Lithospheric structure of the Costa Rican Isthmus: Effects of subduction zone magmatism on an oceanic plateau

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Abstract. We present the results of a multidisciplinary geophysical study, conducted to investigate the lithospheric structure of the Costa Rican Isthmus. The physical properties of the lithosphere are resolved by three-dimensional (3-D) simultaneous inversion of velocity anomalies and hypocenter parameters using local earthquakes and 2-D forward modeling of onshore and offshore seismic refraction and gravity data. According to our results, the northern half of the Costa Rican Isthmus is constituted by a ~40-km-thick crust, with a 6- to 7-km-thick oceanic crust subducting under it. The uppermost level of the basement and most of the marginal wedge show intermediate velocities and high densities, in good agreement with those described for flood basalts. The midlevel shows velocities and densities representative of oceanic crust. The bottommost level (20-40 km) shows high velocities and densities, typical of mafic rocks, and the upper mantle displays anomalously low densities and velocities. Intracrustal heterogeneities at intermediate wavelengths are indicated by prominent velocity anomalies. These results are consistent with a basement beneath the Costa Rican Isthmus being part of the Caribbean plateau, originated at 85-90 Ma with the onset of the Galápagos hotspot. The upper level corresponds to the flood basalts extruded during this phase, and it includes most of the marginal wedge. The second level represents the preexisting oceanic crust. The mafic lower crust, intracrustal heterogeneities, and anomalous upper mantle are interpreted to be built up by underplating, intrusion, and crystallization of basaltic melts, formed under the influence of subducting lithosphere dehydration.

1. Introduction

There is growing evidence that the origin and evolution of new continental crust is related with at least two types of magmatic processes: intraplate magmatism induced by mantle plumes, and island arc magmatism at convergent margins [Rudnick, 1990; Rudnick and Fountain, 1995; Taylor and McLennan, 1995]. The magma produced by these processes contributes to modify and thicken the oceanic crust, preventing its subduction and favoring its accretion to the active continental margins [Ben-Avraham et al., 1981; Reymer and Schubert, 1986; Taira et al., 1992]. Some of the most striking effects of magmatism are observed at lower crustal levels. It has been pointed out that the lowermost crust of the continents, and especially that of island arcs, is constituted by mafic rocks [Holbrook et al., 1992; Christensen and Mooney, 1995]. Different authors postulated that such mafic crust may be generated by underplating of basaltic melt and by crystal fractionation of magmas, which leave behind mafic residuals [Rudnick, 1990; Rudnick and Jackson, 1995]. This interpretation seems to contradict former hypotheses, which affirm that the bulk composition of the continents is of anesitic nature [Taylor, 1977; Taylor and McLennan, 1985]. However, recent studies have shown that the newest continental crust can have a basaltic composition, not a granitic one. The observed structure and geochemistry of older continents may then be the result of arc magmatism and metamorphism, which would modify the original basaltic crust into a lowermost basaltic layer overlain by a more silicic upper crust [Abbott et al., 1997].

The Costa Rican Isthmus (CRI) constitutes a good location to match this scheme of crustal growth, since it has been extensively affected by a mantle plume (origin of the Caribbean plateau) and shows a subduction zone. Several experiments have been carried out in this zone during the past decade. Most of them were centered on the convergent margin off the Nicoya Peninsula. Swath bathymetry surveys [von Huene and Fluhr, 1994], multichannel seismic reflection surveys [Shipley et al., 1992; von Huene and Fluhr, 1994; Hinz et al., 1996], Deep Sea Drilling Project (DSDP) and Ocean Drilling Program (ODP) drilling [von Huene et al., 1985; Kimura et al., 1997], magnetic surveys [Barckhausen et al., 1998], heat flow measurements [Langseth and Silver, 1996], and seismic refraction profiling [Ye et al., 1996; Christeson et al., 1999] have been conducted in the area. These works have allowed determination of the crustal structure of the margin, the geometry of subduction, and the main processes taking place at the subduction zone. Moreover, useful information on the physical properties, composition and origin of the margin wedge, has been extracted.

In addition, some onshore studies have reported different geological structures of the CRI, based on dating and geochemical composition of the Nicoya Complex outcrops [e.g.,

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2. Tectonic Setting

The CRI is located at the boundary between two lithospheric plates: the Caribbean plate, which includes the CRI and extends toward the east until the Lesser Antilles island arc and the Cocos plate, which subducts along the Middle American Trench (MAT) from southern Mexico to southern Costa Rica (Figure 1).

The Cocos plate is generated at the East Pacific Rise (EPR), in the west, and at the Galápagos Spreading Center (GSC), in the south. The boundary between both crustal domains was long believed to be the rough-smooth boundary [Hey, 1977]. However, recent works with better magnetic coverage [Barckhausen et al., 1998], swath bathymetry [von Huene et al., 1995], and dating [Werner et al., 1999] placed this transition slightly northward, at the middle of the Nicoya Peninsula. The plate created at the EPR is typical oceanic lithosphere, while the plate created at the GSC has been overthickened by igneous material extruded at the Galápagos hotspot, as evidenced by numerous small seamounts, oceanic plateaus, and ridges (Figure 1). The Cocos Ridge traces the path of the Galápagos hotspot through the Cocos plate. Based on structural differences of the Cocos plate, three morphological provinces have been distinguished [von Huene et al., 1995]. The northern one subducts under the Nicoya Peninsula and corresponds to a typical subduction zone with a crustal thickness of ~6 km [Ye et al., 1996], dip angle of ~45°, and associated seismicity down to 200 km depth [Pratt et al., 1995b]. The Quepos segment (central part) subducts south of the Nicoya Peninsula. Here the crust has been slightly thickened by Galápagos hotspot-derived material [Canales et al., 1997; Stavenhagen et al., 1998]. The geometry of

![Figure 1. Tectonic map of the study region including the most significant features. RSB, rough-smooth boundary; MAT, Middle America Trench; EPR, East Pacific Rise; GSC, Galápagos Spreading Center.](image-url)
Figure 2. Local and regional maps showing the location of the Costa Rica Isthmus and the main tectonic and geographical features of the region. Triangles show location of the currently active volcanoes. The locations of the igneous outcrops are also shown (Nicoya, Herradura-Quepos, and Osa). Convergence rates off Nicoya and Osa Peninsulas were taken from DeMets et al. [1994].

subduction in this segment is similar to the northern one, although it shows a less inclined and shallower Wadati-Benioff zone when compared to the Nicoya segment [Protti et al., 1995b]. The narrow shelf off the southern part (adjacent to the Osa Peninsula) corresponds to the subduction of the Cocos Ridge. The differences between this segment and the central and northern ones are remarkable. The dip angle of subduction is much lower (15-20°), with associated seismicity down to only 70 km depth [Protti et al., 1995b], and there is no current arc magmatism [Kussmaul et al., 1994].

The tectonic evolution of the CRI is more controversial. Most of the recent geochemical [Sinton et al., 1997; Hauff et al., 1997; Alvarado et al., 1997], geological [Gursky, 1988; Donnelly, 1994] and geophysical [Bowland and Rosencrantz, 1988; Sallarès, 1999; Christeson et al., 1999] data suggest that the basement of the CRI represents the westernmost end of the Caribbean plateau. Therefore the igneous outcrops of the Nicoya Complex (Figure 2) and most of the marginal wedge are thought to be part of the oceanic plateau basement.

The Caribbean plateau is thought to have formed during Late Cretaceous, as a result of the initial phase (the "head phase") of a mantle plume which originated on the later stage the Galápagos hotspot [Duncan and Hargraves, 1984; Richards et al., 1989; Hill, 1993]. Once integrated in the lithosphere, the Caribbean plateau moved toward NE, into the South American and North American plates, and collided with the Lesser Antilles island arc at 75-80 Ma [Burke, 1988; Pinell and Barrett, 1990]. The suture between the plateau and the Lesser Antilles induced the onset of subduction beneath the CRI and the first phases of andesitic volcanism during the Campanian [Lundberg, 1991; Kussmaul et al., 1994].

An alternative description suggests that part of the CRI represents a fragment of oceanic crust or the remnant of an island arc as old as Jurassic. This arc would have originated independently of the Caribbean plateau, being accreted afterward in its western margin [e.g., Wildberg, 1984; Frisch et al., 1992]. This hypothesis is based in the biostratigraphic ages of the radiolarian cherts (100-150 Ma), which apparently separate the upper and lower parts of the Nicoya Complex. However, recent 39Ar/40Ar dating [Sinton et al., 1997] has demonstrated that both the upper and lower parts of the Nicoya Complex have approximately the same age (~88 Ma). Thus the radiolarian
cherts are interpreted as a remnant of the sedimentary cover of the preexisting oceanic crust, which was intruded and deformed by the extrusion of the basement in the Late Cretaceous [Donnelly, 1994].

3. Geophysical Data Sets

This multidisciplinary study benefits from the integration of three independent geophysical data sets: local seismicity, seismic refraction and gravity. The approach was to consider independently these data sets, using different methodologies to process and interpret them. Afterwards, all the results were merged into a single geophysical model.

3.1. Seismic Tomography With Local Earthquakes

3.1.1. Data and method. The local seismicity data includes $P$ waves first arrivals from 5166 events, recorded at 81 stations from the Costa Rican seismic networks between 1991 and 1998 (Figure 3a). Two independent seismic networks currently operate in Costa Rica. The first one is constituted by 32 short period

![Figure 3. Location map of (a) the 5166 local earthquakes from the initial data set, and (b) the 583 local earthquakes used in the tomographic inversion. The 81 seismic stations which contain any readings are also included (triangles).](image)

stations (1 s) and is maintained by the Instituto Costarricense de Electricidad and the Universidad de Costa Rica [Güendel et al., 1989]. The second one includes 31 stations from the Observatorio Vulcanológico y Sismológico de Costa Rica of the Universidad Nacional [Colombo et al., 1997]. The rest of the 81 stations correspond to temporary local networks set near the volcanic arc for distinct microseismicity studies. The whole data set has been compiled and standardized since 1995 by the Centro de Prevención de Desastres Naturales de América Central. This institution provided the data set used in the tomographic inversion.

Travel time residuals were inverted using a tomographic technique to obtain a three-dimensional (3-D) velocity field and the hypocentral parameters of local earthquakes. We followed an inversion scheme similar to that proposed by Kissling [1988] and Kissling et al. [1994]. The tomographic inversions have been performed using VELEST [Kissling et al., 1994] and SIMULPS12 [Thurber, 1993] algorithms of simultaneous inversion.

This inversion method is described in further detail by Sallarès et al. [2000], and thus we only summarize here the most relevant parts of the procedure:

1. An adequate data subset for tomographic inversion is selected (Figure 3b). The events should be recorded by at least 10 stations, showing a gap smaller than 180° and a root-mean-square error of <2.5 s. These criteria are considered to be sufficiently strict to guarantee stability and convergence on the inversion of hypocentral parameters [Solarino et al., 1997].

2. The best reference 1-D velocity model is estimated (Figure 4). This model is achieved checking the space of parameters by

![Figure 4. Velocity-depth profile corresponding to the best reference 1-D model (or minimum 1-D model) inverted from local seismicity data. The a priori models constrain the range of convergence of the model.](image)
Figure 5. Anomalies of the velocity field obtained after 3D inversion, for layers (a) 1 (0-2 km; 1σ=5.2 km/s); (b) 2 (2-6 km; 1σ=6.0 km/s); (c) 3 (6-12 km; 1σ=6.3 km/s); (d) 4 (12-20 km; 1σ=6.5 km/s); (e) 5 (20-30 km; 1σ=6.8 km/s); and (f) 6 (30-43 km; 1σ=7.1 km/s). Anomalies are expressed in percentage of deviation with respect to the best reference 1-D model (Δv/σ in %), only for the well-resolved zones (R>0.4). Dots display location of well locatable local earthquakes within these layers.
performing a simultaneous inversion with a set of distinct 1-D initial models [Kissing et al., 1994]. The best reference 1-D model is considered the one that minimizes residual times and is compatible with the a priori information (e.g., velocity-depth profiles from seismic refraction).

3. The optimal parameterization is determined. The parameterization of the velocity field is set according to the results of checkerboard tests using a synthetic data set. This analysis allows us to estimate qualitatively the minimum size of the anomalies which can be resolved by the data in the different parts of the study area [Sallarès et al., 2000].

4. The resolution is evaluated (Figure 5). The diagonal element of the resolution matrix determines the discrepancy between the obtained solution and the optimal one (that which would fit exactly the observations). The comparison of this parameter with the results of the checkerboard tests allows automatic discrimination between the well-resolved (interpretable) zones and the poorly resolved (noninterpretable) ones.

3.1. Results. The results of the tomographic inversion are shown in Figures 4 (1-D model) and 5 (3-D model). The main global characteristics inferred from the best reference 1-D model are the following: the crudely constrained velocity-depth contrasts obtained at all crustal depths; the high velocities obtained at the lowermost crustal levels (7.1 km/s between 30 and 43 km depth); and the low velocity of the "upper mantle" (7.5 km/s between 43 and 65 km depth). This velocity-depth distribution makes it difficult to distinguish crustal boundaries within the crust and to estimate the location of the Moho. However, the results are consistent with the presence of a remarkably thick crust (240 km) under the CRI, as suggested in most of the previous works [Matumoto et al., 1977; Case et al., 1990; Sallarès et al., 1999].

The best reference 1-D model was used as the initial model for the 3-D simultaneous inversion. Figure 5 shows the results obtained in the well-resolved zones after the 3-D inversion. These zones are defined by nodes displaying diagonal elements of the resolution matrix >0.4. See Sallarès et al. [2000] for additional details.

The 3-D velocity field shows a different behavior depending on the depth range. The uppermost levels (0-6 km depth; Figures 5a and 5b) reflect the most outstanding geological features observed at the surface in the central part of the CRI. Thus the lower velocities are associated with the volcanic arc in the northern half and with prominent Quaternary sedimentary basins. The higher velocities can be related to the basement highs and to the more consolidated Tertiary volcanic rocks. The distribution of hypocenters (Figures 5a and 5b) shows two different zones, separated by a SW-NE seismic alignment: the northern part is seismically active and can be correlated with the southernmost segment of the currently active volcanic arc. The southern part is seismically inactive and is coincident with the gap on the volcanic arc.

The midcrustal levels (6-20 km depth; Figures 5c and 5d) show a very similar distribution of seismicity (active in the north and inactive in the south) but a quite different velocity field as compared with the shallower ones. The velocity field of the northern half is highly heterogeneous, being characterized by conspicuous intermediate-scale (λ~20 km) high-and-low velocity anomalies (Figure 5c and 5d). In contrast, the southern part shows a much more homogeneous velocity field, characterized by relatively high seismic velocities. Beneath these levels (>20 km depth; Figure 5e and 5f) the 3D velocity field is insufficiently resolved, and thus it is difficult to infer its local characteristics.

3.2. Seismic Refraction

3.2.1. Data and method. The seismic refraction data were acquired along two profiles (offshore and onshore) located at the northern part of the CRI across the Nicoya Peninsula (Figure 6). The first part of the experiment comprised the acquisition of the offshore profile in March 1995, during a cruise of the RV Maurice Ewing. A profile 85 km long perpendicular to the coast line was shot, and it was recorded by 10 ocean bottom hydrophones (OBH) from the Geomar Forschungszentrum für marine Geowissenschaften (GEOMAR) [Fiehle and Bialas, 1996], 10 ocean bottom seismographs from the University of Texas Institute for Geophysics (UTIG) [Christeson et al., 1999], and 27 land stations from the GeoForschungs Zentrum Potsdam (GFZ) deployed across the CRI (Figure 6). In this study, we only were able to use the data from the German OBH and the land stations. All the record sections obtained at the OBH were of high quality, but only five of the land stations were good enough to be used in the modeling.

The second part of the experiment included the onshore profile, acquired in March 1996. Eight quarry-blast shots were fired at five different locations along a 170 km long profile, from Marbella to Los Chiles (Figure 6), and they were recorded by 60 landstations with two deployments. Shots A, C, and E were fired twice, resulting in a maximum of 120 recordings along the whole profile, while shots B and D were fired once, and thus they were recorded only by 60 stations in one half of the profile. Therefore the global data set included 20 high-quality record sections acquired along a 255 km long profile, which constituted the data set used in the forward modeling (five of the record sections are shown in Figure 7). A preliminary model using the onshore data and results from former offshore data [Ye et al., 1996] are given by Sallarès et al. [1999]. Nevertheless, current results include the whole data set and show significant differences with the former ones thus we have included here a full description of them.

The most relevant features of the onshore and offshore data are the following:

1. All the OBH record sections show a similar pattern (Figures 7a and 7b). Near the stations, we found very low seismic velocity phases (>1.8 km/s). These phases are interpreted as the refraction and reflection within the unconsolidated sedimentary cover of the continental slope (Pn, PNP). From 10 km and onwards, the first arrivals are correlated with the refraction phase within the marginal wedge (Pm). This phase shows a highly varying apparent velocity (3.5-6.0 km/s) and length (10-45 km), reflecting the thickening of the prism and the velocity increasing as we approach the coast line. First and secondary arrivals at the intermediate range of distances are well correlated with refracted and reflected waves within the oceanic crust (Pm, PP, and PmP). These phases show apparent velocities between 6.0 and 7.0 km/s. The deepest and fastest PmP and PP phases (>8.0 km/s) are observed as first arrivals at growing distances from OBH 2 (15 km) to OBH 20 (50 km), reflecting the deepening of the subducting slab.

2. The record sections from the land stations (Figure 7c) are quite similar to those of the OBH closest to the coastline (18 and 20). The first arrivals are constituted by a long Pm phase (up to 50 km from the receiver) showing a mean velocity of ~5.0 km/s. At greater distances (>45 km), PnPP and PnP phases are observed to more than 100 km from the stations.

3. The record sections from the landshots (Figures 7d and 7e) are very different from the offshore ones. First arrivals for the whole range of distances have been associated with refracted waves at the different crustal levels. Near the source (<10 km), they represent a slow phase (2.5-4.0 km/s) interpreted as the
refraction within the sedimentary cover \((P_s)\). This phase is not observed in shot A (Figure 7d). From this point to 40 km from the source the first arrival corresponds to the refraction within the uppermost level of the basement \((P_s)\), showing an apparent velocity of 5.0-5.5 km/s. Between 40 and 130 km a slightly faster phase (6.0-6.5 km/s) is observed, which corresponds to the refraction at midcrustal levels \((P_s)\). From 130 km to the end of the profile the first arrival is correlated with a refraction within the lower crust \((P_c)\), with an apparent velocity of 7.0 km/s. Shots A and E show also two fast second arrivals (from -100 to 170 km), which have been associated to reflections at the base of the CRI crust \((P_{sP})\) and at the base of the Cocos plate \((P_{wP})\), respectively. These phases are hard to correlate due to the scattered pattern shown in shot E (Figure 7e) and to their low relative amplitude (especially \(P_{wP}\)) in shot A (Figure 7d). Hence the picking uncertainties for these phases are higher than for the first arrivals (0.2-0.3 s), but we considered them in the modeling since they contain information on the lower crustal levels.

The 2-D velocity-depth profile and synthetic seismograms were computed using the forward modeling algorithms of Zelt and Smith [1992].

3.2.2. Results. The 2-D velocity model along profile P1 (Figure 6) with the whole set of OBH and landstations is shown in Figure 8, and the synthetic seismograms and ray tracing obtained with this model are shown in Figure 7.

All the possible refracted and reflected rays across the model have been traced. Most of the synthetic seismic phases fit the arrival times with differences of less than 0.2 s, and the relative amplitudes agree with those observed in the record sections. Therefore the \(P\) phase in the proximal distance range, \(P_sP\) and \(P_c\) in the intermediate range and \(P_{wP}\) and \(P_s\) in the distal range represent the most energetic phases of the offshore part (Figures
Figure 7. Ray tracing for identified phases (a1-e1), record sections (a2-e2), and calculated synthetic seismograms (a3-e3) corresponding to (a) OBH 6, (b) OBH 14, (c) land station 2, and (d) land shots A and (e) E. We included the identified seismic phases in both the record sections (a2-e2) and the synthetic seismograms (a3-e3).

The reflection from the top of the Cocos plate ($P_P$) is modeled in the synthetic seismograms as a phase of high amplitude beneath the water wave (Figure 7a). However, this phase is not observed in the record sections. We believe that it is probably hidden by multiples and reverberations of the water wave.

$P_n$, $P_s$, and $P_l$ constitute the most prominent phases of the synthetic seismograms from the landshots, and the amplitude from the intracrustal reflections is small. The only reflected waves recovered in the synthetic seismograms are represented by the deep $P_{n}P$ and $P_{s}P$ phases (Figures 7d and 7e). The synthetic seismograms show a low relative amplitude for these phases,
especially in shot A (Figure 7d), and they are recovered only at a far distance range (>100 km). These results are in good agreement with the observations of the record sections.

The offshore part of the model is characterized by the subduction of a standard oceanic crust (6-7 km thick) under the CRI, with a dip angle beneath the marginal wedge of ~10°. A thin sedimentary layer of <0.5 km showing quite low seismic velocities (1.9 km/s) covers the crust. The oceanic layer 2 is 2.0-2.5 km thick, and its averaged seismic velocity varies laterally from 5.6 km/s at the trench to 6.3 km/s beneath the coastline. The seismic velocity at the top of this layer is as low as 3.8-4.0 km/s near the trench and shows a remarkable vertical velocity gradient, reaching 6.3 km/s at the base. Layer 3 is 3.5-4.0 km thick, and its seismic velocity varies from 6.7 km/s at the trench to 7.1 km/s beneath the coast line. The velocities of both crustal layers are within the range identified by White et al. [1992] for oceanic
crust. The upper mantle velocity of the Cocos plate reaches typical values of 8.1-8.2 km/s.

The marginal wedge shows remarkable lateral and vertical velocity gradients. Thus velocity increases abruptly from 2.2 km/s at the trench to 4.2 km/s 10 km landward. From this point to the coastline the velocity gradient is smoother, and the velocity reaches maximum values of 4.8 km/s at the surface and 5.9 km/s at the base. This velocity model shows small differences with the model obtained by Christeson et al. [1999] with the full set of OBH and OBS and land stations. Christeson et al. determined higher velocities (6.8 km/s) at the base of the wedge beneath the coast line, but lower averaged velocities in the oceanic layer 2 of the Cocos plate beneath the margin wedge (always lower than 5.0 km/s). Elsewhere, both models are compatible in terms of averaged velocities.

A 40 km thick crust constitutes the CRI in the strict sense. A
sedimentary cap with highly varying thickness (2-4 km) and velocity (2.2-4.0 km/s) mostly covers the crust. The lowest velocities are obtained at the volcanic arc (2.2-2.5 km/s), and the highest ones are obtained at the distal sedimentary basins (3.5-4.0 km/s). The lack of intracrustal reflections in the record sections makes it difficult to distinguish internal crustal boundaries. Nevertheless, we defined three "theoretical" levels tied to velocity but not associated with first-order velocity contrasts to facilitate its description. Level 1 corresponds to the uppermost part of the basement (~2-8 km depth), where the seismic velocity increases from about 5.3 to 6.0 km/s. Level 2 extends from ~8 to 18-20 km depth and shows velocities between 6.1 and 6.8 km/s. Level 3 includes the lower half of the crust (~20-40 km), and the seismic velocity increases from 6.9 to 7.2 km/s. The secondary arrivals identified in the record sections from shots A and E (Figures 7d and 7e) have allowed to estimate the Moho depth (~40 km) and the upper mantle velocity (~7.5 km/s) beneath the CRI. The coincident results between seismic refraction and 1-D
seismic tomography at deep levels (≥20 km depth) validate these results in spite of the low amplitude and picking uncertainties of the deep seismic phases.

3.3. Gravity

3.3.1. Data and method. The location of the gravity profile is coincident with the seismic profile P1 (Figure 6). The gravity data were taken from the data sets of Sallings et al. [1994] and Sandwell and Smith [1997], being projected along this profile with one data point each 5 km. This data set includes free air anomalies at sea and Bouguer anomalies on land, and the reference density is 2670 kg/m³. The forward modeling was accomplished using an algorithm based on the formulation and subroutines of Ramapo and Murthy [1989] and Murthy and Rao [1979]. The geometry and layering of the density profile (Figure 9) were taken approximately from the 2-D velocity profile, and velocities were converted into densities using different empirical velocity-density relations, depending on the estimated...
3.3.2. Results. The velocities of the global model have been converted into densities using different velocity $v$ density $\rho$ relations. Some of the polygons corresponding to the different layers have been divided into smaller ones in order to reflect approximately the lateral velocity (density) variations (Figure 9). For the Cocos plate, we used the $v$-$\rho$ relation of Carlson and Raskin [1984] for oceanic crust. We considered averaged velocities of 6.0 and 6.9 km/s for oceanic layers 2 and 3, which determined averaged densities of 2850 and 2950 kg/m$^3$, respectively. The density of the upper mantle was set at 3300 kg/m$^3$ according to the standard values achieved for the seismic velocity (8.1-8.2 km/s). The density of the sedimentary cap (1900 kg/m$^3$) was fixed later to fit the values of the gravity anomaly.

The polygon including the marginal wedge and the uppermost part of the CRI basement has been divided into three bodies. The first one only includes the frontal part of the wedge (~10 km). Its averaged velocity is ~3.2 km/s and the density is 2400 kg/m$^3$, typical values for sediments. The second body includes the rest of the marginal wedge, and the third one includes the Nicoya Complex and the uppermost basement of the CRI. The averaged velocities of these two bodies are 4.4-4.6 and 5.4-5.6 km/s, respectively.

It is accepted that most of the margin wedge represents an offshore extension of the Nicoya Complex, and thus they must be composed of similar materials [von Huene et al., 1985; Hinz et al., 1996; Kimura et al., 1997; Christeson et al., 1999]. The dominant components of the Nicoya Complex are basalts, as is described in numerous works [Wildberg, 1984; Frisch et al., 1992; Donnelly, 1994; Sinton et al., 1997]. The main difference between these works concerns the possible origin of the basalts.

On the one hand, Donnelly [1994] and Sinton et al. [1997] consider that the Nicoya Complex is part of the basement of the Caribbean plateau and thus it originated by the onset of a mantle plume. According to this hypothesis, flood basalts would mostly compose the basement of the CRI and the marginal wedge, instead of mid-oceanic ridge basalts (MORB).

On the other hand, Wildberg [1984] and Frisch et al. [1992], among others, proposed that the CRI represents the remnant of an island arc built up on a “normal” oceanic crust. This hypothesis would imply that the basement of the CRI is a fragment of oceanic crust, and thus its uppermost level would be composed essentially of MORB (oceanic layer 2).

Hence we have considered the previous hypotheses to convert velocities into densities:

1. The basement is composed of flood basalts. To test this, we considered a $v$-$\rho$ relation determined from analysis of wire line logs at the flood basalts of the Vering volcanic margin [Planke et al., 1994].

2. The basement is composed of MORB. In this case, we considered the relation of Johnston and Christensen [1997],
which shows the physical properties of oceanic layer 2 from the most significant DSDP/ODP holes. The results are summarized in Table 1.

The density of the "upper mantle" is considered to be lower than normal, according to the values of the seismic velocity obtained from 1-D seismic tomography and seismic refraction (7.5 km/s). This velocity would correspond to a density of ~3100 kg/m³ [Gladchensko et al., 1997]. Therefore the densities of crustal levels 2 and 3 have to be between the values of level 1 (~2900 kg/m³) (Table 1 and Figure 9) and upper mantle (~3100 kg/m³). The sedimentary cover has been divided into four bodies to reflect the lateral velocity contrasts. Their densities have been adjusted to fit the gravity anomalies. The lowest values (~2100 kg/m³) are obtained beneath the volcanic arc, while the highest ones (~2500 kg/m³) are obtained at the distal zones.

Figure 9 shows the best fitting 2-D density-depth model along

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**Figure 9.** (a) The 2-D×2 density-depth profile obtained along profile P1 and (b) best fit of gravity anomaly. The geometry of layers is taken from the velocity-depth profile (Figure 8).
the whole profile. This model confirms some of the most outstanding features inferred from the seismic methods. Hence the high densities of the marginal wedge (2730 kg/m³) and of the CRI uppermost basement (2910 kg/m³) are in good agreement with those described for flood basalts (Table 1). Moreover, we did not find important density-depth contrasts within the crust, and the density of the upper mantle is lower than normal (3100 kg/m³ instead of 3200-3300 kg/m³). The offshore section is in agreement with the expected values at crustal and upper mantle levels.

The resolution of these results has been estimated considering two alternative models to that of Figure 9 (see Figure 10). These models hold the same distribution of bodies but have different densities. The first model (model A) was built to check the influence of level 1 density on the computed anomalies. Thus, if we consider densities compatible with the velocity-density relation inferred for MORB (2600 kg/m³ within the wedge and 2800 kg/m³ within the CRI basement) (Table 1), it is not possible to fit the anomalies (Δg < 45 mGal), even if we try to adjust the densities of the rest of bodies. This fact indicates that the influence of the uppermost level density is very important, especially over the outcrops of the Nicoya Complex. Therefore the density and thickness of the basement mainly control the gravity anomaly thus they are well controlled in the model.

In the second model (model B), we considered higher densities in the upper mantle (3200 and 3300 kg/m³), which produced remarkable discrepancies between observed and computed anomalies (≥30 mGal). This indicates that a low density in the "upper mantle" is also necessary to fit the gravity anomalies.

3.4. Joint Interpretation

The results from the three individual data sets have been integrated into a global one showing the structure, geometry, and physical properties of the different lithospheric levels. The offshore section is represented in Figure 11 and the onshore section is represented in Figure 12.

3.4.1. Offshore section. This part of the model includes the subducting Cocos plate and the marginal wedge, from the MAT to the coastline (Figure 11). The Cocos plate is 6-7 km thick and shows a dip angle of ~10° beneath the margin. It is covered by a thin sedimentary layer of unconsolidated marine deposits (<0.5 km) [see also Kimura et al., 1997]. The most significant feature of these layers is the increase in average seismic velocities from the west of the MAT (5.5-5.6 km/s in layer 2 and 6.7 km/s in layer 3) toward the coast line (6.3 km/s in layer 2 and 7.1 km/s in layer 3). This fact can be related to other observations. Langseth and Silver [1996] showed that the heat flow in the frontal part of this subduction zone is anomalously low (~14 mW/m²), and they suggested hydrothermal cooling due to water circulation within the crust to explain it. The presence of extensional faults in the subducting crust caused by lithospheric flexure [McAdoo et al., 1996] seems to correlate with this description. Hence we suggest
that the low velocities found westward from the trench are also related to the presence of fluids within the oceanic crust, which considerably diminish the seismic velocities (5-10%). The progressive increase of seismic velocities beneath the margin and that of heat flow (~30 mW/m² at 15 km from the MAT) could indicate a gradual ejection of the fluids due to compaction and metamorphism of crustal rocks as the Cocos plate enters the subduction zone. Two more observations seem to be consistent with this description: geochemical data at Site 1039 of ODP Leg 170 showed evidence of a seawater source deep in the section, implying flow of water through seafloor conduits into the upper part of the crust [Kimura et al., 1997]; and DSV Alvin submersible dives discovered widely distributed low temperature fluid vents in the frontal part of the prism [Kahn et al., 1996], although the source of the fluids (crust, sediments) is unknown.

The marginal wedge is 75 km long and extends from the MAT to the coastline. The frontal part (~10 km) shows a remarkable velocity gradient. From this point to the east the velocity gradient is smoother and the density is higher (Figures 8 and 9). These results are in general agreement with previous seismic reflection [Hinz et al., 1996] and refraction [Ye et al., 1996; Christeson et al., 1999] studies. These works suggested that only a small portion of the marginal wedge (8-10 km) is formed by accreted or recycled slope sediments, while the rest of the prism (~65 km) represents an offshore extension of the Nicoya Complex. Moreover, the analysis of core samples from ODP Leg 170 [Kimura et al., 1997] confirmed that the external part of the marginal wedge contains gabbroic and basaltic fragments which are similar to those outcropping in the Nicoya Complex and that the complete section of incoming sediments is subducting beneath the toe of the wedge.

Hence we suggest that the frontal section of the wedge is effectively composed by recycled slope sediments (as proposed by Kimura et al. [1997]), progressively consolidated eastward from the trench. The rest of the prism would be mostly composed by fractured or altered flood basalts. The velocity and density increases as we approach the coast are probably related to the diminishing of porosity (fracturing) of the basalts (Table 1). The lower velocities and densities coincide approximately with the location of fluid vents, where fluid ejection over the wedge is more significant. In contrast, the porosity of flood basalts near the coastline is practically insignificant, and thus velocities and densities are very similar to those obtained in the Nicoya Complex and in the CRI uppermost basement.
3.4.2. Onshore section. On the basis of our modeling, the CRI crust is around 40 km thick. The igneous crust is mostly covered by sedimentary and volcanic rocks (2-4 km thick). The lateral variations in the physical properties and the thickness of this layer reflect the differences in age and composition of the rocks. Thus the lowest densities and velocities (2.2 km/s, 2000 kg/m^3) are found in the volcanic arc, where the most recent accumulation of volcanic sediments (ashes) occurs [Kussmaul et al., 1994]. Velocities and densities increase at both sides of the volcanic arc. The proximal part shows intermediate velocities and densities (3.0 km/s, 2300 kg/m^3) that are associated with more consolidated volcanic sediments and alluvial deposits. The distal part (i.e., the Tempisque basin) is characterized by relatively high velocities and densities (3.8 km/s, 2500 kg/m^3) and can be related with Tertiary volcanic rocks corresponding to earlier magmatic phases of the volcanic arc [Kussmaul et al., 1994]. The 3-D seismic tomography results at the uppermost layers (0-6 km depth) also agree with this description (Figures 5a and 5b). Thus the low-velocity anomalies are concentrated around the currently active volcanic arc and at the most prominent sedimentary basins, while the high velocities can be associated with Tertiary volcanic rocks and basement highs.

The extrusive basement (level 1 in Figure 12) is found beneath the sedimentary cover, and it crops out along the coast (i.e., the Nicoya Complex). This level is ~6 km thick, with seismic velocities between 5.2 and 6.0 km/s and an averaged density of 2910 kg/m^3. Level 2 is 10-12 km thick and shows velocities ranging between 6.1 and 6.8 km/s and a density of 2950 kg/m^3. Level 3 extends from ~20-40 km depth and displays velocities between 6.8 and 7.2 km/s and a density of 3000 kg/m^3. As we stated above, all these intracrustal levels do not correspond to sharp velocity (density) contrasts, and thus the crust is defined by velocity gradients instead of first-order velocity contrasts from the top to the bottom. The upper mantle is characterized by low velocities and densities (7.5 km/s, 3100 kg/m^3).

These results are in good agreement with those inferred from 1-D tomography (Figure 13). This implies that the CRI lithospheric structure determined from the seismic profile P1 is
related with the head phase of a mantle plume which intruded the oceanic crust at ~90 Ma [Richards et al., 1989; Hill, 1993]. The second phase of the plume (the tail) then corresponds to the Galápagos hotspot [Duncan and Hargreaves, 1984], and thus the Cocos Ridge would mark approximately the trace of the Caribbean plateau from its initial to the final locations [Pindell and Barrett, 1990]. Most of the basaltic outcrops described throughout the Caribbean plate (e.g., the Nicoya Complex) show a similar geochemical composition, which is compatible with the magmas currently extruded at the Galápagos hotspot [Hauff et al., 1997; Sinton et al., 1998; Werner et al., 1999]. Moreover, the estimated age of the Nicoya Complex is ~88 Ma [Alvarado et al., 1997; Sinton et al., 1997], very similar to the estimated onset of the Galápagos mantle plume.

Our results are in good agreement with this description. The physical properties of the two uppermost levels at the CRI are compatible with the "oceanic plateau" scheme of crustal growth [Saunders et al., 1996], showing an upper level (level 1 in Figure 12) mainly constituted by flood basalts and a bottom one (level 2 in Figure 12) corresponding to the preexisting oceanic crust which was intruded and thickened by the extrusion of flood basalts in the Late Cretaceous. These two levels are commonly found in all the oceanic plates, such as Kerguelen [Charvis et al., 1995], Ontong-Java [Gladczenko et al., 1997], and the North Atlantic [Elidholm and Greu, 1994]. The crustal thickness and the physical properties of the lowermost crustal levels (level 3 in Figure 12) vary considerably depending on the tectonic setting of the area. It has been described that high seismic velocities (>7.3 km/s; Kerguelen, North Atlantic) indicate that this level is probably composed by a mixture of isotropic gabbros (oceanic layer 3) and ultramafic rocks from the mantle [Elidholm and Greu, 1994; Charvis et al., 1995]. Lower velocities (<7.3 km/s; Ontong-Java, CRI) are consistent with a lower crust formed by underplating, fractionation and crystallization of basaltic or picritic magmas [Gladczenko et al., 1997].

3.6. Origin and Composition of the CRI Lower Crust and "Upper Mantle"

There is some petrological [Kussmaul et al., 1994] and geophysical [Matumoto et al., 1977; Case et al., 1990; Sallarès, 1999] evidence indicating that the CRI lithospheric structure suffered important changes after the eruption of the Caribbean plateau in the Late Cretaceous. This evidence is reflected in our results by different observations: the crustal thickness of the CRI, nearly twice the averaged crustal thickness of the Caribbean plateau; the intracrustal heterogeneities observed at intermediate wavelengths (~20 km); the high velocities and densities found at lower crustal levels; and the anomalous physical properties of the "upper mantle" (low velocities and densities).

The most outstanding tectonic event effecting the CRI since the Late Cretaceous is the initiation and development of the subduction zone. This is estimated to have begun about 75-80 Ma, as a result of the collision and suture between the Caribbean plateau and the Lesser Antilles island arc [Burke, 1988]. The first phases of magmatism at the volcanic arc began ~70 Ma [Lundberg, 1991]. The dominant components of these oldest magmas are andesites, which have been related with the presence of a young island arc. During the Neogene the composition of extruded magmas transformed progressively from andesitic to calc-alkaline, implying a geochemical and structural evolution of the crust [Kussmaul et al., 1994]. Therefore we suggest that most of the structural and geochemical changes that have taken place in the CRI from the Late Cretaceous to the present may be related to subduction, according the scheme of evolution that we summarized below.
It is widely accepted that the formation of basaltic magmas responsible for building up of volcanic arcs takes place by partial melting of asthenospheric rocks at the mantle wedge above subduction zones [Wilson, 1989]. The melting processes are triggered by high-temperature fluids released from the subducting plate, owing to metamorphic dehydration of crustal rocks at depths greater than 80-100 km (726°C). The basaltic magmas rise through the lithospheric mantle and are underplated at the base of the crust. This process leads to a substantial diminishing of the seismic velocities and densities of the upper mantle as is observed, for example, in Alaska [Kissling and Lahr, 1991] or the Ryukyu [Iwasaki et al., 1990] and Aleutian [Fledner and Klemperer, 1999] island arcs. Another process contributing to the low velocities and densities of the upper mantle is the presence of water and the formation of hydrous minerals (i.e., serpentine) within the mantle wedge by fluids released from the subducting crust [Peacock and Hyndman, 1999].

Owing to their low viscosity, the underplated magmas can easily intrude the overlying crust. In the CRI, occurrence of crystal fractionation processes as the magma ascends through the crust is revealed by the composition of the lavas which have been erupted in the volcanic arc [Kussmaul et al., 1994]. Hence only the lighter components of the magma reach the surface, while most of them are trapped at or near the crust-mantle boundary, crystalizing and forming mafic rock types [Rudnick, 1990; Rudnick and Jackson, 1995]. The mafic residuum is then considered responsible for the progressive thickening of the lowermost crustal levels, while the intracrustal heterogeneities revealed by the seismic tomography in the NW part of the CRI (Figures 5c and 5d) may represent magmatic intrusions.

Figure 14 shows the phase diagram and the seismic velocities corresponding to underplating of tholeiitic basaltic magmas [Furlong and Fountain, 1986]. For a high-temperature regime characteristic of subduction zones, the expected seismic velocities and composition of the rocks at lower crustal and upper mantle levels (30-50 km depth) are compatible with our results and with the predictions of this scheme of crustal growth. Thus velocities at the lowermost crust (30-40 km) are between 6.9 and 7.2 km/s, and the dominant components are expected to be mafic granulites. At greater depths (40-50 km) the estimated velocities are just slightly higher (<7.6 km/s), and the bulk composition of these levels would be similar to that of the lowermost crust.

The obtained velocity-density relations in the lower crust and upper mantle of NW Costa Rica also suggest that they could be mainly composed by mafic rocks rather than felsic or ultramafic rocks, as is illustrated in Figure 15 [Rudnick and Fountain, 1995]. These observations show that the lowermost levels of the CRI crust could have an overall mafic composition, which would be comparable to that described in continental island arcs [Holbrook et al., 1992; Christensen and Mooney, 1995; Fledner and Klemperer, 1999]. Therefore we suggest that the CRI can be considered as a fragment of new continental crust with a bulk basaltic composition, as described by Abbott et al. [1997].

3.7. Seismic Regime and Nature of the Moho (M2)

The distribution of seismicity beneath the NW part of the CRI provides additional evidence for the existence of a remarkably thick crust. Figure 16 displays the hypocentral projection of local earthquakes that occurred between 1965 and 1995 in the vicinity (<40 km) of the seismic profile P1 [Rojas et al., 1993]. The hypocentral distribution allows us extend the structural model obtained from seismic refraction down to ~150 km depth. Most of the seismicity can be directly correlated with the subduction, but not all of it. It is noticeable that intracrustal seismicity beneath the CRI extends down to 35-40 km depth, which is in good agreement with the crustal thickness inferred from seismic refraction and tomography. All these events have magnitudes over 4.0, and they were located using mainly seismic stations placed throughout the CRI. Therefore events located onshore should have much lower location uncertainties (≤5 km) than events located offshore (≥10 km). This would indicate that the lower crust in the NW part of the CRI is seismically active, while the upper mantle is inactive.

Different processes have been suggested to account for intracrustal seismicity of the CRI. There is back arc thrusting in SE Costa Rica, which has produced recent big earthquakes.
SALLARÈS ET AL.: LITHOSPHERIC STRUCTURE OF COSTA RICA

[Protti and Schwartz, 1994]; there is also a seismically active developing boundary through the central valley, connecting to the back arc thrusting [Protti et al., 1995a], and there is evidence of general right-lateral shearing associated with oblique plate convergence [Lundgren et al., 1999].

These tectonic processes can account for most of the differences in the intracrustal seismic activity between the NW and SE parts of the CRI, which are especially notable at upper and middle crustal levels (see Figures 5a and 5c). However, we suggest that at least part of the upper crustal seismic activity in the NW part of the CRI could also be a consequence of magmatic processes related to the development of the volcanic arc, which are currently lacking in the SE. These processes would include magma transport through the crust, emplacement of magmatic intrusions, and readjustment of material induced by phase changes (crystallization of magmas) [see, e.g., Shaw, 1980; Chouet, 1996].

It is more difficult to explain the origin of the lower crustal seismic activity (Figure 16). The existence of a brittle lower crust would require a mafic composition and a low heat flow [Lamontagne and Ranalli, 1996]. The first condition agrees with the estimated composition of the CRI lowermost crust (Figure 15), but the second one seems to conflict with the repeated intrusion of magmas and with the presence of an active volcanic arc. Therefore, this interpretation remains speculative.

According to our results, we suggest that the “Moho” of the CRI (M2) does not represent a first-order boundary between two levels of different origin and composition, but it shows more likely the maximum depth at which the underplated magmas begin to crystallize. Thus the lowermost part of the crust could represent a “transition zone”, constituted by layered sequences of mafic residues and solidified melting products with intrusion of mantle-derived igneous stratiform bodies [Meissner and Kuszniir, 1987]. Such a transition zone would also explain the scattering of the deep seismic phases (P, D, S, and P, S) that are observed in the seismic records (Figures 7d and 7e) to be the result of superposition of seismic reflections in the distinct layers within the transition zone [Holliger et al., 1993].

3.8. Crustal Domains Within the CRI

The results from the simultaneous tomographic inversion allow us to distinguish at least two crustal domains within the CRI, which are separated by the SW-NE seismic alignment (Figure 5a). The NW domain corresponds to the area in which the Cocos plate subducts under the Nicoya Peninsula and the Quepos segment. As we stated above, this subduction zone has a dip angle of ~35°, with associated seismicity to more than 150 km depth [Protti et al., 1995b], and shows an active volcanic arc. Moreover, the 3-D velocity field inferred from seismic tomography is highly heterogeneous, being characterized by conspicuous velocity anomalies of intermediate scale, and shows significant intracrustal seismicity (Figures 5c and 5d).

The SE domain corresponds to the subduction of the Cocos Ridge. This part is characterized by the subduction of an overthickened oceanic crust [StevensJen et al., 1998], having a dip angle of 15-20°, with subduction-related seismicity to <70 km depth and shows a volcanic gap. The 3-D velocity field is much more homogeneous than that of the northern half, and it shows scarce intracrustal seismicity (Figures 5c and 5d).

These observations seem to indicate that the differences in the structure and geometry of the subducting plate between NW and SE Costa Rica may play an important role to determine the crustal structure, arc magmatism, intracrustal seismicity, and heterogeneities in the velocity field of the overlying plate. The subduction of young, overthickened and thus more buoyant plates, such as the Cocos Ridge beneath SE Costa Rica, is characterized by low subduction angles [Protti et al., 1995b]. This implies that the subduction zone is generally shallower (e.g., ~70 km in SE Costa Rica); thus it is difficult to reach conditions adequate to dehydrate the crust and trigger partial melting reactions within the mantle wedge. These processes are estimated to occur deeper than 80-100 km depth [Wilson, 1989; Peacock and Hyndman, 1999]. Therefore subduction-related magmas are unlikely to form in SE Costa Rica, in contrast to the NW part. This situation could account for the gap in the volcanic arc, the lack of magmatic intrusions (homogeneous velocity field), and, in turn, the lower seismic activity within the upper crustal levels in SE Costa Rica.

4. Conclusions

The results of this multidisciplinary study have allowed a description of the structure and physical properties (v, p) of the CRI lithosphere and its convergent margin. According to our results the CRI crust is ~40 km thick in its NW half. It can be divided into four internal levels, which have been built up in three periods of evolution.

The first period took place about 150 Ma, and corresponds to
the generation of a normal oceanic crust in an oceanic ridge located at the southeastern Pacific [Hull, 1995]. The second period initiated at 90 Ma and corresponds to the fast extrusion of a huge volume of igneous material, which intruded and thickened the preexisting oceanic lithosphere [Findell and Barrett, 1990]. This interval is thought to represent the head phase of the mantle plume which originated in a later stage the Galápagos hotspot [Duncan and Hargraves, 1984]. These two initial periods of growth built up the Caribbean plateau as a large igneous province with ~15–20 km of averaged crustal thickness and an extension of ~10^3 km.

We suggest that the basement of the CRI constitutes the westernmost edge of the Caribbean plateau. The upper level (~2–8 km) would then correspond to the extrusive basement, and the second level (~8–20 km) would represent the preexisting oceanic crust, which was modified and thickened by the extrusion of flood basalts in the Late Cretaceous. Accreted or recycled slope sediments formed the first 8–10 km of the marginal wedge. The rest of the wedge (~65 km) is more likely an offshore extension of the extrusive basement, and thus it would be composed mainly by fractured or altered flood basalts.

The third period of evolution initiated about 80 Ma, and we suggest that it is related with the subduction zone magmatism. The fluids released by processes of metamorphic dehydration of the subducted lithosphere produced partial melting reactions within the mantle wedge above the subducted slab. These magmas are underplated at the base of the overlying plate, as revealed by the low velocities and densities of the upper mantle. Another process contributing to these low velocities and densities is the presence of water and the formation of hydrous minerals (i.e., serpentinite) within the mantle wedge by fluids released from the subducting plate [Peacock and Hyndman, 1999]. The underplated magmas can easily intrude the crust, crystallizing and forming mafic rock types at or near the crust-mantle boundary, which are likely to build up the lower crustal levels [Rudnick and Jackson, 1995].

Emplacement of migmatic intrusions within the crust could also account for intracrustal heterogeneities that are revealed by seismic tomography and for part of the upper crustal seismicity in the NW part of the isthmus. The lighter components of the magmas are erupted in the volcanic arc, building up the volcanic edifice. This scheme of crustal evolution could explain some of the differences in the crustal structure, velocity field and magmatic and seismic activity between the NW and SE parts of the CRI in terms of the differences in the crustal structure of the subducting plate and in the geometry of subduction.

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