassing some offshore motion, but probably coming onshore north of Limón [Plafker and Ward, 1992]. The extension of the thrust belt from where it comes onshore is unclear, but it has been suggested that the NPDB may terminate through a diffuse set of strike-slip faults in central Costa Rica [Ponce and Case, 1987; Jacob et al., 1991; G. Suarez et al., unpublished manuscript, 1992]. An alternative suggestion is that the thrust belt extends as far north as the Hess Escarpment [Escalante, 1990; Plafker and Ward, 1992]. It is within the apparent onshore extension of the thrust belt that the Valle de la Estrella earthquake is located.

Large back arc thrusting events are quite rare, with the best studied similar earthquakes being the 1983 Akita-Oki (M_w = 8127
Fig. 1. Map illustrating the plates and tectonic features of the Costa Rica-Panama region. The Cocos-Nazca plate boundary is the Panama Fracture Zone (PFZ). Most seismic activity is associated with the Middle America Trench (MAT), where the Cocos plate subducts beneath the Caribbean plate. The epicenters of two well located earthquakes (1950: $M_s=7.7$, 1983: $M_s=7.2$) along this zone are marked by solid stars. The location of other plate boundaries is less clear. The epicenter of the April 22, 1991 ($M_w=7.7$), Valle de la Estrella, Costa Rica earthquake (solid star) is located away from the MAT, near the western edge of the North Panama Deformed Belt (NPDB). Three other historical earthquakes (two in 1916: $M_s=7.1$ and $M_s=7.4$, and one in 1882; all indicated by open stars with poorly constrained locations) have been associated with motion along the NPDB.

7.8) [Kanamori and Astiz, 1985; Satake, 1985] and 1964 Niigata ($M_w = 7.6$) [Abe, 1975; Satake and Abe, 1983] earthquakes in the Sea of Japan. The occurrence of such large back arc events is often associated with the early stages in the development of a new plate boundary. A search of the Harvard centroid moment tensor (CMT) and the Costa Rica regional network catalogues reveals that the NPDB had a low background seismicity level before the 1991 Valle de la Estrella earthquake. However, there is evidence of large historic earthquakes that may be associated with underthrusting of the Caribbean plate beneath Panama. de Pinilla and Toral [1987] and Mendoza and Nishenko [1989] have investigated a 1882 earthquake that caused extensive damage in northern Panama and was most likely located within the central or eastern part of the NPDB (the source location from Mendoza and Nishenko [1989] is shown in Figure 1). On April 24 and April 26, 1916, two major earthquakes ($M_s = 7.4$ and 7.1 [Abe, 1981]) occurred on the western edge of the NPDB [Kirkpatrick, 1920; Güendel, 1986; de Pinilla and Toral, 1987] (the source locations in Figure 1 are from intensity patterns studied by de Pinilla and Toral). This doublet has also been associated with thrusting motion along the NPDB [Güendel, 1986], and Jacob et al. [1991] report a thrust fault mechanism for the April 24, 1916, event. All of these events occurred east of the Valle de la Estrella epicenter where the thrust belt widens and probably takes up more of the regional convergent and rotational motion [Silver et al., 1990]. Montero et al. [1991a] report that the Limón region, north of the epicenter, had suffered earlier earthquakes in 1798, 1822, and 1953, with the 1822 event having similar damage patterns to the 1991 event (G. Suarez et al., unpublished manuscript, 1992). In this context, the event of April 1991 is particularly significant, since it indicates that the NPDB is much more active than previously acknowledged, and it gives evidence that the NPDB extends at least as far north as Limón. The relatively high level of seismic activity on the NPDB suggests that this region may be an emerging plate boundary. We investigate seismic waves from the 1991 event to constrain the faulting process and then discuss the tectonics further.

There is an extensive ground motion data set available for this earthquake, including body wave, surface wave, geodetic (both Global Positioning System (GPS) trilateration and coastal uplift observations), and aftershock data. This allows for a study of the source rupture process over a broad range of frequencies. Figure 2 shows the locations of well-located aftershocks determined from portable seismometers deployed in the vicinity of the mainshock [Schwartz and Protti, 1991], along with the best double-couple mechanisms from all reported centroid moment tensor solutions for events in the sequence [Dziewonski et al., 1992]. NEIC locations are shown for the mainshock and for large aftershocks in the first 2 weeks of activity. The aftershocks extend along the coast for
over 100 km, but there is substantial diversity in the mechanisms. One of the largest aftershocks, on April 24, has a strike-slip mechanism and locates near the western edge of the aftershock activity. Other strike-slip mechanisms are found in this region of the aftershock zone [Montero et al., 1991a], so it is clear that induced slip on various faults occurred. In May 1991, several large aftershocks appear to have extended the rupture zone eastward along northern Panama. In this paper, we investigate the compatibility of source models for the Valle de la Estrella event derived from broadband body waves (2-30 s) with those derived from long-period Rayleigh and Love waves (157-288 s) in an effort to detect any source finiteness or unusual source complexity. We compare our results with the geodetic and aftershock data processed by other investigators. We then consider whether the 1991 earthquake gives us any insight into the complicated tectonics and nature of the motion along the NPDB.

SURFACE WAVE ANALYSIS

We analyze long-period (157-288 s) Rayleigh and Love waves obtained from the Global Seismic Network (GSN), GEOSCOPE (Institut National des Sciences de l'Univers, France), and International Deployment of Accelerometers (IDA) networks to determine the faulting duration, centroid depth, moment tensor, and centroid location of the 1991 earthquake. We utilize a moment tensor spectral inversion method originally developed by Kanamori and Given [1981] and later modified to a two-step procedure which isolates the duration calculation from the determination of the centroid depth and moment tensor [Romanowicz and Guillemant, 1984; Zhang and Kanamori, 1988a,b]. We also make use of an optimal centroid location search [Velasco et al., 1992], which stabilizes the moment tensor determination. The long-period source parameters depend significantly on the assumed source location, propagation correction models, source velocity structure, and attenuation models [Wallace et al., 1991; Velasco et al., 1993]. We consider a suite of Earth models and assumed source locations to estimate the confidence bounds of our source parameters.

Summary of Surface Wave Results

Allowing for the model and centroid location dependence, our preferred solution from surface waves has a major double couple with strike ($\phi$) = $107 \pm 5^\circ$, dip ($\delta$) = $21 \pm 10^\circ$, rake ($\lambda$) = $56 \pm 11^\circ$, centroid depth ($h$) = $22 \pm 8$ km, trapezoid source time function duration ($\tau$) = $40 \pm 6$ s, and seismic moment ($M_o$) = $3.7 \pm 0.5 \times 10^{20}$ N m ($M_w = 7.7$), with an 8% non-double-couple component (Figure 3). Our moment tensor elements as well as those for Harvard's centroid moment tensor (CMT) are listed in Table 1, with good agreement between the two long-period analyses. The Harvard CMT moment estimate is $3.3 \times 10^{20}$ N m [Dziewonski et al., 1992] for a best double couple with $\phi = 103^\circ$, $\delta = 25^\circ$, $\lambda = 58^\circ$. The error estimates on our double couple parameters encompass the range of results obtained using a variety of models. We find a centroid location approximately 30 to 40 km northeast of the NEIC epicentral location. $M_o$ has considerable uncertainty due to strong trade-offs with the dip and depth estimates. Furthermore, shallow dipping, shallow depth earthquakes do not strongly excite the surface wave moment tensor terms $M_{xz}$ and $M_{yx}$ (dip-slip components) and thus have large uncertainties in dip and moment. However, the surface wave source parameter estimates are consistent with a shallow, southwest dipping thrust plane, as expected for the NPDB. The excellent fit to the data of our model (Figure 4) and the lack of scatter in this high-quality long-period data set suggest that no complex rupture process occurred (i.e., no significant directivity, change in focal mechanism, or unusually slow
rupture). We briefly outline below the procedures followed in obtaining these long-period source parameters and their uncertainties for this earthquake.

Procedure, Results, and Uncertainties

In the first step of the surface wave inversion method, we invert the phase spectra of Rayleigh and Love waves for the best trapezoid duration. We assume a simple trapezoid source function since the long-period data are insensitive to details of the source. Our primary sensitivity is to the centroid time of the source function, or τ/2. We invert for duration at seven different periods (157, 175, 200, 225, 256, 275, 288 s) utilizing a suite of phase velocity models for propagation corrections: the radially symmetric model PREM [Dziewonski and Anderson, 1981], and the aspherical earth models M84C [Woodhouse and Dziewonski, 1984], and MPA [Wong, 1989] of spherical harmonic degree 8 and 12, respectively. The duration is directly affected by the phase velocity model used to correct phase back to the source. The duration estimates are given by minimizing the normalized RMS error in the first-step inversion, σ [see Zhang and Lay, 1989], which measures the phase misfit. For MPA, we estimate τ = 40 s (Figure 5a), while PREM and M84C give τ = 44 s and τ = 50 s, respectively. Model MPA produces a particularly good fit to the phase data, as has been shown for other earthquakes [Velasco et al., 1992, 1993]. Thus, we prefer this model for propagation corrections. Scatter exists between estimates at different periods, mostly due to inadequacies of the phase propagation corrections, but there is no systematic trend suggesting a complex source process. We therefore estimate τ = 40 ± 6 s with the error encompassing the scatter in duration estimates for model MPA. The corresponding centroid time estimate is 20 ± 3 s.

In the second-step inversion, we solve for a point source depth and the moment tensor utilizing both amplitude and phase information along with the source duration estimate obtained in the first step. Depth and moment tensor estimates depend on the choice of the global attenuation (parameterized by the attenuation quality factor, Q(τ)) and source velocity structure models [Wallace et al., 1991; Velasco et al., 1993]. Large uncertainties exist for long-period surface wave depth determinations due to the uncertainties in the models [Velasco et al., 1993]. The least constrained aspect of the long-period spectral inversion method is the source excitation structure, from which we calculate excitation functions. This is due to the lack of a priori knowledge of the regional crust and upper mantle velocity structure. In this particular case, the velocity structure below the Caribbean coast of Costa Rica is poorly documented, with only one investigation publishing results for northern Costa Rica [Matumoto et al., 1977]. We combined this crustal model with upper mantle models for tectonically active North America in an attempt to construct a regional source model. Using this model for computing surface wave excitation resulted in a doubling of the second-step error, ρ, relative to standard Earth model structures RA of Regan and Anderson [1984] and PREM of Dziewonski and Anderson [1981]. The second-step inversion error (ρ) is a weighted RMS error which measures the misfit between the data and the modelled spectra [see Zhang and Lay, 1989]. The standard Earth models apparently represent better approximations of upper mantle P and S velocity structures, which affect the long-period excitations more strongly than crustal structure. We thus restrict ourselves to the standard Earth models, recognizing these are unlikely to be very accurate representations of the local crustal structure. Depth estimates vary from 15 km to 30 km depending on the choice of global Q models (MG, Masters and Gilbert [1983]; DS, Dziewonski and Stein [1982]; and PREM, Dziewonski and Anderson [1981]) and the choice of excitation structures (RA and PREM) (Figure 5b). These uncertainties are enhanced by the shallow depth and shallow dip of the source, where the dip-slip moment tensor terms (Myz and Mxz) become very difficult to resolve. Our final result, h = 22 ± 8 km, encompasses the range of results obtained for the models considered, but shallower source depths are plausible, since this type of long-period analysis often results in overestimates of the source depth [Wallace et al., 1991].

Optimizing the centroid location is important for stabilizing the second-step moment tensor inversion, as it reduces effects of model uncertainties [Velasco et al., 1992]. To locate the centroid, we contour the residual variance in the second-step moment tensor inversion, ρ, over a 130 x 100 km grid of assumed point source locations near the epicenter (Figure 6b). Since ρ is a measure of how well the data are fit by the associated point source moment tensor for that source position, the location of the minimum of ρ gives the optimal centroid location and moment tensor. The minimum ρ locates just offshore, approximately 30 to 40 km northeast of the onshore NEIC.

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TABLE 1. Moment Tensor Solutions for the Valle de la Estrella Earthquake

<table>
<thead>
<tr>
<th>This Study</th>
<th>Resultant Terms</th>
<th>CMT</th>
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<tbody>
<tr>
<td>Mxx + Myy</td>
<td>1.00 ± 0.02</td>
<td>2.39</td>
</tr>
<tr>
<td>Myy - Mxx</td>
<td>-2.39 ± 0.03</td>
<td>-1.70</td>
</tr>
<tr>
<td>Mxy</td>
<td>-1.07 ± 0.01</td>
<td>0.43</td>
</tr>
<tr>
<td>Myz</td>
<td>1.58 ± 0.24</td>
<td>1.84</td>
</tr>
<tr>
<td>Mxz</td>
<td>1.06 ± 0.02</td>
<td>1.99</td>
</tr>
<tr>
<td>Mzz</td>
<td>1.06 ± 0.02</td>
<td>1.07</td>
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In units of 10²⁰ N m.
Fig. 4. Moment tensor inversion results at periods of 200 and 256 s from simultaneous inversion of Rayleigh and Love waves. The curves are the theoretical fits determined from the solution found using the Regan and Anderson [1984] excitation structure, the Dziewonski and Stein [1982] global Q, and MPA [Wong, 1989] phase velocity models. There is no suggestion of directivity in the data.

epicenter (Figure 6a). At the optimal centroid location p is reduced by a factor of 2 compared with p at the epicentral location. Locating the earthquake at the epicenter also results in a southwest dipping plane with a steep dip (Figure 6a) which violates the teleseismic first-motion data (Figure 8), while the solution at the centroid is consistent with first motions. The small variations in the major double couple solutions near the optimal centroid (Figure 6a) demonstrate the stability of the moment tensor solutions near the centroid. In this case, a shift in source location from the epicenter is necessary to obtain a stable source model. Although some propagation model deficiency probably contributes to the centroid shift, the northeast relative location of the centroid is consistent with our body wave results discussed in the next section, observed coastal uplift, and the small tsunami produced by this event [Pfaffker and Ward, 1992]. Thus, while we do not resolve any directivity in the long-period waves for this event, it is likely that at least some of the centroid location shift corresponds to a seaward shift of the true physical centroid of moment release relative to the NEIC hypocenter. The Harvard CMT centroid location is about 60 km northeast of the epicenter, even further offshore. Velasco et al. [1992] discuss the significance of these centroid estimates at greater length.

**BODY WAVE ANALYSIS**

Body waves can further constrain details of the source rupture process. To study these shorter period waves, we analyze P and SH waves from 11 broadband GSN and GEOSCOPE stations. We invert these data using the iterative method developed by Kikuchi and Kanamori [1982, 1986, 1991] to
Fig. 5. Duration and depth estimates for the Valle de la Estrella Earthquake. (a) Normalized error in the first-step source duration inversion, \( \sigma \), versus assumed trapezoid source duration using model MPA for propagation corrections. The minimum of each curve is the duration estimate from the inversion of waves of that particular period. The optimal mean duration over the period range is \( 40 \pm 6 \) s, with the error encompassing the scatter in estimates for each period. (b) Normalized error in the second-step moment tensor inversion, \( \rho \), versus point source depth for various combinations of excitation structures (RA, Regan and Anderson [1984] and PREM, Dziewonski and Anderson [1981]) and global Q models (MG, Masters and Gilbert [1983]; DS, Dziewonski and Stein [1982]; PREM, Dziewonski and Anderson [1981]), using MPA for propagation corrections. Lack of a priori knowledge of source velocity structure makes depth determination difficult, and we estimate the centroid depth \( (h) = 22 \pm 8 \) km.

Fig. 6. (a) Map illustrating the sensitivity of the focal mechanism determinations to point source location. The major double couple solutions are plotted for point source locations at the epicenter and near the optimal centroid location. Note the steep southwest dipping plane obtained at the epicenter, in contrast with the shallow southwest dipping planes for solutions near the optimum centroid. (b) Contours of the residual error in the second-step inversion for the full moment tensor as a function of source centroid location for propagation model MPA, excitation structure RA, and global Q model DS. The minimum error gives the optimal centroid location for a particular combination of models, and is located 30 to 40 km in the updip direction from the epicentral location.
determine the focal mechanism, source time function, and spatial distribution of moment release on the fault.

Summary of Body Wave Results

The body wave mechanism we obtain for this event indicates oblique thrust motion ($\lambda = 50-80^\circ$) on a shallow dipping ($\delta = 15-20^\circ$), southeast striking ($\phi = 90-120^\circ$) fault plane (Figure 7). No change of mechanism during the rupture process is unambiguously resolved, although we cannot preclude some minor subevents having different mechanisms. The extent of the rupture defined by subevent directivity is quite small for an $M_w = 7.7$ event. The resolvable rupture area is no larger than 80 x 60 km$^2$, and most of the moment release occurred within an area of about 45 x 45 km$^2$. The rupture appears to have been quite smooth, propagating somewhat faster downdip than updip. Two pulses of high moment release several seconds apart are discernable. The first occurred approximately 15 km downdip of the hypocenter (subevents 2 and 3 in Figure 7) and the second ruptured an area 15 to 30 km updip (subevents 4, 5, and 6). The total duration of the body wave radiation was approximately 40 s, but the two significant episodes of moment release occurred in the first 20 s. The seismic moment determined from the body waves is between $1.2 \times 10^{20}$ and $2.0 \times 10^{20}$ N m. We prefer the higher of these estimates as it provides a superior fit to the P wave amplitudes. Below, we outline the procedure used for determining body wave source parameters, and then present the results of these inversions.

Procedure

Because of initial indications that the rupture took place on a shallow listric fault [Montero et al., 1991b], i.e. a fault plane with a dip that increases towards the surface, we chose to use the Kikuchi and Kanamori [1982, 1986, 1991] body wave inversion technique which allows us to solve for spatial characteristics of the rupture and possible changes in mechanism. Kikuchi and Kanamori's method solves for the rupture process in terms of a series of subevents. Given a fixed focal mechanism and trapezoidal time function, the method inverts for moment, onset time relative to a reference time (here taken to be the P wave onset times of the seismograms), and location on the fault plane relative to the hypocenter (for which we must assume an absolute depth) for each subevent. With the most recent inversion technique of Kikuchi and Kanamori [1991], it is also possible to solve for separate moment tensors for each subevent. At each iteration, parameters of one extra subevent are determined by minimizing the residual difference between observed and synthetic seismograms. There are trade-offs between timing, location and focal mechanism (if included in the inversion) of each subevent [Kikuchi and Kanamori, 1991; Young et al., 1989]. These trade-offs have to be evaluated carefully when assessing the significance of resulting solutions.

Synthetic seismograms that include direct P and S arrivals as well as surface reflected phases Pp, Sp, and S are calculated using geometric spreading factors (calculated from PREM) for mantle propagation, frequency independent constant $t^*$ operators for attenuation ($t^*_p = 1$ s; $t^*_s = 5$ s), and horizontally layered source and receiver structures. We use a simple continental model for both source and receiver structures, with a P wave velocity of 6.4 km/s in a 40-km-thick crust and 8.15 km/s in the mantle. Matsumoto et al. [1977] developed a velocity model for northern Costa Rica but there is no specific model for the source region of the Valle de la Estrella earthquake. We tried the Matsumoto et al. [1977] model as well as a model suggested by F. Tajima and M. Kikuchi (1983 and 1991 Costa Rica earthquakes: Modeling of source ruptures with time-variant fault mechanisms, submitted to Tectonophysics, 1992). In both cases we obtained an inferior fit to the data compared with the two-layered source structure. Given the lack of a more detailed velocity model that is appropriate for the source region we use the simple model in all our subsequent inversions. The values for P and S velocities for this model agree well with the averaged velocities for the more complicated source structures.

Each subevent is a point source with a trapezoidal source-time function to be located on a prescribed grid which represents the fault plane. We use a grid spacing of 15 km both along dip and along strike. The total extent of the grid is varied to ensure that its dimensions do not affect the solution. The final grid has five grid points along dip and seven along strike. Using subevent trapezoids with a rise time of 4 s and a total duration of 9 s gives the best overall match to the waveforms. The 9-s subevent trapezoid duration is short enough to model most details of the P waves and in combination with a $t^*_s$ of 5 s gives a reasonable fit to the SH waves, which are relatively broad.

Finally, the hypocentral depth ($h_0$), which is the reference point for the spatial distribution of the subevents, and the orientation of the fault plane must be specified. Since these parameters are not well constrained by independent knowledge for this earthquake, we solve for $h_0$ and the focal mechanism using an error minimization strategy. After these parameters are constrained, the source-time function and the spatial rupture pattern are solved for.

Representative seismograms of our data set are shown in Figure 8 (the complete data set is shown with our final fit in Figure 11). The traces are deconvolved by the instrument responses to obtain ground displacements. For an accurate assessment of the spatial distribution of moment release, determination of accurate onset times is important but is somewhat difficult due to the emergent character of many of the arrivals, especially the SH waves (Figure 8). All onset times are determined from short-period channels if available or picked to be consistent with expected arrival times for the NEIC origin parameters. We have nonuniform azimuthal coverage, with few stations located at southern azimuths and many stations at northwestern azimuths (Figure 8). To reduce bias from this irregular station distribution, the northwestern stations PAS, COL, ANMO, and COR are given half the weight of the other stations in the waveform inversion. Furthermore, all traces are weighted by the inverse of their maximum amplitude. This does not change the actual amplitudes, but in the calculation of the misfit it deemphasizes records with higher amplitudes, especially those of the SH waves, so that they do not dominate the inversion.

It is clear from Figure 8 that only one nodal plane of the focal mechanism is well-constrained by the first motions at these stations. The strike of the other plane can vary between about 50° and 190°, i.e. dipping between east-southeast and west, and still satisfy the P wave polarities. Thus, we use the waveforms to determine the focal mechanism. We note that the first motion solution from local seismic stations [Montero et al., 1991a] suggests a thrust mechanism with close to 45° dip, which is not consistent with our teleseismic first motions. Given the poor constraint on the velocity structure
in the source region it is likely that this reflects errors in the takeoff angles to the local stations; however, it is possible that the very high frequency radiation had a slightly different initial motion than what is observed teleseismically.

**Results**

We first perform point source inversions to constrain the hypocentral depth and focal mechanism. We invert for a range of mechanisms for a variety of fixed depths and estimate the total misfit, which is given by $\Delta = \sum (\text{obs}_j - \text{syn}_j)^2 / \sum \text{obs}_j^2$ after six iterations (i.e., with a series of six subevents for the synthetics). The summation is over the weighted P and SH waves for all stations. Unconstrained point source moment tensor inversions result in thrust mechanisms, but fail to precisely match the emergent P wave first motions, so we choose to constrain our source to a double couple mechanism. For our range of mechanisms, we keep the steeply dipping nodal plane fixed, consistent with the first motions, and vary the orientation of the fault plane. The mechanism is assumed to be constant throughout the rupture process.

A point source approximation gives a reasonable fit to the data, which can be explained by the relatively confined area of the high-frequency moment release in the final finite source result, and the insensitivity of the very smooth SH waves to details within this small area. Figure 9 shows the misfit between observed and synthetic seismograms as a function of focal mechanism for different values of $h_0$. The mechanism is expressed in terms of the strike of the fault plane. The best overall fit to the waveforms is obtained for a mechanism with $\phi = 111^\circ$, $\delta = 16^\circ$, $\lambda = 72^\circ$, and $h_0 = 5$ to 10 km. However, misfit curves for different $h_0$ are generally broad, and the data can be fit nearly as well with fault planes striking between 90$^\circ$ and 120$^\circ$. The misfit for an incorrect choice of hypocentral depth increases because of subtle interference effects, especially near the onset of the P waveforms. Figure 9 shows how well the onset of the P waves is fit for $h_0 = 10$ km. The fit for $h_0 = 5$ km is equally good, but it degrades both for shallower depths and for depths greater than 10 km. Only P waves are shown since they are most sensitive to these details of the rupture process. In the smooth SH waveforms (Figure 8), the small features of the first pulse of ground motion can not be distinguished separately.

If we allow for spatial extent of the source, either as a one-dimensional line source along dip of the fault plane or a two-dimensional grid on the fault plane, we find that the misfit for $h_0 = 10$ km is reduced more than for $h_0 = 5$ km (Figure 10). The relatively good fit for the 5 km deep point source reflects the fact that the major moment release is located updip of the hypocenter in the final finite source solution. A 10-km-deep hypocenter gives a better overall fit for a source with finite spatial dimension and still provides good fits to the onsets of the P waves.
We invert for spatial distribution of moment release for both P and SH waves as well as for only P waves. Allowing for spatial propagation of the rupture with the hypocenter at 10 km depth improves the fit only slightly (6% for the best fit mechanism and six iterations) relative to a point source when both P and SH waves are used. The P waves alone, however, give a 25% improvement in fit when a planar source is used instead of a point source (Figure 10). The lack of constraint for spatial distribution inversions provided by the combined P and SH inversion can be attributed to several factors. First, the onset times are determined more accurately for the P waves than for the SH waves. Second, propagation effects influence the SH waves more strongly than the P waves, such as attenuation which filters out higher frequencies. The very smooth SH waves for this event do a poor job in constraining the spatial distribution of moment release, where in fact the less attenuated P waves alone can constrain the spatial extent quite well. In spite of the small reduction of misfit in the combined inversion, the general spatial pattern obtained from this inversion agrees very well with what we obtain from the P waves alone. The detailed times and positions of the individual subevents from the P wave inversion of course differ from the details of the solution from the joint inversion, but the general interpretation of the rupture process is the same. In both inversions the updip and downdip patches of moment release, as shown in Figure 7, are well located in depth, but the along strike distribution cannot be very well constrained. Details of the solution from the joint inversion are shown by the numbered subevents on the source time function and the fault plane solution (Figures 7a and 7c). We prefer, however, to emphasize a more general interpretation of the results which summarizes the features that we think are well resolved by the inversions. We therefore emphasize the slip distribution along dip, although the rupture does have some poorly resolved extent along strike.

Allowing for the spatial distribution does not further constrain the mechanism. The best overall fit is obtained for a mechanism with $\phi = 102^\circ$, $\delta = 17^\circ$, and $\lambda = 63^\circ$. The fit between data and synthetics for this final body wave inversion is good (Figure 11). Inverting the P waves alone results in a higher moment than for the joint P and SH inversions, $M_o = 2.0 \times 10^{20}$ N m instead of $1.2 \times 10^{20}$ N m. This is probably a better estimate of the total body wave moment since it gives the best fit to the amplitudes of the waves. The total duration of the rupture was around 40 s, but most of the energy was released in the first 20 s, which is also the portion of the rupture for which the location could be well constrained. This energy was confined within an area of about $45 \times 45$ km$^2$ that had two pulses of high moment release, one updip and one downdip from the hypocenter, several seconds apart. A second, smaller, peak of moment release between 20 and 40 s is resolved but cannot be well located on the fault plane.

Because of indications of a possible change in dip during the rupture [Monteiro et al., 1991b] or the presence of minor strike slip motion (F. Tajima and M. Kikuchi, submitted manuscript, 1992) (or consistent with aftershock mechanisms northwest of the rupture plane), we also allowed for a variable mechanism solution, where each subevent may have a different
HYPOCENTRAL DEPTH AND FIT FOR A POINT SOURCE

Fig. 9. (a) Misfit ($\Delta$) between observed and synthetic seismograms for point sources at different depths, $h_0$. The misfit is plotted as a function of the fault mechanism, characterized by the strike of the fault plane. Focal spheres show the range in mechanisms (same as shown by the dashed lines in Figure 8) and the mechanism that gives the minimum in misfit. (b) Fit to the onset of several representative P wave seismograms when the point source is placed at 10 km depth. Solid lines are the observed waveforms; dashed lines are the synthetic seismograms.

MISFIT FOR POINT, LINE OR PLANE SOURCE

COMBINED P AND SH

P-WAVES ONLY

Fig. 10. Similar to Figure 9, but for different degrees of spatial distribution of the subevents: a point source, a line distribution along the dip of the fault plane, and a two-dimensional planar distribution along strike and dip. Misfits are shown for inversions with both P and SH waves as well as using only P waves, with hypocentral depths of 5 km and 10 km.

orientation. Solving for different subevent moment tensors adds five parameters to be determined in each step of the inversion. The trade-off between the distribution of moment release in time and space that we have discussed above is exacerbated to a trade-off between time, location, and focal mechanism for each subevent. We perform this inversion along a grid having the orientation of our preferred fault plane ($\phi = 102^\circ$, $\delta = 17^\circ$) with $h_0 = 10$ km. To test the sensitivity of the focal mechanism determinations to the assumed source structure, we ran inversions using several different source velocity models [Matumoto et al., 1977; F. Tajima and M. Kikuchi, submitted manuscript, 1992] including the layer over a half-space velocity model described earlier. We also vary the spatial distribution of the grid of subevent locations. The composite source function for any of these inversions has a moment between $1.0$ and $1.4 \times 10^{20}$ N m, and the summed mechanisms are within the range determined by the fixed mechanism inversions. This indicates that the overall faulting is consistent with the first motions, thereby justifying our constraint of the steeply dipping nodal plane in the fixed mechanism inversions.
The subevent parameters found in the first two or three iterations (which have the largest moments) are always very close to the parameters that yielded the smallest error in the fixed mechanism inversions. The major subevents also have spatiotemporal distributions of moment release that correspond closely to the earliest subevents found with the fixed mechanism finite source inversions. Minor subevents are much more unstable and their significance is hard to assess. Focal mechanisms determined for minor subevents range from pure thrust, to oblique thrust, to strike-slip mechanisms with either right or left-lateral motion. These minor events are often located updip, but variation in mechanism can suppress the evidence for updip rupture. Varying the source velocity structure and the data weighting scheme results in very different parameters for the later subevents, demonstrating large trade-offs between the many
parameters. The slight improvement in the fit to the signals obtained for the variable mechanism inversions does not offset the increase in degrees of freedom. Thus, we find no compelling indication of a significant change in mechanism during the rupture process, in contrast to the results of F. Tajima and M. Kikuchi, submitted manuscript, (1992). Given the instability of the variable mechanism inversions, we prefer the fixed mechanism body wave solution, where we solve only for the spatial and temporal distribution of moment release.

**DISCUSSION**

**Comparison of Body Wave, Surface Wave, and Geodetic Results**

The agreement between body and surface wave results for the 1991 Valle de la Estrella, Costa Rica, earthquake is quite good. The body wave solution gives more details of the rupture process than the long-period surface waves, while the overall mechanism and total duration are better resolved by the long-period signals. Thus, the results from both wave types complement and reinforce one another. The seismic waves reveal a shallow dipping thrust plane, with some oblique slip. The surface waves constrain the strike and rake better than the body waves, which is probably a result of the limited station distribution in the body wave inversion. On the other hand, the moment and dip obtained from the surface waves have large uncertainties. The dip determined from body waves appears to be slightly better constrained, but the body wave moment (2.0 x 10^20 N m) is below the lower bound of the estimated surface wave moment (3.8 ± 1.5 x 10^20 N m). Constraining the dip in the surface wave inversion to have the shallower dip of the body wave solution results in an increase in moment, so it is likely that the body wave moment estimate is biased low, probably a result of the limited bandwidth of the deconvolution procedure. Some smooth long-duration slip can be imposed on the body wave source function without significantly degrading the fit to the data. The updip location of the surface wave centroid is compatible with the concentration of moment release updip found for the body waves. The half duration (centroid time of the moment release) of the surface wave source function is 20 ± 3 s, while the body wave centroid time is approximately 16 s, demonstrating the general compatibility of the duration estimates. The hypocentral depth determined from the body waves is shallower than the surface wave centroid depth, but there are uncertainties in both estimates due to the lack of a priori knowledge of the source velocity structure.

If we use the source orientation determined from the teleseismic waves and find a finite fault model with uniform displacement which fits the coastal uplift pattern from Plafker and Ward [1992], we find good agreement, as shown in Figure 12. Also included in Figure 12 is a comparison of this model with the GPS trilateration data from Lundgren et al., [1993]. The oblique component of the seismic mechanism improves the agreement with the GPS and uplift observations relative to the pure thrust geometries considered by Plafker and Ward [1992]. The azimuth of the horizontal deformation near Limón is fairly well predicted, especially allowing for the 17° uncertainty in rake. The surface wave solution predicts amplitude of the Limón GPS data slightly better. The uniform slip model gives a moment intermediate to our body wave and surface wave estimates. A nonuniform slip model, with concentrated slip near Limón, as suggested by the body waves may be able to explain the peak of uplift which reached a maximum of about 1.5 m near the town of Limón [Plafker and Ward, 1992; Plafker et al., 1991; de Obaldia et al., 1991a,b], although some local near-surface complexity may have caused some of the localized uplift. Concentration of slip below Limón may also explain the magnitude of the horizontal motions, which are under predicted by 30% by the uniform slip model. The existing data, however, do not permit a stable inversion for a nonuniform slip model. The dimensions of the seismically determined fault plane are very similar to those found from this dislocation model as well as those from Plafker and Ward [1992] and agree with the aftershock data if we allow for separate fault activity west of the epicenter (Figure 2) [Güependel et. al., 1991; Montero et al., 1991a,b; Schwartz and Protti, 1991].

**Tectonics, Aftershocks, and the Termination of the NPDB**

The precise plate boundaries between the Cocos, Caribbean, Nazca, and South American plates in this region are not very clear. Though the NPDB has generally been recognized as a zone of active underthrusting and possibly subduction [e.g., Silver et al., 1990; Wolters, 1986; Wadge and Burke, 1983], the role it has been assigned in the regional tectonics varies widely. Suggestions for its role range from being a major plate boundary separating the Caribbean and the Nazca plate [Bowin, 1976] to being relatively unimportant compared to the more traditional plate boundaries which lie along the MAT for the Cocos-Caribbean boundary and south of Panama along a supposed left-lateral transform fault for the Nazca-Caribbean boundary [Molnar and Sykes, 1969; Jordan, 1975; DeMets et al., 1990]. Intermediate interpretations partition the Cocos-Caribbean and Nazca-Caribbean convergent motions between the NPDB and the other boundaries on the west coast of Costa Rica and the south coast of Panama. The southern central American region in this context is viewed as one or a few separate blocks [Adamek et al., 1988; Silver et al., 1990; Vergara-Muñoz, 1988], or as a region of diffuse deformation [Pennington, 1981; Mann and Burke, 1984].

Very little seismic activity has in fact been registered along the northern Nazca boundary south of Panama, where there is no strong seismological or geological evidence for a major strike slip fault. There is evidence for a buried trench south of Panama [Lowe, 1978], which refraction data and background seismic activity suggest may still be active [Silver et al., 1990]. In contrast, the NPDB seems quite active considering the 1991 Valle de la Estrella event along with the large historic earthquakes in the NPDB in 1882 and 1916 (Figure 1). The direction of slip during the 1991 earthquake is closer to the direction of Nazca-Caribbean convergence (74° at the epicenter, calculated from DeMets et al. [1990]) than to the Cocos-Caribbean convergence (23° at the epicenter, calculated from DeMets et al. [1990]) (Figure 13). The direction of slip indicates that the motion along the NPDB may be governed for a large part by Nazca-Caribbean relative motion. This is in contrast with assumptions by Plafker and Ward [1992] and G. Suarez et al. (unpublished manuscript, 1992), that the motion along the NPDB is part of the Cocos-Caribbean convergence. The uncertainties in the exact
Fig. 12. (Top) Comparison of predicted horizontal surface motions for a uniform slip model using the geometry found by the seismic wave analysis with the observed motions from GPS analysis (bold arrows near Limón and lower right hand corner of the fault plane) reported by Lundgren et al. [1992, 1993]. Letters L, C, and B on the map correspond to the places Limón, Cahuita, and Bocas del Toro, respectively. The solid star represents the NEIC epicentral location and the dashed box the fault plane. The computed (solid line) and observed (triangles) [Plafker and Ward, 1992] coastal uplift are compared in the bottom figures. The fit to the uplift data determined the parameters of the uniform slip model, including the average slip of 2.8 m, seismic moment of $2.6 \times 10^{20}$ N m and the fault plane dimensions (58x48 km², and a 3-km minimum depth).
direction of slip for 1991 and the earlier events however, leave open the possibility of a separate Panama Block, as proposed by several workers [Adamek et al., 1988; Silver et al., 1990; Vergara-Muñoz, 1988], which moves independently of the major plates. The existence of this microplate might be a transitional situation, as motion that previously was taken up by the southern margin of Panama is transferred to the already more active northern thrust belt. The southern boundary of the Panama block would be formed by what appears to be a thrust and strike-slip zone [Silver et al., 1990]. On the east it would be bounded by a shear zone in northwestern Colombia, which is marked by diffuse seismicity and several strike-slip faults. There is no clear crosscutting surface structure that could define the western boundary of the Panama Block. This may be either because it is still developing or because it is a hinge region with diffuse seismicity and strike-slip faulting that characterizes central Costa Rica.

The aftershock activity (Figure 2) [Montero et al., 1991a; Schwartz and Protti, 1991] is located mostly on land, where thrust faults have previously been identified as the onshore extension of the NPDB [e.g., Ponce and Case, 1987], with little activity in the inferred part [Plafker and Ward, 1992] of the thrust belt off the coast of Limón. The deformation belt which is very wide near Colombia narrows toward the Costa Rican coast [Silver et al., 1990], suggesting that the onshore extension of the NPDB represents a hinge of the relative motion between Panama and the Caribbean plate. This hinge may be the termination of the thrust belt. Little aftershock activity occurred to the north of the mainshock rupture area [Montero et al., 1991a; Schwartz and Protti, 1991] supporting termination of the thrust belt here, although some offshore seismic activity does extend northward along the Caribbean coast of Costa Rica (note the April 27, 1991, event shown in Figure 2). Many aftershocks form a northeast-southwest trending diffuse zone on the northwestern end of the rupture (Figure 2), with left-lateral strike-slip mechanisms [Dziewonski et al., 1992; Montero et al., 1991a; Fan et al., 1992]. This supports the hypothesis that the NPDB terminates with a diffuse set of left-lateral strike-slip faults [Ponce and Case, 1987], perhaps traversing Costa Rica and merging into the Middle American Trench [Jacob et al., 1991].

An alternative suggestion for the termination of the NPDB is that the thrust belt extends as far north as the Hess Escarpment [Escalante, 1990; Plafker and Ward, 1992]. Plafker and Ward [1992] suggest that the thrust belt most likely comes onshore beneath thick Quaternary deposits north of Limón and terminates in northern Costa Rica against the onshore extension of the Hess Escarpment. Onshore geologic data do not indicate the thrust belt continues north of Limón [Ponce and Case, 1987], but offshore thrusting may still be occurring. Although we cannot preclude this hypothesis, we feel the evidence discussed above suggests that the termination of the NPDB occurs in a diffuse zone of strike slip faults. Thus, the location of the Valle de la Estrella earthquake mostly like represents the northwest termination of the NPDB.

**Comparison With Other Back Arc Thrusting Events**

Subduction accompanied by back arc thrusting is quite rare. Two regions with some resemblance to what is taking place in Panama are the Sunda Arc and the Japan Sea. Along the Sunda Arc, proximity of the Australian continent to the subduction zone at the Timor Trough may cause convergent motion to be transferred to the Flores Thrust in the back arc [McCaffrey and Nábělek, 1984]. Subduction of the Cocos Ridge and the Panama Fracture Zone in the Middle American Trench may cause a similar effect near Costa Rica. In the Japan Sea, two major back arc thrusting earthquakes occurred in 1964 and 1983, the Niigata (Mw = 7.6) and Akita-Oki (Mw = 7.8) events, respectively. These two events have been suggested to represent subduction between the Eurasian and North American plates, where the convergent motion is transferred from the subduction zone east of Japan to the back arc thrust faults [e.g., Kanamori and Astiz, 1985]. Cook et al. [1986] suggested that the area between these two major fault zones actually constitutes a separate Okhotsk plate, which makes

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**Fig. 13.** Final interpretation of our body and surface wave solutions projected onto a map of the epicentral region. The dashed box represents the fault plane area in the finite source inversion, where the star in the middle coincides with NEIC epicenter. The cross-hatched areas represent the regions where most of the moment was released. The solid arrow gives the direction of slip. Areas where coastal uplift was observed [Plafker and Ward, 1992] are marked with dark stippled triangles. Our results are consistent with this observed uplift and with the orientation of the observed thrust faults in the area that have been associated with the NPDB. The upper inset shows the relative plate motion directions for the Cocos and Caribbean plates (COC-CAR) and the Nazca and Caribbean plates (NAZ-CAR) for the NUVEL model [DeMets et al., 1990] calculated near the 1991 source location. The COC-CAR vector has a magnitude of 9.1 cm/yr, the Nazca-Caribbean convergence is 4.9 cm/yr. The slip direction of the 1991 event (not scaled) is similar to the NAZ-CAR relative motion. The small map at the top shows the various plates in the region, including the conjectured Panama Block, bounded by shear zones marked by dark stippled areas. The tectonic regime is discussed in the text.
estimation of the back arc hazard as difficult as it is along the NPDB.

We compare the Valle de la Estrella event to the well-documented Niigata and Akita-Oki earthquakes by considering the stress drops for these three events. Approximating the ruptures by circular faults with area equal to the main slip zones gives stress drops of 6.5, 5.6, and 5.4 MPa for the Niigata, Akita-Oki, and Valle de la Estrella earthquakes, respectively (parameters for the Japan earthquakes are from Abe [1975], Satake and Abe [1983], and Satake [1985]; for the Costa Rica earthquake we use the body wave results $M_o = 2.0 \times 10^{20}$ N m and rupture area is $45 \times 45$ km$^2$, which give a displacement of $3.3$ m for a rigidity of $3 \times 10^{10}$ Pa). The stress drop for the Valle de la Estrella event is thus similar to those for the Niigata and Akita-Oki earthquakes. Using the larger surface wave moment and a larger fault area for the 1991 event gives a comparable stress drop. The stress drops for these back arc events are more consistent with stress drops characteristic of interplate events than intraplate events [Kanamori and Anderson, 1975], suggesting that all of these earthquakes represent motion on active plate boundaries.

CONCLUSIONS

Body and surface wave inversions give consistent results for the rupture process of the April 22, 1991, Valle de la Estrella, Costa Rica, earthquake. The event was a back arc thrusting event on a shallow, southwest dipping plane with a small oblique component. The rupture appears to be relatively simple over a broad range of frequencies. We find that the rupture began at a shallow depth, 10 to 20 km and propagated from the hypocenter both updip and downdip along the fault plane. Most of the moment was released in two pulses only a few seconds apart. The first pulse ruptured a patch at about 15 km depth; the second one ruptured a larger patch near 5 km depth. The shallow slip is consistent with the observed surface deformation and tsunami. The duration of the rupture was about 40 s and the total moment is $3.8 \times 10^{20}$ N m.

The earthquake appears to have ruptured the westernmost end of the North Panama Deformed Belt. This thrust belt has in the past generated major, but infrequent earthquakes from its eastern end in the Colombia basin westward to the Limon Basin. The NPDB takes up a significant portion of the complicated convergent and rotational motions that characterize the tectonics of the region, and may be an active plate boundary at the north end of the Panama Block.

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S.D.B. Goes, T. Lay, S.Y. Schwartz, and A.A. Velasco, Institute of Tectonics, University of California, Santa Cruz, CA 95064. (Received May 15, 1992; revised November 16, 1992; accepted December 29, 1992.)