Chapter 4

Seismogenic Zone Structure of the Southern Middle America Trench, Costa Rica

4.1 Abstract

This study provides structural constraints on the seismogenic zone of the Middle America Trench offshore central Costa Rica and insights into the physical and mechanical characteristics controlling seismogenesis. I have located ~300 events that occurred following the M_w 6.9 August 20, 1999, Quepos, Costa Rica underthrusting earthquake using 3D velocity models and arrival data recorded by the Osa Peninsula CRSEIZE network of land and ocean bottom seismometers. Aftershock locations define a plane dipping at 19° that marks the interface between the Cocos Plate and the Panama Block. The majority of aftershocks occur below 10 km and above 30 km depth below sea level, corresponding to 30-35 km and 95 km from the trench axis respectively. Relative event relocation produces a seismicity pattern similar to that obtained using absolute locations, increasing confidence in the geometry of the seismogenic zone. The aftershock locations spatially correlate with the downdip extension of the oceanic Quepos Plateau and reflect the structure of the mainshock rupture asperity. This strengthens an earlier argument that the 1999 Quepos earthquake ruptured specific bathymetric highs on the downgoing plate. The subduction of this highly disrupted seafloor has established a set of conditions that presently limit the seismogenic zone to be between 10-35 km below sea level.

4.2 Introduction

On August 20, 1999, a M_w 6.9 underthrusting earthquake occurred offshore the city of Quepos in central Costa Rica (Figure 4.1). The earthquake occurred at the southern edge of the morphologically defined seamount-dominated segment of the Costa Rica margin [von Huene et al., 1995] along the downdip extension of the incoming Quepos Plateau, a highly disrupted string of seamounts and bathymetrically high crust (Figure 4.1). The event was preceded by a M_w 5.5 foreshock on 8/10/99 and generated much aftershock activity. Best depth estimates using the Observatorio Vulcanológico y Sismológico de Costa Rica (OVSICORI) epicenter place the event at 21±4 km [Bilek et al., 2003], near the plate interface as defined by available refraction information [Stavenhagen et al., 1998] and consistent with the Harvard CMT thrust focal mechanism. The moment release history consisted of two main subevents with a total rupture length equal to the width of the Quepos Plateau [Bilek et al., 2003]. Bilek et al. [2003] proposed that the Quepos event represented rupture of specific topographic highs within the Quepos Plateau seamount chain acting as asperities at depth and that the size of these asperities limited the mainshock rupture extent. The magnitude of the mainshock is consistent with the largest magnitudes (M_w 7.0) recorded along the central Costa Rica margin [Protti et al., 1994, 1995a], and hence, the aftershock sequence was assumed to have ruptured much of the interplate seismogenic zone.
The Quepos earthquake occurred 19 days prior to scheduled station deployment of the Osa Peninsula CRSEIZE seismic experiment, originally designed to be placed on and offshore the Osa Peninsula. The land and ocean bottom seismic stations were hence relocated to better record the aftershock sequence of the Quepos earthquake and provide a more favorable network geometry for aftershock relocation. In this chapter I present aftershock locations of the Quepos underthrusting earthquake and use these small magnitude events to define the geometry of the seismogenic plate interface northwest of the Osa Peninsula. The goals of this study include: 1) determining a 3D velocity $P$-wave model for the southern Costa Rica margin; 2) locating small magnitude ($1.0 \leq M_L \leq 4.0$) earthquakes within this velocity model; 3) exploring the trade-off between absolute and relative earthquake locations in a heterogeneous subduction environment; 4) determining the updip and downdip extent of seismogenic zone earthquakes offshore central Costa Rica; 5) recognizing patterns in the seismicity defining the subducting plate interface; and 6) correlating seismicity with existing thermal and mechanical models for seismogenic zones.

4.3 Earthquake Location Methods

The Osa Experiment recorded over 1000 local, regional, and teleseismic earthquakes including aftershocks of the 1999 Quepos earthquake, Panama Block, Caribbean, and Cocos intraplate earthquakes, and activity along the outer rise (Figure 4.1). The Osa seismic array covered ~1600 km$^2$ and extended from the central Costa Rica coast and northwest Osa Peninsula to seaward of the Middle America Trench (MAT) (Figure 4.2). Aftershocks and oceanic intraplate activity dominate the dataset with ~600 events occurring within the boundaries of this study (Figure 4.2). I focus event relocations on those events within the station network coverage recorded by both land and ocean stations with a minimum of 10 arrivals, including both $P$- and $S$-waves. These earthquakes occurred between 9/24/1999 and 11/20/1999.

I relocate events using both absolute and relative earthquake location techniques to best resolve the aftershock pattern of the 1999 Quepos earthquake. High-resolution earthquake relocation requires either accurate $a$ $priori$ knowledge of velocity structure or a high quality dataset allowing for joint hypocenter-velocity model inversion. Velocity models are related to hypocentral parameters through the calculation of travel times, and there is a strong non-linear trade-off between velocity and hypocenter location [e.g., Crosson, 1976; Kissling, 1988; Thurber, 1992]. When network coverage or data resolution do not allow for joint inversion of travel times, $V_P$, $V_S$, or $V_P/V_S$ must be accounted for in other ways, either through layered 1D models or more complex 2D/3D models constrained by velocity information from seismic refraction, reflections, and teleseismic modeling. For the Osa Experiment, I construct an $a$ $priori$ 3D $P$-wave velocity model using velocity information from local reflection and refraction studies (Figure 4.2) for use in the non-linear, grid-searching location program QUAKE3D [Nelson and Vidale, 1990]. Evidence from Chile and Japan [Husen et al., 1999; Shinohara et al., 1999] suggests $V_P/V_S$ varies considerably in subduction zones. Refraction information does not provide independent $S$-wave velocities, and the effects of modeling a variable ratio with a constant ratio will have to be assessed. In order to test
the geometry of the absolute locations along the plate interface and explore spatial clustering within the aftershock sequence. I also relocate hypocenters using hypoDD, a relative relocation program that utilizes an arrival time differencing technique within a 1D P-wave velocity model [Waldhauser and Ellsworth, 2000].

4.3.1 3D Velocity Model and Location Technique

A 3D P-wave velocity model is developed utilizing a combination of 2D refraction data and surface geology for geometric and velocity constraints (Figure 4.2 and 3.3). Numerous published and unpublished refraction-based P-wave velocity data exist for offshore central and southern Costa Rica [Ye et al., 1996; Stavenhagen et al., 1998; Walther, 2003] (Figure 4.2), providing absolute velocity values, velocity gradients, and depths of sharp velocity contrasts. Stavenhagen et al. [1998] provide velocity data through the region of aftershock activity, and data from Walther [2003] provide velocities along the crest of the Cocos Ridge and perpendicular to the strike of the ridge near the MAT. Where velocity information does not exist, velocity layers are linearly interpolated along strike. I model the Osa Peninsula as a rectangular box of higher velocity material as surface geology indicates the peninsula is primarily composed of ophiolitic material [Gardner et al., 1992]. Refraction information from Ye et al. [1996] forms the north boundary of the model and allows for accurate representation of the steepening slab from the Osa to the Nicoya Peninsula. The resultant 3D velocity model is 176 km x 226 km x 80 km, with a grid spacing of 1 km. It encompasses the Osa Peninsula, the central seamount-dominated oceanic crust, and the deformed forearc from the trench to the central volcanic chain (Figure 4.3). The model is composed of constant velocity polygons, and P-wave velocity gradients have been accounted for within the limitations of the design program. I use the $V_p/V_S$ average value of 1.78 to calculate a corresponding S-wave velocity model, consistent with the 1D velocity model inversion findings for this region [Chapter 3] (Figure 4.4).

The QUAKE3D suite of programs combines a finite-difference travel time calculator [Vidale, 1988; Vidale, 1990; Hole and Zelt, 1995] with a grid-search earthquake location algorithm [Nelson and Vidale, 1990] to locate events within an arbitrarily complex 3D velocity volume. The finite-difference travel time calculator solves for P- and S-wave travel times from each station to all points in a pre-defined grid (x, y, and z) by solving the 3D eikonal equation for raytracing,

$$
\left( \frac{\partial t_{\text{calc}}}{\partial x} \right)^2 + \left( \frac{\partial t_{\text{calc}}}{\partial y} \right)^2 + \left( \frac{\partial t_{\text{calc}}}{\partial z} \right)^2 = s^2(x, y, z)
$$

where $t_{\text{calc}}$ is the calculated travel time from source to receiver and $s$ is inverse velocity or slowness, here held fixed [Vidale, 1990]. The finite difference approach has many advantages over traditional raytracing: 1) it automatically follows the first arrival and ignores multipathed arrivals; 2) it follows diffractions through shadow zones; 3) it solves for any number of points within a volume; and 4) it inherently addresses curved wavefronts [Vidale, 1990]. Algorithm improvements made by Hole and Zelt [1995] increase the processing speed by solving for multiple sources simultaneously and by better accommodating sharp velocity contrasts and resultant arrivals from reflections.
The earthquake location algorithm finds those grid points that minimize arrival time residuals for all station-arrival pairs using either L1 or L2 norm criterion and then interpolates between these grid points to find the local minimum residual [Nelson and Vidale, 1990]. The L1 norm criterion minimizes the sum of the absolute values of travel-time residuals
\[ r = \frac{1}{n} \sum_{i=1}^{n} |t_{\text{obs}} - t_{\text{calc}} - t_{\text{org}}| \]
while the L2 norm criterion minimizes the mean square travel time residual
\[ r^2 = \frac{1}{n} \sum_{i=1}^{n} (t_{\text{obs}} - t_{\text{calc}} - t_{\text{org}})^2 \]

where \( r \) is the event residual, \( n \) the number of receivers, and \( t_{\text{c}} \) the observed travel time, calculated travel time, and earthquake origin time [Nelson and Vidale, 1990]. Comparison of the two methods has shown that the L1 norm method is less influenced by outliers in a dataset and hence is better for sparse station coverage [Prugger and Gendzwill, 1988; Nelson and Vidale, 1990]. Theoretically, minimum location error using either method should be 0.1-1 km based on the 1 km grid spacing of the velocity model, and \( P, S \), or \( S-P \) travel times, or any combination of these, can be used in the location procedure. I expect higher errors however due to the large gap in station coverage, velocity model uncertainty, and reading errors.

### 4.3.2 Relative Relocation Technique

I compute relative locations using hypoDD (version 1.0), an arrival time differencing location algorithm that solves for the relative locations of event pairs within a closely spaced cluster [Waldhauser and Ellsworth, 2000]. HypoDD uses raytracing to calculate travel times within a layered 1D velocity model similar to the technique used by VELEST (see discussion in Chapter 3). The double-difference technique solves for the relative location of two spatially related events, \( i \) and \( j \), recorded at a common station, \( k \), following a linear matrix solution,

\[ \frac{\partial t_{k}^{i}}{\partial m} \Delta m^{i} - \frac{\partial t_{k}^{i}}{\partial m} \Delta m^{j} = d_{k}^{ij} \left( t_{k}^{i} - t_{k}^{j} \right)_{\text{obs}} - \left( t_{k}^{i} - t_{k}^{j} \right)_{\text{calc}} \]

where \( t \) is travel time, \( \Delta m \) represents perturbations to the four hypocentral parameters (latitude, longitude, depth, and origin time), and \( r \) represents the travel time residuals. For closely spaced events, raypaths from each event to each common station should be nearly identical, and differences in observed and predicted travel times should only reflect the relative difference in event location. Linking common arrivals and stations through a nearest neighbor approach creates event-station pairs. HypoDD iteratively minimizes arrival time residuals using weighted least squares methods, either a Singular Value Decomposition (SVD) approach or a conjugate gradient approach (LSQR). SVD performs well for small systems (100s not 1000s of events) and produces more accurate error estimates than the computationally efficient LSQR method [Waldhauser and Ellsworth, 2000]. I use the minimum 1D \( P \)-wave velocity model and a constant \( V_P/V_S \) of 1.78 discussed in Chapter 3 for relative earthquake locations (Figure 4.4).
4.4 Results

4.4.1 3D Velocity Model and Hypocenters

Absolute earthquake relocations using QUAKE3D are ranked based on arrival quality, greatest azimuthal $P$-wave separation (GAP), reported location errors, and event RMS residual. High quality hypocenters use $>10$ $P$- and $S$-wave arrivals combined, exhibit a GAP $\leq 180^\circ$, and have final RMS arrival time residuals within one standard deviation of the mean RMS arrival time residual for the dataset. For the Osa dataset, calculating earthquake locations using the L2 norm vs. the L1 norm criteria produces a greater number of high quality events, 399 vs. 381 respectively. Comparison of absolute locations using the L1 and L2 norm residual computation methods show a mean epicenter difference of 2.6 km and depth change of 3.4 km (Table 4.1), primarily resulting from a difference in the number and distribution of phases retained in the location determination. As both methods produce similar locations using the same average number of phase arrivals, 14.95 and 14.67 arrivals per event for L2 and L1 respectively, the L1 norm locations are favored due to a significant decrease in arrival time residual for each event. Average RMS residuals calculated using the L1 norm method (0.16 secs) are of the same order as, but slightly higher than, the median arrival time reading error for $P$- and $S$-wave data (0.11 and 0.13 secs respectively), as expected for well-constrained event locations. However, I assign the highest quality rating (A) only to those events that are retained in both the L1 and L2 datasets (Figure 4.2 and Appendix 7.1). I include a cluster of outer rise events (B quality) that occur near the outer boundary of the OBS array for interpretation purposes, although these events violate the maximum GAP criteria, and both errors and hypocenter locations are therefore poorly constrained, particularly in depth (Appendix 7.1). I use the highest quality events (267) for interpretation purposes, to determine earthquake location sensitivity to changes in velocity model and arrival information, and to estimate true location errors.

In order to explore the effect of the a priori velocity model on earthquake location within the Osa dataset, I compute event locations with QUAKE3D using 2D and several 3D model variants. The initial 3D model based on refraction information (referred to herein as OSA3D) contains a high level of structural detail that may affect event locations. Such details include a plate interface low velocity zone, a shallowing of the oceanic plate dip to the south, and velocity gradients within both the oceanic and continental plates (Figure 4.3 and 4.5). Most events within the Osa network locate within the model space most heavily influenced by the velocities and plate geometry (slab dip 17°) reported by Stavenhagen et al. [1998]. I test for sensitivity to the dip of the slab in the aftershock region by increasing slab dip throughout the model using the northernmost refraction information from Ye et al. [1996]. In effect I create a 2D model that eliminates the Osa Peninsula structure; the revised velocity model is herein referred to as OSA2D (Table 4.1). Average event distance and depth differences between locations through the 3D and 2D models are 2.9 km and 4.9 km respectively (Table 4.1). Removing velocity gradients from the initial 3D model, which increases velocity layer thickness and causes a decrease in resolution, or removing the low velocity zone between the oceanic and continental plates along the thrust interface results in average location changes of 1.6-4.8
km (Table 4.1) and negligible residual changes. Application of station corrections, calculated from the average residuals for each station for $P$- and $S$-wave arrivals, results in a mean epicentral distance change of 1.4 km and depth change of 1.7 km (Table 4.1). Average RMS residuals calculated for the 2D and 3D velocity model variants range between 0.18 to 0.20 seconds, greater than the OSA3D average RMS value of 0.16 seconds. Event locations are not sensitive to reasonable modifications of the 3D structure but are sensitive to the extension of a 1D or 2D velocity model to 3D. The hypocenters calculated using velocity model OSA3D retain a larger number of arrivals and possess the smallest event RMS residuals; this model is therefore preferred over 2D and 3D model variants and used in all further testing and interpretation.

Earthquake hypocenters and associated errors reflect arrival time type and quality as well as network coverage. I relocate events within the OSA3D model using only $P$-wave data to test location sensitivity to $S$-wave data; Gomberg et al. [1990] showed that inclusion of well-determined $S$-wave arrival data can significantly improve earthquake depth estimates while inclusion of poorly picked $S$-wave data can lead to significant location biases. Resulting locations have a mean epicentral distance difference of 4.4 km and a mean 4.7 km shift downward in depth (Table 4.1) and an average RMS residual of 0.21 secs. I test the importance of onshore vs. offshore data by the removal of land station data, though this creates a large gap in network coverage. Hypocenters move an average of 7.5 km in epicentral distance and 6.8 km in depth, illustrating the importance of station geometry for azimuthal coverage and of having both onshore and offshore network information for calculating local offshore earthquakes. Comparisons using land stations only were not carried out due to the geometry of the land network and the small number of events recorded by >10 $P$- and $S$-wave arrivals at land stations.

Final error estimates for absolute locations using the best 3D model (OSA3D) are conservatively 3 km in epicentral distance and 5 km in depth. These error estimates reflect the mean event error values reported for distance and depth within QUAKE3D and also incorporate the error estimates due to using L1 vs. L2 norm criteria, varying velocity model structure and dimension, and using arrival quality information. As such, these values likely represent a maximum error estimate. Errors of 3 km and 5 km reported here apply to those events located on or near the subducting plate interface; errors for events outside the station coverage, such as outer rise activity, would be greater.

The majority of events within the Osa dataset form a plane dipping at 19° interpreted as the seismogenic interface between the Cocos Plate and Panama Block (Figure 4.5a), in good agreement with dip values computed from seismic reflection, refraction, and geodetic estimates. Previous location studies using land network data were only able to resolve a cloud-like pattern of seismicity near the shallow plate interface [e.g., Protti et al., 1994, 1995a]. Scatter in the QUAKE3D dataset is asymmetric with more outliers occurring within the oceanic plate rather than the upper plate, and scatter increases outside the coverage of the station network, especially downdip. Oceanic intraplate earthquakes occur within the oceanic crust directly below the seismogenic zone and appear in all QUAKE3D iterations using a variety of velocity models. A few events locate deeply enough to occur within the oceanic mantle based on comparisons with
refraction information; these events are left uninterpreted due to the small number of events and potential for large depth error.

### 4.4.2 Relative Relocation of Hypocenters

I compute relative relocations of events using hypoDD [Waldhauser and Ellsworth, 2000] to further identify plate interface events and to explore the relative error estimates for well resolved interplate aftershocks. Location error within hypoDD is highly dependent on station geometry, data quality, and the maximum separation between events in a pair, where maximum separation is small compared to typical event-station distances [Waldhauser and Ellsworth, 2000]. The data quality and station geometry of the Osa experiment have been discussed previously (see Chapters 2 and 3). The mean event separation within the clustered Osa dataset is 5.5 km, well within the average station separation of ~20 km of the Osa array, and events within the aftershock region are linked by 14 arrivals, a value similar to the 14.67 average arrivals/event used for location by QUAKE3D. Locations are calculated within the minimum 1D velocity model using the LSQR and singular value decomposition (SVD) matrix inversion methods, and S-wave weighting of 0.50 relative to P-wave weighting. I retain the P- and S-wave qualities defined by the analyst and used within QUAKE3D. The LSQR method solves for relative relocations quickly while the SVD method produces more reliable error estimates for small datasets. The number of events located within the Osa station array is very small (<300), and I therefore focus discussion on the SVD results.

HypoDD does not solve for absolute location of hypocenters, and therefore cluster locations need to be shifted to visualize results in absolute space. I correct cluster locations based on the uniform shift between the hypoDD relocations and corresponding QUAKE3D absolute locations. This is an arbitrary correction, and therefore hypoDD results are not interpreted for absolute location. 138 of the 267 high quality QUAKE3D hypocenters are contained within the 224 relative relocations, and I use these locations to correct cluster centroids. The maximum cluster shift was 2.1 km both horizontally and in depth with the hypoDD clusters uniformly locating slightly deeper and landward from the QUAKE3D locations. Shifting cluster location does not affect the error calculations for individual events as reported errors are relative errors based on the relocations of events within a given cluster rather than the absolute location of the events. Most aftershocks lie within ~10 km of the interface shown in Figure 4.5b based on scatter within the relative relocation results. Mean relative errors for these events were 0.7 km epicentral distance and 0.8 km depth (see Appendix 7.1). Figure 4.5b shows the shifted hypoDD plate interface event locations with associated relative error bars plotted on a cross-section through the OSA3D velocity model.

### 4.5 Discussion

I define the up and downdip rupture limits of the 1999 Quepos earthquake using the statistical approach outlined in Pacheco et al. [1993] and applied by Husen et al. [1999] to the 1995 Antofagasta aftershock sequence. QUAKE3D hypocenters with depths within 5 km of the low velocity layer defining the plate interface are interpreted as interplate earthquakes (boxed events in Figure 4.5a) and plotted by depth distribution.
using a bin size of 2.5 km (Figure 4.6). The depth distribution of this dataset is best fit by a double Gaussian, consistent with reported distributions for other subduction zones [Pacheco et al., 1993; Husen et al., 1999]. The 5th percentile of the depth distribution defines the updip limit of seismicity for the Osa dataset at 10 km depth, 30-35 km from the Middle America Trench, and the 95th percentile defines the downdip limit at 30 km depth, ~95 km from the trench. Use of the 5th and 95th percentiles accounts for location errors and incompleteness within the dataset [Pacheco et al., 1993]. Systematic focal mechanism determinations for these events are currently underway to further constrain the nature of interplate seismicity.

Characteristics of the 1999 Quepos mainshock rupture [Bilek et al., 2003] and its aftershock sequence appear to be strongly influenced by the morphology of the downgoing plate. Deformation of the margin and uplift of the forearc along central Costa Rica have led investigators to suggest that incoming seamounts reach seismogenic depths offshore Costa Rica [Protti et al., 1995b; von Huene et al., 1995; von Huene et al., 2000; Husen et al., 2002]. Analysis of the spatial and temporal patterns within the aftershock sequence reveals details of the structure of the subducted plate. The along-strike extent of the aftershock pattern coincides with the width of the mainshock rupture determined from waveform inversion [Bilek et al., 2003] and with the along-strike width of the Quepos Plateau (Figure 4.7). Temporal relationships within the aftershocks indicate a fine-scale structure to the mainshock rupture asperity at depth. Most large aftershocks occur within the first month of recording, corresponding to 30-60 days after the mainshock, and almost all events downdip of the mainshock and events on the outer rise occur at this time (Figure 4.8a). Later aftershocks, more than 60 days after the mainshock, generally occur updip of the mainshock and define two linear streaks that lie parallel to the incoming Quepos Plateau (Figure 4.8b). These parallel streaks have a spatial separation similar to bathymetric highs within the Quepos Plateau suggesting the morphology of the subducted Cocos Plate beneath the mainshock mimics that of the incoming oceanic plate. Therefore, aftershock relocations support the Bilek et al. [2003] interpretation that the 1999 Quepos earthquake ruptured topographic highs at depth.

If bathymetric highs within the Quepos Plateau act as rupture asperities, or areas of concentrated moment release, and if asperity size limits the extent of rupture, then the 1999 Quepos earthquake may not have ruptured the updip and downdip extent of the plate interface capable of stick slip behavior. Instead, the limiting conditions controlling the transition from stick-slip to stable sliding behavior may change over the seismic cycle, and the subduction of highly disrupted seafloor in the vicinity of the 1999 Quepos earthquake has established a set of conditions that presently limit the seismogenic zone to be between 10-35 km below sea level. In this scenario, different segments of plate boundaries in different stages of the earthquake cycle would display spatial variations in up and downdip limits of seismicity. Such along-strike variability in the updip limit of the seismogenic zone offshore Costa Rica is supported by the initial shallow location and depth (10-15 km) of a $M_w$ 6.4 June 2002 underthrusting earthquake that occurred to the southeast of the 1999 Quepos aftershock area (Figure 4.7). Variability in the updip limit of interplate seismicity was also reported in northern Costa Rica where Newman et al. [2002] found evidence of an abrupt transition in the updip limit of microseismicity under
the Nicoya Peninsula. The evolving image of the seismogenic zone is one in which updip and downdip limits vary as a function of time within an earthquake cycle, and perhaps over longer periods, and these limits reflect temporal variations in critical parameters influencing the transition from stick-slip to stable sliding behavior. Although these critical parameters change with time, exactly what these conditions are, how long they will persist, and how they may change over time is unknown. It is therefore still instructive to compare the snapshot of the seismogenic zone illuminated by the aftershocks of the 1999 Quepos earthquake with various models that seek to describe the static depth extent of seismogenic zones.

Possible mechanisms controlling the transition from aseismic to seismic behavior along the updip limit of seismogenic zones include the mechanical backstop model [Byrne et al., 1988], temperature-controlled mineral transition models [Vrolijk, 1990; Hyndman et al., 1997], and combinations of mechanical and thermal controls [Hyndman et al., 1997; Moore and Saffer, 2001]. Byrne et al. [1988] suggested that unconsolidated, overpressured sediments support aseismic slip along the plate interface below the accretionary wedge while stronger, more coherent rocks in the crystalline upper plate backstop support higher levels of shear stress and therefore support stick-slip earthquakes. I disregard this model in south central Costa Rica because wedge sediments only extend to ~5 km depth here [Stavenhagen et al., 1998] while interplate seismicity begins near 10 km depth. Correlations between updip limits of seismicity and thermal modeling of the 100-150°C isotherms have been observed in Chile, Alaska, southwest Japan, and northern Costa Rica [Oleskevich et al., 1999; Newman et al., 2002], supporting the idea of a temperature influenced updip limit. Vrolijk [1990] suggested the clay-mineral transition of smectite to illite between 100-150°C controls the transition from aseismic to seismic slip, while Moore and Saffer [2001] provided a number of examples of diagenetic to low-grade metamorphic and consolidation processes that occur between 100-150°C. Plate interface temperatures for central Costa Rica just north of the Quepos mainshock have been estimated using conductive thermal models to be between 120-185°C at 10 km depth under variable values of shear stress (0-50 MPa) along the plate interface [Peacock et al., 2004]. Additionally, heat flow and temperature estimates calculated in the region of the 1999 Quepos aftershocks from the depth to the bottom-simulating reflector (BSR), a commonly noted reflection within forearcs that marks the bottom of the methane stability field, place the 100-150°C isotherms much shallower than 10 km [Pecher et al., 2001; Ingo Grevemeyer, University of Breman, Germany, personal communication, 2002]. Possible discrepancies between temperature estimates from BSR data and conductive thermal modeling not incorporating such data will need to be resolved in order to reliably correlate the updip extent of aftershock rupture to temperature isotherms.

The lack of a well-defined continental Moho and the dearth of well-constrained thermal models for central Costa Rica leaves little downdip information with which to compare the 30 km depth limit of the aftershock sequence. Tichelaar and Ruff [1993] suggested that the downdip transition from seismic to aseismic behavior correlates with mineral property transitions from stick-slip to stable sliding and conditionally stable behavior as the subducting plate becomes ductile at higher temperatures (~350-450°C)
and pressures. Hyndman et al. [1997] suggested an alternative process for low temperature subduction zones in which the downgoing plate encounters the upper forearc mantle before temperatures reach 350°C. If the forearc mantle wedge contains serpentinite, a rock believed to exhibit both stable sliding and strain rate dependent conditionally stable behavior under laboratory conditions, stick-slip behavior would no longer be supported along the plate interface once it is in contact with the forearc wedge. Thermal models for central Costa Rica indicate interface temperatures of 220-250°C at 30 km depth for 10-20 MPa of shear stress [Peacock et al., 2004]. Increasing the amount of shear stress upwards of 50 MPa is required to obtain temperatures of ~350°C at 30 km depth using conductive thermal models. This amount of shear heating would result in a temperature near 180°C at 10 km depth [Peacock et al., 2004], making the updip value more consistent with estimates made from BSR data. Although the continental Moho under central and southern Costa Rica has never conclusively been imaged, it most likely intersects the subducted plate deeper than 30 km [Matumoto et al., 1977; Protti et al., 1995b; Stavenhagen et al., 1998]. If the south central Costa Rican subduction zone is hotter than characterized by present thermal models, as indicated for the shallow portion of the seismogenic zone by BSR data, the downdip limit of seismicity at 30 km may have a strong thermal influence.

4.6 Conclusions

Four types of earthquakes occurring offshore central Costa Rica within the station coverage of the Osa experiment: 1) aftershocks of the 8/20/99 Quepos underthrusting earthquake, 2) outer rise earthquakes, 3) intraplate oceanic events below the interplate seismogenic zone, and 4) intraplate upper crust events. Individual depth error estimates for events located within the oceanic mantle are difficult to constrain, and these events are left uninterpreted. Intraplate oceanic events correlate spatially and temporally with the aftershock sequence of the Quepos earthquake and may be the oceanic plate response to changes in strain and fluid flow within the system. The small magnitude outer rise events reported here also occur within ~60 days of the mainshock and may have been triggered by the Quepos event, following a previous suggestion for outer rise activity recorded after the Mw 7.0 Nicoya Gulf event in 1990 [Protti et al., 1995b]. Upper plate activity has been previously noted in microseismicity studies of subduction zones [e.g., Shinohara et al., 1999], and the relationship of these events to general seismicity patterns in Costa Rica is an avenue of further research. Errors reported for well-constrained aftershocks along the plate interface, estimated from consideration of velocity model error, reading error, and station geometry, are 3 km horizontally and 5 km in depth. These errors are consistent with microseismicity errors reported within other subduction zones using OBS data in conjunction with non-simultaneous inversion location techniques [e.g., Hino et al., 1996; Shinohara et al., 1999]. Relative relocation of interplate events significantly improves the resolution of the aftershock rupture pattern however, with relative error estimates for closely spaced interplate events of <1 km in distance and depth.

Interplate aftershocks appear confined to a narrow zone corresponding to the interface between the Cocos Plate and Panama Block. The majority of well-located aftershocks occur below 10 km depth, 30-35 km from the trench and above 30 km depth,
95 km from the trench (Figure 4.7). The aftershock sequence correlates spatially with the downdip extension of the Quepos Plateau, and the locations of these events (Figure 4.7) may reflect the size and detailed structure of topographic features at depth that ruptured in the 1999 Quepos event. Slip during the 1999 mainshock and aftershock sequence represents the present limits of stick-slip behavior along this portion of the MAT; however, these limits very likely change over the earthquake cycle as physical and chemical parameters influencing the transition from stick-slip to stable sliding vary. Comparisons of recorded seismicity to available thermal and mechanical models for the updip and downdip limits of the seismogenic zone place the updip limits of seismicity at 120-185°C and the downdip limit between ~250-350°C. Neither result is inconsistent with proposed temperature-influenced models for the transition from stick-slip to stable sliding behavior and vice-versa, but further modeling is necessary to fully resolve the temperature influence on seismicity in this region.
Figure 4.1 Overview map of the Osa CRSEIZE experiment. Cocos plate oceanic crust formed at the Cocos-Nazca Spreading Center from 22.7-19.4 Ma and from 19.5-14.5 Ma (CNS-1 and CNS-2 from Barckhausen et al. [1998, 2001]) subducts along the Middle America Trench (MAT) offshore central and southern Costa Rica. Thickened crust of the Cocos Ridge subducts beneath and uplifts the Osa Peninsula. The Osa Experiment recorded the aftershock sequence of the 1999 Quepos underthrusting earthquake; shown is the local OVSICORI location (yellow star) paired with the Harvard Centroid Moment Tensor solution. Initial database locations of aftershocks through the 1D IASP91 model are scaled by local magnitude with maximum magnitude $M_L=4.3$ (red circles). Orange stars: Post-1980 large earthquakes ($M_w>7.0$); Black triangles: CRSEIZE seismometer locations; Black squares: OVSICORI network; Grey hexagons: RSN network; Bathymetry is from von Huene et al. [2000].
Figure 4.2 The Osa experiment primarily recorded aftershocks of the 8/20/99 Quepos earthquake (open star). Triangles denote seismic stations. The solid box marks the boundaries of the 3D velocity model compiled from refraction information (heavy dashed lines) [Ye et al., 1996; Stavenhagen et al., 1998; Walther, 2003] used to relocate seismicity with QUAKE3D. High quality, L1-norm computed QUAKE3D locations (red circles) are subdivided from the entire dataset (grey dots) and shown scaled by local magnitude (maximum $M_L$ 3.6).
Figure 4.4 Minimum1D P-wave velocity model for the Osa Experiment. $V_S$ is calculated using a constant $V_P/V_S$ ratio of 1.78.
Figure 4.3 3D velocity model representation of velocities along cross-sections defined by P-wave refraction data.  a) Walther [2003] data perpendicular to the Cocos Ridge.  b) Walther [2003] data parallel to the crest of the Cocos Ridge.  c) Stavenhagen et al. [1998] data along the flank of the Cocos Ridge near the location of the 1999 Quepos earthquake.  d) Ye et al. [1996] data defines the northern boundary of the 3D velocity model.  Depth is reported in km below sea level rather than in model z units.  Triangles: seismic stations; Star: 1999 Quepos earthquake.
Figure 4.5 QUAKE3D (a) and hypoDD (b) earthquake relocations plotted on a cross-section of the preferred 3D velocity model (OSA3D), approximately corresponding to the Stavenhagen et al. [1998] refraction line. Inverted triangles indicate locations of the seismic stations; T and C mark the trench and coastline respectively. a) Locations using QUAKE3D and the 3D velocity model (dots) define a dipping plane corresponding to the plate interface but show significant scatter around the 8/20/99 mainshock (star). The dashed box surrounds events interpreted as interplate aftershocks. b) HypoDD relative locations (black error bars centered on corresponding hypocenter) using the SVD method have error estimates for interface events on the order of 1 km. Note, relative hypocenters were solved for through the preferred 1D velocity model (shown in Figure 4.4), and event clusters were adjusted in absolute space using QUAKE3D absolute locations.
Figure 4.6 Depth distribution of interplate seismogenic zone earthquakes. The number of events located within 5 km of the Cocos/Panama Block plate interface (box in Figure 4.5a) is plotted versus depth using 2.5 km depth bins. A double Gaussian (solid line) distribution best fits the dataset. The 5th percentile occurs at 10 km below sea level and the 95th percentile at 30 km, defining the statistically significant updip and downdip limit of aftershock rupture.
Figure 4.7 Best QUAKE3D earthquake locations. Earthquakes occurring along the seismogenic plate interface (red circles) define the region ruptured in the 8/20/99 main shock (dashed line). Blue circles indicate the outer rise events, oceanic intraplate events, and continental intraplate events discussed in the text. Some events are left uninterpreted within the aftershock rupture area due to inconsistent depth locations with map view locations. The aftershock sequence occurs directly downdip of the extension of the Quepos Plateau and associated seamounts (QP as outlined) and has an aspect ratio similar to incoming seamounts. In addition to the 1999 Quepos main shock, a $M_w$ 6.4 underthrusting earthquake occurred south of the study area in 2002 (grey star, NEIC location and focal mechanism).
Figure 4.8 Temporal aftershock pattern, scaled by depth. Colors reflect data of event; darker colors occur closer to the date of the Quepos earthquake (star). a) Events recorded between 9/24/99 and 10/20/99 include earthquakes downdip of the mainshock and a number of outer rise earthquakes. b) Events recorded 3 months after the mainshock (10/21/99-11/20/99) occur updip of the mainshock and are concentrated along two linear streaks that correspond with an updip extension of two main patches of moment release defined by Bilek et al. [2003] (star and box).
<table>
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<tr>
<th>Models</th>
<th>Absolute Lat.,a</th>
<th>Absolute Lon.,a</th>
<th>Distance, km</th>
<th>Absolute Depth,a</th>
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<tr>
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<td>km</td>
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<td>km</td>
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<td>0.9(n)</td>
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<td>2.4(-)</td>
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<tr>
<td>L1, NoGrad-L1c</td>
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<td>L1, OBSonly-L1</td>
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<td>7.6</td>
<td>6.9(-)</td>
</tr>
</tbody>
</table>

a Shift of first model relative to second: n = no shift; - N,E, deeper; + S,W, shallower
b No low velocity zone above the plate interface
c No gradient smoothing between velocity layers
d Station corrections applied

**Table 4.1:** Earthquake location differences due to model and norm criteria