Tectonic erosion and consequent collapse of the Pacific margin of Costa Rica: Combined implications from ODP Leg 170, seismic offshore data, and regional geology of the Nicoya Peninsula

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Abstract. The convergent margin off the Pacific coast of the Nicoya Peninsula of Costa Rica exhibits evidence for subduction erosion caused by the underthrusting Cocos plate. Critical evidence for efficacy of this process was recovered at the Ocean Drilling Program (ODP) drilling Site 1042 (Leg 170), positioned ~7 km landward of the Middle America trench axis off the Nicoya Peninsula. The primary drilling objective at this site was to identify the age and origin of a regionally extensive and prominent seismic discontinuity, the so-called base-of-slope sediment (BOSS) horizon or surface. The BOSS horizon, which can be traced landward from near the trench to the Nicoya coastal area and parallel to it for hundreds of kilometers, separates a low-velocity (~2.0-2.5 km s⁻¹) sequence of slope sediment, from an underlying sequence of much higher-velocity (~4.4-5 km s⁻¹) rock. Site 1042 reached the acoustically defined BOSS horizon at a below sea level depth of ~3900 m and yielded a carbonate-cemented calcarenitic breccia of early-middle Miocene age. Sedimentological, geochemical, paleontological, and cement paragenesis data document that the breccia accumulated in a shallow water depositional environment. On the basis of coastal exposures, the BOSS horizon, as a margin-wide geologic interface, can be temporally and lithostratigraphically correlated to a regional angular unconformity. This unconformity, known as the Mal Pals unconformity, separates Neogene and younger shelf-to-littoral beds from the underlying mafic units of the Mesozoic Nicoya Complex and Cretaceous and early Tertiary sedimentary sequences. At Site 1042 it is inferred that tectonism caused the vertical subsidence of the early Neogene breccia from a shallow to a deep water setting. The Mal Pals unconformity of the BOSS horizon thus connects the rock fabric of the outermost part of margin to that of coastal Nicoya and implies that in the early Neogene the Nicoya shelf extended seaward to near the present trench axis. This circumstance requires that the early Neogene trench axis was at least 50 km seaward of where it is now located. The long-term effects of subduction erosion, similar to those described for the scientifically drilled Japan, Tonga, and Peru margins, best account for offshore and onshore evidence for a post-Paleogene history of crustal thinning and landward trench migration of Costa Rica's Pacific margin. During the past 16-17 Myr the calculated mass removal and landward migration rates are 34-36 km³ Myr⁻¹ km⁻¹ of margin, and 3 km Myr⁻¹, respectively. These values are similar to those found for other Pacific margins dominated by nonaccretionary subduction zone processes.

1. Introduction

The Pacific margin of Costa Rica has long been considered a favorable place to study processes of subduction accretion (Figure 1). In part for this reason a large quantity of geophysical data have been gathered in this offshore region. On the basis of this information, and Deep Sea Drilling Project (DSDP) Leg 67 and 84 drilling results, two dominant, but very different, models have emerged during the past decade regarding the structure of the Costa Rica margin and the tectonic mechanisms shaping its form and evolution [Shipley et al., 1990, 1992; Von Huene and Flueh, 1994; Hinz et al., 1996; Ye et al., 1996]. One model ascribes margin evolution to subduction accretion, in particular to massive underplating of oceanic sediment since the early Tertiary. A contrasting or nonaccretionary model contends that the margin is mostly underlain by Costa Rican basement rocks (i.e., chiefly mafic ocean crustal units of Cretaceous age), which extend seaward from coastal outcrops to near the trench axis. Figure 2, adapted from Von Huene and Flueh [1994], diagrammatically contrasts the accretionary and nonaccretionary models. It is important to recognize that for Costa Rica the accretionary and nonaccretionary models are based on the interpretations of seismic reflection and refraction information; that is, they are effectively unconstrained by geologic observations. As emphasized below, this paper focuses new geologic information on testing the viability of the nonaccretionary model.

A major scientific objective of Ocean Drilling Program (ODP) Leg 170 was thus to supply geologic information off the Nicoya Peninsula of Costa Rica. Drilling was targeted along a traverse from the trench to the midshelf area. Beyond drilling to investigate the relative importance of different tectonic mechanisms forming the margin's rock and sediment framework, Leg 170 science also focused on mass balance issues dealing with the volume of solid and fluid material added to, subtracted from, or that bypassed the margin during much of Cenozoic time [Kimura et al., 1997].
The initial attempt to investigate the geologic history of the Costa Rica margin was at DSDP Site 556, Leg 84. This site was drilled in response to the findings of the transect of holes drilled by DSDP Leg 67 across the Guatemala margin northwest of Costa Rica [Von Huene et al., 1980]. The objective of Site 556 drilling, located on the middle slope, was to reach a prominent, subsurface reflection horizon, variably called the rough, smooth, or base-of-slope sediment (BOSS) surface (Figure 3). At the time of the drilling the surface was, on geophysical grounds, thought to be either the top of a subduction complex buried beneath a 500-1000-m thick apron of slope deposits, or a regional unconformity beneath which Costa Rican basement rock would be encountered. Unfortunately, drilling did not penetrate the overlying apron [Von Huene et al., 1985]. A decade later, in 1994, visual data were collected during 20 Alvin dives exploring for fluid vents [Silver, 1996]. Thus, in effect, geological data resulting from Leg 170 drilling are among the first to provide physical observations about the nature of the BOSS horizon and thus to decipher the geologic evolution of the Costa Rica margin. Crucial information about the nature of the BOSS horizon was recovered near the bottom of Site 1042, which, located near the base of the landward trench slope, was sited to penetrate this surface.

The existence of the BOSS horizon has been cited as evidence for subduction erosion [Vannucchi et al., 1998; Meschede et al., 1999]. In this paper we assemble and interpret the geologic and tectonic implications of a combination of newly acquired Leg 170 findings and onshore mapping and rock information. These data sets are focused to test the viability of the nonaccretionary model, which can be approached because this model predicts that the margin's rock framework or wedge underlying the BOSS horizon is exposed along the Nicoya coast and extends seaward to near the base of the slope. We conclude, on the sedimentary, structural, and micropaleontological evidence described and interpreted in this paper, that the predictions of the nonaccretionary model are consistent with our findings. With respect to the larger implications of our conclusions the Cenozoic evolution of the Costa Rican margin can be viewed as a manifestation of the tectonic processes of subduction erosion and sediment subduction.

2. Tectonic Setting and Previous Geophysical Studies

The Nicoya Peninsula, which rises above the leading or western edge of the Caribbean plate, is underthrust at a velocity of ~90 mm yr^-1 (90 km myr^-1) by the eastward subducting Cocos plate [DeMets et al., 1990]. A detailed bathymetric swath map of the Pacific side of Costa Rica reveals that relief of the incoming Cocos plate determines
much of the morphology of the continental margin [Von Huene et al., 1995, Fisher et al., 1998; Von Huene et al., 2000] (Figure 1). Southeast of the Nicoya Peninsula the Cocos plate is populated by seamounts, that inboard of the Middle America trench (MAT) penetrate the lower trench slope and leave subduction tracks appearing as embayments [Von Huene and Scholl, 1991, Von Huene et al., 1995, Fisher et al., 1998] (Figure 1). The topography of the Cocos plate seaward of the Nicoya Peninsula is, in contrast, rather smooth, and the corresponding Costa Rican continental margin has a nearly uniform sedimentary cover [Mcintosh et al., 1993]. Along the Nicoya margin slumps and other gravitational processes erode the middle and upper slope and drive materials to the lower slope and onto the trench floor [Fisher et al., 1998; Von Huene and Ranero, 1998] (Figure 1).  

A number of seismic investigations of the Nicoya offshore have been completed during the past decade. Stoffers et al. [1991] and Shipley et al. [1992] first conducted a three-dimensional (3-D) seismic survey in this area. Figure 3 shows a 2-D seismic reflection profile off the Nicoya Peninsula. In this profile the margin's subsurface velocity structure is well constrained from -20 km landward of the trench axis to the coastal area (Figure 2). Four velocity domains can be distinguished: that of the lower or incoming Cocos plate, the lowermost continental slope, below the BOSS horizon, the inner mass of the margin of the middle-upper continental slope, and, above this horizon, the apron of slope sediment [Ye et al., 1996]. The incoming Cocos plate, which, at the location of Leg 170, was generated at the East Pacific Rise [Lonsdale and Klüg, 1978, Meschede et al., 1998; Von Huene et al.,
Beneath the region of the middle to upper continental slope, at a water depth range of 3500-1000 m and from 55 to 25 km from the shoreline, three main sedimentary and rock units are recognized on seismic reflection and refraction profiles (Figures 2 and 3). From top to bottom these are (1) the sedimentary deposits of the slope apron, (2) the landward thickening, wedge-shaped unit below the BOSS horizon that forms the margin's forearc basement, and (3) beneath the base of the margin, the underthrusting surface of the oceanic slab. Beneath the middle slope, poor reflectivity and incoherent stratification characterize the apron sequence, but below the upper slope lateral continuity of internal reflectivity, i.e., stratification, is well expressed. The thickness of the apron sequence ranges from 0.5 to 2.6 km, thickening landward such that the apron of slope sediment beneath the upper slope is more than twice that of the lower slope. A bottom-simulating reflector (BSR) is present at a subsurface depth of ~150 m (Figure 3).

As previously emphasized, the lower boundary of the apron sequence is coincident with the strong reflector called BOSS horizon or surface, beneath which is the forearc basement wedge. The basal sediment of the slope apron...
drapes the BOSS horizon, and locally it is offset by a few tens of meters of reverse and normal faults [Meschede et al., 1999]. Stratification within the apron sequence, as noted, is weak, and correlations from the upper slope to the lower slope are unreliable. Locally, the slope apron sediment onlaps the BOSS surface [Hinz et al., 1997], documenting a downslope, time-transgressive burial of the BOSS horizon.

The BOSS horizon marks an abrupt change in velocity from 2.0-2.5 km s⁻¹ for lower beds of the slope sediment to 4.5 km s⁻¹ and higher, with a maximum of 5.9 km s⁻¹, in the underlying forearc basement wedge (Figure 2). Below the lower and middle slope the surface of the BOSS horizon exhibits an irregular morphology, but beneath the upper slope the profile of the BOSS surface changes to a smooth, continuous surface (Figure 3). What is important about the BOSS horizon is that it can be traced landward to near the Nicoya coast and parallel to it along much of Costa Rica [Barboza and Zucchi, 1994; Hinz et al., 1996] and northward off at least southern Nicaragua [Ranero et al., 2000].

The nature of the forearc basement wedge underlying the BOSS reflector has been a matter of debate for decades. The interpretation of the high-velocity and indistinct structures imaged on seismic profiles is the focal point of the controversy. Shipley et al. [1992] and Hinz et al. [1996] agreed that below the middle slope the main fabric of the forearc basement wedge consists of landward dipping horizons that generally do not offset the BOSS interface. Beneath the upper slope they recognized that landward the internal reflectors become more gently dipping approaching horizontal attitudes. Shipley et al. [1992] interpret the seaward part of the interior wedge as an accretionary structure, constructed of larger coherent units separated by landward dipping, low-angle thrusts. This hypothesis states that packets of offscraped and underplated oceanic sediment, now highly consolidated, form the margin's interior wedge, with accretionary buildup of the forearc basement wedge since the Eocene (Figure 2). Accretionary buildup of the forearc basement wedge has been under way since the Eocene. Von Huene and Flash [1994], Hinz et al. [1996] and Ye et al. [1996] confuse the accretionary processes (i.e., the tectonic addition of lower plate material to the upper plate) to the lowermost slope of the margin. More importantly, they consider the bulk of the forearc basement wedge to be the offshore extension of the so-called Nicoya Ophiolite Complex. This effectively nonaccretionary model explains the few indistinct landward dipping reflectors beneath the middle-upper slope as part of the basement's fabric of rock and structural units, and it explains the BOSS surface as a major unconformity separating two very different stratigraphic sequences. More recently, enhanced seismic processing has imaged thick (1-2 km) lenticular bodies along the interplate surface that are interpreted by Ranero and Von Huene [2000] as displaying upper plate material being transported downdip by processes of subduction erosion.

3. Leg 170 Results
3.1. Lower Slope
Leg 170 drilled five sites (Figure 1): one reference site on the subducting Cocos plate (Site 1039); two sites through the toe of the landward trench slope, the décollement, the underthrust sedimentary section, and into the underlying igneous oceanic crust (Sites 1040 and 1043); and two mid-to-lower slope sites (Sites 1041 and 1042). The reference Cocos plate section at Site 1039 and the comparison underthrust section at Site 1043 document that no more than 5-10 m, if any at all, of the top of the underthrusting reference section are missing at Site 1043. This observation implies either a negligible to zero volume of frontal accretion, and that sediment recovered at the toe of the lower slope is mass-wasted debris from the lower slope [Kimura et al., 1997; Vannucchi and Tobin, 2000]. It is possible that erosion of the uppermost sediment by trench axis bottom currents occurred prior to underthrusting [Silver et al., 1997]. However, it is evident that the décollement, which is an overpressured horizon documented by its negative polarity reflection (Figure 3), allows the lower plate to slip with little friction beneath the upper plate of the continental margin.

The discovery that, effectively, frontal accretion does not occur along the drilling transect has important meaning for the observation that inboard of the Nicoya Peninsula, ⁹⁰⁷Be is not detected in young arc eruptive rocks, although this radiogenic isotope, which indicates subduction of young ocean sediment to the depth (~ 100 km) of arc magmatism, occurs in abundance in sediment entering the Costa Rica subduction zone [Valentine et al., 1997].

3.2. Stratigraphy of Middle Slope Sites
ODP Site 1041 was drilled into the middle slope of the Nicoya margin to reach, sample, and determine the geologic nature of the BOSS surface and to determine the stratigraphy of the overlying sedimentary apron. Unfortunately, after a good recovery of the slope apron sediment, because of poor hole conditions, the BOSS horizon (~550 m below sea floor (mbsf)) and underlying rock units of the forearc basement wedge could not be reached. Hence site 1042 was drilled near the BOSS horizon oceanward closure (~7 km landward from the trench), where it was expected at only 300 mbsf (Figure 3). At this location the setting of BOSS horizon can be distinguished by the termination of the landward thickening, wedge-shaped unit that forms the margin's forearc basement, and, in fact, a thrust surface was drilled [Kimura et al., 1997]. The slope apron sequence at site 1042 was cored every 50 m. Thus Site 1041 is the reference site for the apron section (Figure 4). Site 565 of DSDP Leg 84 was devoted to the same objectives as Sites 1041 and 1042, and, even though Site 565 did not reach the basement, a continuous apron sequence of latest Miocene to Quaternary age was recovered [Von Huene et al., 1985]. At Leg 170's Site 1041, the complete Pleistocene to Miocene slope apron sequence was penetrated (Figure 4). The sequence consists of olive green claystone and siltstone with minor sandstone, limestone, and volcanic ash (Figure 5a). A relative abundance of coarser material, medium to coarse sandstone containing granule to pebble-sized clasts of claystone, is present in the lower part of the unit. Ash layers are common in the upper part of the apron sequence, and granules of volcanic glass are common throughout this section. Apron sedimentation seems to have been characterized by episodes of mass flow intercalated with accumulation of pelagic/hemipelagic sediment, especially in the upper part of the column. The cores are generally of poor quality, with intense drilling disturbance. Also, dissociation of
gas hydrate, found in the interval between 116 and 184 mbsf, contributes further to core disturbance. Although complicated by these disrupting factors evidence for low-grade tectonic deformation was only recognized as microfaults, minor fractures and changes in bedding dip. This evidence for tectonism is concentrated in two intervals where two faults cutting through the apron sequence are suspected.

Benthic foraminiferal assemblages identified in Site 1041 (S. Hasegawa, unpublished data, 1998) indicate a general shallowing trend from upper middle bathyal depths (500-1500 m) to...
m) in the middle Miocene to abyssal depths (>4000 m) in the Pleistocene. The Hole C (samples from 405 mbsf to 424 mbsf) and the lower part of Hole B (from 280 mbsf to 386 mbsf) are dominated by species which have upper depth limits in the upper middle bathyal (500-1500 m) and upper bathyal (150-500 m) biofacies as defined by Ingle [1980]. These species include Cassidulina laevigata carinata, Epistominella exiqua, E. californica, E. pacifica, and Miogypsina, Archaias, and Amphistegina have been recognized among the benthic foraminifera (E. Robinson, written communication, 2000). Approximately 5% to 25% of the calcarenite debris are volcaniclastic components, mostly plagioclase feldspar, pyroxene, amphibole, and volcanic fragments, and <2% are siliciclastic components and glauconite (percentages were calculated by N. Lindsay-Griffith, written communication, 1998). All the components are unaltered and thus fresh in appearance. Angular claystone clasts were also observed.

The calcarenite clasts constituting the breccia are well cemented without showing evidence of compaction. Sedimentary structures, such as laminations and vadose silt filling of cavities (Figure 6a), are preserved in the clasts. Petrographic, X-ray, and isotopic analyses were used to identify the depositional and cementation environment of the clasts. The cement is composed both of fibrous to bladed calcite crystals grown around the clasts and of dirty micrite filling the pores (Figure 6a). The original micrite of some of the breccia fragments is completely recrystallized to a neomorphic sparite. Polygonal sutures are not observable in the pore spaces. Pellets and microfossil chambers in the clasts commonly bear brownish to green glauconite (Figure 6b). Rarely, glauconite is present as intergranular cement.

Secondary porosity is widespread, and calcite filling of fractures (veins) took place after primary cementation of the calcarenite. These calcite veins are preserved as prebrecciation structures together with microfaults and mud injections (Figure 6c). The calcite veins are constructed of syntaxial (grown from the edges to the center of the vein) fibrous crystals that are strain free except for rare twinning on the e planes, a pattern that reflects growth twinning [Laurent et al., 1990]. Veins do not exhibit a preferred orientation, even though truncations are observed. The only consistent relation is that thicker veins, up to 1 cm, are younger than thinner veins of submillimeter width. The thickest veins are constructed of drusy calcite that incompletely fills the void (Figure 6d). The structures in the clasts lost their persistency because of the brecciation, and few postbrecciation structures cut through both clasts and matrix.

The matrix of the carbonate breccia is composed of carbonate mudstone containing microfossils and bioclasts.
Patchy calcite aggregates and common glauconite testify to a rudimentary cementation process (Figure 6e). Syntactical veins and healed fractures are common and preferentially located at the boundary between clasts and matrix, or they occur entirely within the matrix (Figure 6e). Veins located at the boundaries of the clasts exhibit asymmetrical growth: They have a syntactical edge linked to the clast wall and they show euhedral crystal terminations in the matrix (Figure 6e). Boundary veins are also present at the edges of clayey clasts. Cathodoluminescence observations were made to distinguish calcite generations on the basis of trace elements distribution, particularly Mn and Fe, the principal activator and quencher, respectively, in calcite. The observations carried out on the cement and veins show low luminescence relative to all other carbonate phases.

The breccia has been dated using both planktonic and benthonic foraminifera. The planktonic foraminifera in sediment immediately overlying the breccia are assigned to the N8-N10 zone, i.e., ~12.6–16.4 Ma by Ibaraki (2000). On the basis of his study of thin sections of the breccia, E. Robinson (written communication, 2000) estimates a similar age for the limestone. To more definitely date the carbonate-
the lithologic components show a medium to high degree of alteration. The clasts constitute about 40-50% of the breccia; glauconite and olivine occur in trace amounts (Figure 7). All minerals and amoeboid crystals (Figure 9). In both examples they exhibit both ribbon-shaped crystals alternating with clay and a minor amount of calcite occurs within the matrix (Figures 8a and 8b). The shapes of the clasts are various, but the serpentinized mafic rocks are spherical. The size of the clasts ranges from submillimeters up to 5 cm.

The clasts of the chert-basalt breccia contain prebrecciation structures such as laminations in the sedimentary components and common veins. Veins are quartz in the chert, calcite in the limestone, and zeolite in the volcanic fragments. Zeolite veins exhibit both ribbon-shaped crystals alternating with clay minerals and amoeboid crystals (Figure 9). In both examples the zeolite is clinoptilolite. Veins are present also in the matrix within which a complex history of mineralization can be traced. All of the carbonate veins sampled in the chert-basalt breccia can be classified into four types based on their morphology, structural association with the host sediment, and intersections (Figure 9). Type 1 veins occur in all the examined samples and cross-cutting relations document they are the older veins (Figure 9). Type 1 veins are constructed of micritic mosaic calcite incorporating clasts of the breccia, they are anastomosing, up to 0.1 mm thick, and have diffuse edges. These veins have no preferred orientation, and where they surround clasts their thickness is greatest and the calcite becomes clean and fibrous. Type 1 veins resemble cement precipitation in irregular fractures, implying an early origin.

Type 2 veins also occur in all the examined samples; they consist of dirty, pervasive veinlets with mosaic and antitaxial calcite (Figure 9). These veins do not show a preferred orientation. Type 3 veins are zeolite, calcite, or composite calcite-zeolite veins that cut through the type 1 and type 2 veins as straight fracture fillings (Figure 9). Type 3 veins are mosaic or fibrous in texture, but always extensional with no pressure-solution seams. A type 4 vein occurs in one sample at 343.39 mbsf. This vein consists of fibrous calcite with extensional, antitaxial growth (Figure 9). It is anastomosing, 0.5 mm thick, and perpendicular to the top of the core.

### Table 1. Sr Isotope Ratio in Samples from Core 1042B 1R, Section 2W

<table>
<thead>
<tr>
<th>Interval, cm</th>
<th>Depth, mbsf</th>
<th>Sr isotope ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>10-13</td>
<td>317.17</td>
<td>0.708740±2x10⁻⁵</td>
</tr>
<tr>
<td>80-87</td>
<td>317.85</td>
<td>0.708715±2x10⁻⁵</td>
</tr>
<tr>
<td>91-94</td>
<td>317.98</td>
<td>0.708699±2x10⁻⁵</td>
</tr>
</tbody>
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*Here mbsf is meters below sea floor

Cemented breccia, three samples were analyzed for $^{87}$Sr/$^{86}$Sr ratio by A. Paytan (written communication, 2000). The average ratio is 0.70872, which corresponds to an age of 16-17 Ma, the top of the early Miocene [Paytan et al., 1993], the age of the samples increases slightly with depth (Table 1).

Transition sediment to subunit 2B, the chert-basalt breccia, was not recovered; hence the change is stratigraphically abrupt. Subunit 2B is a polymictic breccia of clasts of doleritic basalt, palagonite, devitrified pumice, red and white chert, claystone, siltstone and sandstone, serpentinized mafic, and quartz-mica schist. Clasts of micritic limestone, feldspar, glauconite and olivine occur in trace amounts (Figure 7). All the lithologic components show a medium to high degree of alteration. The clasts constitute about 40-50% of the breccia; they are tightly packed but rarely in contact. Brecciated clasts are also present. The matrix is colored green by chloritized clay, and a minor amount of calcite occurs within the matrix (Figures 8a and 8b). The chert-basalt breccia is a poorly sorted aggregation of angular and subangular clasts (Figure 8b). The shapes of the clasts are various, but the serpentinized mafic rocks are spherical. The size of the clasts ranges from submillimeters up to 5 cm.

Veins are quartz in the chert, calcite in the limestone, and zeolite in the volcanic fragments. Zeolite veins exhibit both ribbon-shaped crystals alternating with clay minerals and amoeboid crystals (Figure 9). In both examples the zeolite is clinoptilolite. Veins are present also in the matrix within which a complex history of mineralization can be traced. All of the carbonate veins sampled in the chert-basalt breccia can be classified into four types based on their morphology, structural association with the host sediment, and intersections (Figure 9). Type 1 veins occur in all the examined samples and cross-cutting relations document they are the older veins (Figure 9). Type 1 veins are constructed of micritic mosaic calcite incorporating clasts of the breccia, they are anastomosing, up to 0.1 mm thick, and have diffuse edges. These veins have no preferred orientation, and where they surround clasts their thickness is greatest and the calcite becomes clean and fibrous. Type 1 veins resemble cement precipitation in irregular fractures, implying an early origin.

Type 2 veins also occur in all the examined samples; they consist of dirty, pervasive veinlets with mosaic and antitaxial calcite (Figure 9). These veins do not show a preferred orientation. Type 3 veins are zeolite, calcite, or composite calcite-zeolite veins that cut through the type 1 and type 2 veins as straight fracture fillings (Figure 9). Type 3 veins are mosaic or fibrous in texture, but always extensional with no pressure-solution seams. A type 4 vein occurs in one sample at 343.39 mbsf. This vein consists of fibrous calcite with extensional, antitaxial growth (Figure 9). It is anastomosing, 0.5 mm thick, and perpendicular to the top of the core.

### 4. Discussion and Interpretations of Offshore Data

Site 1042, positioned ~7 km landward of the trench axis off the Nicoya Peninsula of Costa Rica, recovered a calcareous sedimentary breccia at a depth of ~3900 m. The breccia is dated paleontologically as just older than ~15 Ma, and directly via its strontium isotope ratio at 16-17 Ma. The breccia is representative of a basal well-cemented unit of the...
apron sequence of slope sediment and its velocity is high, 3.5-
4.0 km s\(^{-1}\). Thus, from a velocity perspective the basal breccia beds are part of the prominent velocity jump represented by the acoustically defined BOSS horizon. However, the basal breccia units are temporally congruent with the overlying beds of the apron. As a consequence, and in our view, the greater geologic significance of the BOSS horizon, which underlies virtually the width and breadth of the Costa Rica margin, is that it is a regionally extensive unconformity separating rock units of contrasting lithology and age. Drilling at Site 1042 thus penetrated into the top of the acoustic BOSS horizon but not the more important geological or age-contrasting interface included within the velocity transition.

The carbonate-cemented breccia contains neritic shell fragments, all well preserved. The Archaias and the Amphistegina are abundant in the sediment and they are both shallow inner to middle shelf genera at the present day and frequent components of carbonate beach sands (E. Robinson, written communication, 2000). The sedimentary suite of the clasts of the carbonate-cemented limestone breccia includes fine-laminated, unbioturbated and well-sorted calcarenite, claystone chips, sediment injections, and a rich photic-zone fauna. These sedimentary characters are associated with thorough lithification. DSDP and ODP samples testify that deep-sea carbonate units are weakly indurated except for Cretaceous and Jurassic sediment [Larson et al., 1992]. Moreover, deep-sea carbonate beds display evidence of widespread dissolution as the main diagenetic process. In contrast, the carbonate-cemented breccia recovered at Site 1042 is 16-17 Myr old, and dissolution is not observed. The cement paragenesis in the carbonate-cemented breccia, with fibrous to bladed calcite surrounding the grains and micrite filling the pores, suggests the circulation of waters supersaturated with respect to CaCO\(_3\) and vigorous fluid fluxing in order to maximize CO\(_2\) degassing. High-energy settings are favorable places where both the sedimentary facies and cementation processes displayed by the 1042 calcarenite characteristically form. This recognition is consistent with the observation that the original calcarenite of the breccia clasts, although both porous and permeable, exhibits few inclusions of fine material, such as clay particles. High-energy environments are associated with oxidizing conditions that, precluding incorporation of Fe and Mn in the carbonate lattice, provide a good explanation for the lack of cathodoluminescence of the breccia carbonate grains [Moore, 1989].

At least three generations of calcite veins testify to fracture of the calcarenite soon after cementation, but prior to its brecciation. Other fractures in the clasts are produced by sediment injection. These fractures reveal interplay of overpressuring and displacement operating in the calcarenite.

The sedimentary suite and the kinetic factors necessary for cementation of the clasts limit their depositional environment to shallow marine conditions. Distinguishing the many possible settings is, however, difficult. The intertidal-shorefaced zone, in association with beach sedimentation and the shelf margin are both possible habitats for the Archaias and the Amphistegina and are high-energy places where deposition and cementation of the calcarenite could have taken place. Brecciation also occurred in an environment where high-energy cementation was able to operate, as indicated by the syntaxial asymmetric veins developed at the grain matrix boundaries and the patchy calcite aggregates in the matrix. The angular boundaries of the clasts, the intrabasinal nature of the breccia, and the lack of exotic, younger fauna suggest a low level of reworking and transportation of the clasts after they were formed. Also, the Sr isotope ratio data indicate the carbonate-cemented breccia gets older downward (Table 1). This is an evidence that the section is stratigraphically ordered and thus most likely intact. The age progression is also not consistent with the possibility that the breccia section is part of a slide mass of nearshore deposits that ended up in deep water. The high-energy settings invoked for the deposition and cementation of the calcarenite forming the clasts may well have produced the brecciation as well. Processes that can disrupt lithified bands are associated with shoreface or shelf margin environments, where storm action or unstable slopes can brecciate beds of calcarenite deposits.

Our stratigraphic, sedimentological, cementation, and rock
| Type 1 | Present in all the samples, they are the older veins.  
Micritic mosaic calcite incorporating clasts of the breccia, up to 0.1 mm thick with undefined edges. The calcite is clean and fibrous around clasts.  
No preferred orientations, anastomosing.  
Type 1 veins resemble cement precipitation in irregular fractures implying an early origin. |
|---|---|
| Type 2 | Present in all the samples.  
Dirty, pervasive veinlets with mosaic and antitaxial calcite.  
No preferred orientation. |
| Type 3 | Zeolite, calcite or composite calcite-zeolite veins cutting through the former ones as straight fractures.  
Mosaic or fibrous calcite, but always extensional with no pressure-solution seams. |
| Type 4 | Only in one sample at 343.39 mbsf.  
Fibrous calcite with extensional, antitaxial growth.  
Anastomosing, 0.5 mm thick, perpendicular to the top of the core.  
Around the vein there are clasts with evidence of shear.  
Type 4 vein is the only one with a clear tectonic origin. |

![Figure 9. Synoptic diagram of textures and relationships of veins in the chert-basalt breccia.](image)

structure observations are incompatible with downslope transportation of the breccia from shallow to deeper water, and in particular to a recovery depth that would have been below the Miocene calcite lysocline where interclast calcite cementation could not have taken place. Related evidence that distant downslope transport of the calcarenite clasts did not take place includes (1) cementation was active after brecciation, (2) hemipelagic and pelagic sediment are not intercalated with the breccia, (3) the pronounced angularity of the clasts, and (4) the lack of tractive features in the matrix and size gradation of the clasts.

The occurrence of common glauconite implies a low to restricted sedimentation rate or restricted sediment movement. For glauconite to form, in fact, the sediment-water interface must be maintained for an appropriate length of time to assure stabilization. From the samples examined it is not possible to
Figure 10. Photomicrograph of the well-indurated silty-clay recovered at 343.42 mbsf of Site 1042 (parallel nicols). Hydraulic brecciation and shearing predate vein formation. The older mineral precipitation consists of quartz pods, cut by extensional calcite veins perpendicular to the top of the core. The younger veins are calcite, quartz, and composite calcite-quartz bearing, and they occur with no preferred orientation and crosscutting relationship.

determine if glauconite formed prior to or after cementation, but the mineral has either a brown limonitic or green appearance, suggesting different degrees of oxidation. Because glauconite is readily weathered, it is not likely that individual grains were transported under subaerial conditions, but a marine reworking may well have occurred.

The scenario envisaged by the diagenetic features of the carbonate-cemented breccia is thus that of a shallow water, high-energy environment, where the deposited and cemented bioclastic sediment could be brecciated and redeposited in the same place.

The passage from the carbonate-cemented breccia to the underlying chert-basalt breccia marks a prominent change from bioclastic to coarse detrital. The clasts of the chert-basalt breccia define an assemblage of material derived from outcrops of well-lithified source or parental rocks that includes chert, limestone, sandstone, doleritic basalt, and serpentinized mafic units. The source rock sequence of the chert-basalt breccia is similar to that of an oceanic or ophiolitic assemblage of igneous basement units and overlying sedimentary cover. The basalt clasts have zeolite veins implying ocean floor alteration and fluid circulation, while the brecciated clasts and the quartz-mica schists indicate tectonic deformation and metamorphism. On the Nicoya Peninsula a suitable source sequence, the ophiolitic Nicoya Complex, is widely exposed [Dengo, 1962; Baumgartner et al., 1984]. The Nicoya Complex, possibly a fragment of an oceanic plateau, is largely of Cretaceous age [Sinton et al., 1997], although chert masses of Jurassic age are reported.

The chert-basalt breccia recovered at Site 1042 is very immature both in lithology and in texture: The clasts do not exhibit evidence of distant transport. The clay-rich matrix and the numerous veins argue for an undrained deposit with episodes of high fluid pressure. Mud and sand injections and type 1 veins suggest escape paths as consequence of fast compaction. The described features are compatible with a debris flow mechanism where the sediment mobilization was helped by the fluid overpressure that allowed mass movement of material even on a low dipping plane. The silty clay-

Figure 11. (a) Site 1042 shear zone at the base of the chert-basalt breccia. (b) Photomicrograph of the shear zone fabric.
indurated horizons recovered as interlayers in the chert-basalt breccia confirm the presence of high fluid pressure and help define the sedimentary environment. In fact, even though the nature of the silty clay level is difficult to infer, this band suggests pelagic sedimentation following events that replaced the chert-basalt breccia. The presence of this fine-grained intercalation within the chert-basalt breccia suggests a submarine environment located between the proximal and the intermediate slope, where pelagic, hemipelagic, and debris flow alternate.

Summarizing, the sedimentary units corresponding to the acoustic BOSS horizon are indicative of an intermediate-upper slope environment, the chert-basalt breccia passing to a shallow water depositional environment for the carbonate-cemented breccia. The overlying apron sequence marks another variation in the sedimentary regime; in fact, the brecciated carbonate sandstone of the subunit 2A passes upward to the olive green clay(stone) and silt(stone) of unit 1A. The bottom part of the slope apron sequence, unit 1A, is characterized by coarser material than the top, with layers interpreted as debris flow deposits. These have different characters than the basal chert-basalt breccia because they lack an assemblage of hard-rock clasts and only include pebble-sized clasts of claystone having the same composition as the matrix material. This part of the apron sequence is thus envisaged as an accumulation of remobilized continental slope sediment.

The up-section passage from the carbonate-cemented breccia to the slope apron sediment is a major change both in the lithological types supplied by the source areas and in the sedimentation environment. Data on the benthic foraminifera reveal that the apron sequence deepens upward.

5. Geological Setting of the Nicoya Peninsula

The carbonate-cemented and chert-basalt breccias recovered at Site 1042 place the source areas on the upper or Caribbean plate and suggest that equivalent units might be exposed on the nearby Nicoya Peninsula. Fundamentally, Costa Rica is an igneous massif of arc volcanic rocks of Cenozoic age built over a basement of Mesozoic oceanic crust and superimposed sedimentary sequences of Cretaceous and early Tertiary age. Basin uplift, i.e., accretion or subduction, occurred at an intraoceanic subduction zone sometime during or after the late Cretaceous [Lundberg, 1982; Meschede and Frisch, 1998; Sinton et al., 1997].

The accreted mafic crust, the Nicoya Complex of Dengo [1962] (Figure 12), is well exposed on the Santa Elena, Nicoya, and Osa Peninsulas just seaward along the Pacific coast of Costa Rica (Figures 1 and 12).

The Nicoya Complex is subdivided into two lithostratigraphic sub-complexes: the Lower and the Upper Nicoya Complex [Wildberg et al., 1981; Gursky et al., 1984] (Figure 12). The Upper Nicoya Complex consists of pillow and flow basalt that includes small masses of pelagic sediment of Jurassic-Cretaceous age [Burr and Escalante, 1969]. The igneous units of the Upper Nicoya Complex are mainly those of a primitive island arc and oceanic plateau flood basalt [Wildberg, 1984; Sinton et al., 1997]. In northwestern Costa Rica (S. Elena Peninsula area, Figures 1 and 12), the Lower Nicoya Complex structurally overlies the Upper Nicoya Complex as a nappe [Bourgois et al., 1984]. The Lower Nicoya Complex consists of pillow and flow basalt representing oceanic basement [Wildberg et al., 1981; Gursky et al., 1984; Frisch et al., 1992]. It is associated with plagiogranite, diorite, and pelagic sediment, mainly radiolarian chert of early Jurassic to Santonian age [Schmidt-Effing, 1979; Tournon, 1984; DeWever et al., 1985; Baumgartner, 1987].

In the field, the imbricate architecture of the Nicoya Complex and structural as well as petrographic features define a south vergent stack, interpreted as an intracratonic accretionary prism of mostly mafic igneous rocks [Baumgartner et al., 1984; Frisch et al., 1992]. An overlap sequence of younger sedimentary deposit of late Cretaceous and early Tertiary age unconformably overlies the imbricate structure of the Nicoya Complex [Baumgartner et al., 1984]. The overlap sequence starts with the Garza Supergroup [Sprechmann, 1984] (Figure 12). The lowest formation of the Garza Supergroup is a discontinuous basal breccia, mainly formed by basaltic clasts. The Garza Supergroup can be subdivided into two groups [Baumgartner et al., 1984]. The lowest unit, the pelagic sedimentary deposits of the Sabana Grande Group, rests unconformably on the breccia or directly on the basement (Figure 12). The Sabana Grande Group is composed of limestone, radiolarian chert, and fine to coarse distal turbiditic beds ranging in age from Santonian to the Maastrichtian-early Paleocene [Baumgartner et al., 1984]. The overlying Samara Group [Sprechmann, 1984] represents the highest part of the Garza Supergroup (Figure 12). Clastic sediment typical of gravity flows characterizes the Samara Group. Conglomerate and proximal turbiditic beds both siliceous and calcareous, and ranging in age from upper Santonian to upper Eocene, formed the Samara Group [Baumgartner et al., 1984].

Starting from the Middle Eocene, angular unconformities and changes in the depositional environment marked the end of the Garza Supergroup sedimentation [Lundberg, 1982] (Figure 12). Carbonate and siliciclastic sediment, lying unconformably on the Garza Supergroup or directly on the basement, form the heterogeneous Mal Pais Supergroup, which is constituted of neritic to nonmarine depositional sequences (Figure 12). Baumgartner et al. [1984] recognized at least three shallow water events recorded by the Mal Pais deposits in the Nicoya Peninsula. The first event in southern Nicoya occurred during the middle-upper Eocene with burial of the Sabana Grande Group in the Cabo Blanco-Montezuma area. The second event is recorded in the central part of the Nicoya Peninsula from Nosara, to Cabo Blanco, to Paquera, where shallow water sediment of upper Oligocene-lower Miocene age accumulated on the Sabana Grande Group. The third and youngest event records deposition of shallow marine beds in the Cabo Blanco-Paquera area during the middle Miocene-Pleistocene.

The Cenozoic shallow water deposits of the Mal Pais Supergroup document uplift of much of the area of the shelf, which is underlain at depth by folded and faulted Mesozoic units of the mafic Nicoya Complex and Cretaceous and early Tertiary pelagic and turbiditic sediment of the overlying Garza Supergroup. Toward the end of Eocene time these deep water deposits accumulated in areas immediately adjacent to where deposition of the shallow water Mal Pais sequences was getting under way. The tectonic scenario envisaged by the sedimentary record of the Nicoya coast is thus one
Figure 12. Summary geological map of the Nicoya Peninsula and lithostratigraphic columns at the passage between Nicoya Complex-deep water sediment and shallow water sediment. Columns refer to the Punta Pelada, Tango Mar, and ODP Site 1042 sections.

involving the vertical movement of adjacent crustal blocks [Baumgartner et al., 1984] (Figure 12). Differential vertical tectonism continues today as the late Holocene deposits of the Cabuya-Montezuma coast of southern Nicoya (Figure 12) suggest net uplift rates in this area of 1.7-4.5 m Kyr\(^1\) (i.e., 1.7-4.5 km Myr\(^{-1}\)) since the middle to late Pleistocene [Marshall, 1991; Marshall and Anderson, 1995]. These high rates of uplift and the laterally discontinuous nature of exposed shallow water deposits of middle and late Cenozoic age open questions about the underlying mechanisms
involved. Because of the high rates and spatial limitations observed, for the southeastern Nicoya peninsula attention has been focused on vertical Nicoya peninsula linked to the underthrusting of seamounts and seamount groups [Marshall, 1991; Gardner et al., 1992; Marshall and Anderson, 1995]. The oversteepened trench slope offshore of Cabo Blanco [Von Huene et al., 1995; Fisher et al., 1998; Von Huene and Ranero, 1998] (Figure 1) may have developed because of localized rapid uplift and linked mass slumping related to subduction of the Fisher Seamount immediately to the southeast of the peninsula (Figures 1 and 12) [Marshall, 1991; Marshall and Anderson, 1995].

6. Outcrops of the BOSS Horizon along the Nicoya Coast

At Site 1042 the top of the acoustically defined BOSS horizon was evidently reached. However, a prominent unconformity separating rocks of contrasting age and lithology was not penetrated. What this unconformity might separate is implied by the contrasting lithologic composition of the breccia of shallow water limestone and the underlying chert-basalt breccia of Nicoya Complex debris. Because the BOSS horizon can be traced landward, without interruption, to near the Nicoya coast, we searched the Pacific margin of the peninsula for outcrops of neritic beds of the Mal Pais Supergroup unconformably overlying Nicoya basement complex or temporally related beds of the Garza Supergroup.

Neritic sedimentary deposits of the Mal Pais Supergroup crop out in restricted areas along the southwest coast of the Nicoya Peninsula (Figure 12), most notably at Punta Pelada (Figure 13a) northwest of Samara and in the Mal Pais-Cabo Blanco-Montezuma area along the southwestern and southern edges of the peninsula (Figure 13b). At Punta Pelada the Mal Pais Supergroup, represented by the Punta Pelada Formation, unconformably overlies the Garza Supergroup. The Garza Supergroup is here represented by the Barco Quebrado Formation, a distal facies of massive turbidite beds of Maastrichtian-early Paleocene age [Baumgartner et al., 1984]. The Barco Quebrado Formation thinly overlies the Nicoya basement complex, which is exposed just inland (1-2 km) of Punta Pelada. At Punta Pelada the Mal Pais Supergroup, and Barco Quebrado Formations. A 1-2-m thick basal breccia of the Punta Pelada Formation directly overlies the Garza Supergroup sediment (Figs. 12 and 13a). The breccia is composed of subrounded to angular sandstone clasts set in a sandstone matrix. The remainder of the overlying Montezuma Formation is fine grained and highly bioturbated. The Montezuma Formation is demonstrably a transgressive shallow marine deposit, with the lower part representing a high-energy, wave base environment [Lundberg, 1982, 1991]. Conglomerates with the same characteristics as the Montezuma basal part occur in several localities along the Montezuma-Cabuya coast to the SW (Figure 12; Lundberg, 1982).

At Tongo Mar, the Montezuma-Nicoya contact is exposed near sea level at the base of the modern sea cliff (Figure 13b). Just inland, at an elevation of 160-180 m, at the base of a Pleistocene terrace of marine origin, the Cobano surface, Nicoya Complex crops out under the Montezuma Formation (J. Marshall, personal communication, 1998). The trend of the Montezuma-Nicoya contact implies that the Montezuma Formation was deposited over an emergent early Miocene erosional surface that had appreciable relief (~ 10-100 m). The Montezuma Formation was deposited on this surface during its submergence in late Miocene to Quaternary time, interrupting the uplifting trend [Marshall, 1991; Marshall and Anderson, 1995]. A return to uplift beginning in the middle to late Pleistocene, possibly, as earlier noted, caused by underthrusting seamounts, led to emergence of the Montezuma Formation and erosion of the Cobano surface during the late Pleistocene [Marshall, 1991; Marshall and Anderson, 1995].

The deposition of the neritic and shelfal Punta Pelada and Montezuma Formations over much older and deep water rock sequences implies at least two cycles of uplift-subidence-rupture of coastal Nicoya since the early Tertiary. The recovery of evidence for depositionally, temporally, and lithologically similar relations at the base of Site 1042 implies thereby that a major unconformity of the first cycle, i.e., the geologic BOSS horizon, was closely approached at the bottom of this hole near the modern trench floor. Projected shoreward, the BOSS horizon can thus be tied to the Punta Pelada unconformity buried by late Oligocene shelfal beds of the Mal Pais Supergroup or to the Montezuma unconformity buried by late Cenozoic shallow beds of younger Mal Pais deposits. The circumstance of similar geologic rock sequences explored at the base of Site 1042 and along the nearby coastal area of the Nicoya Peninsula is thus in keeping with the predictions of the nonaccretionary model of Von Huene and Flueh [1994]
Figure 13. (a) Photograph and interpretative diagram of field relations where the Punta Pelada Formation (Mal Pais Supergroup) unconformably overlies the Barco Quebrado Formation (Garza Supergroup) at Punta Pelada. (b) The Montezuma Formation (Mal Pais Supergroup) unconformably overlies basalts of the Nicoya Complex at Tango Mar.
and as elaborated on more recently by Vannucchi et al. [1998], Meschede et al. [1999], and Raniero and Von Huene [2000].

7. Subsidence of the Costa Rica Margin: A Case For Subduction Erosion

Subduction erosion has been described as the major tectonic process involved in shaping the rock fabric of many convergent margins, as, for instance, the Japan, Tonga, Chile, and Peru margins [Von Huene et al., 1982; Von Huene and Scholl, 1991], and on the basis of regional coastal and offshore information, the margin bordering the Middle America trench [Scholl et al., 2000]. The process of subduction erosion is defined as the removal and transport of upper plate material toward subcrustal and mantle depths. The primary evidence for this process is the documentation of substantial (i.e., by 3-5 km) subsidence of the outer forearc during relatively short periods (20-50 Myr) of geologic time. Coastal uplift can be paired with offshore subsidence as documented by the coastal observations of Marshall [1991], Gardner et al. [1992], and Marshall and Anderson [1995]. Subduction erosion is also documented by vertical subsidence of the submerger margin in conjunction with progressive seaward tilting toward the trench [Clift and MacLeod, 1999]. Other fundamental evidence testifying to the efficacy of subduction erosion is the landward migration of the coastline and the arc magmatic front over substantial periods of geologic time [Rutland, 1971; Von Huene and Scholl, 1991].

For Costa Rica, Alvaredo [1993] has documented a landward shift of the Costa Rica volcanic arc of ~40-50 km since the middle Miocene. With respect to these primary observations concerning the consequences of subduction erosion at convergent margins, Leg 170 drilling results combined with coastal studies of the adjacent Nicoya Peninsula can be cited as matching evidence for substantial amounts (3-4 km) of ocean margin subsidence during Neogene and Quaternary time. The carbonate-cemented breccia drilled at Site 1042 exhibits shallow water sedimentological features typical of those described for the Punta Pelada and the Montezuma Formations. The chert-basalt breccia recovered underneath shows affinities with the Nicoya Complex rocks. Studies on the benthic foraminifera in the slope apron sediments show a deepening upward succession. The situation described at Site 1042 is thus similar to that for the Mal Pais Superterg, where this shallow water sediment rests on oceanic crust or depositionally kindred deep water sedimentary units along the Nicoya coast. The evidence of ocean floor alteration in the chert-basalt breccia of Site 1042 argues for a source terrane of altered and brecciated oceanic crust.

The present position of the carbonate-cemented breccia is ~3900 m below sea level, which is thus roughly equal to the amount of subsidence that affected the middle and lower slope since 16-17 Ma. Sedimentological and geochemical analyses on the carbonate-cemented breccia imply that the BOSS horizon is geologically a deep submerged regional surface of wave base and subaerial erosion. The erosional surface is time transgressive or diachronic as argued on the basis of the burial age of its exposed manifestation as the Mal Pais unconformity. Thus uplift and subsidence of the Mal Pais unconformity are discontinuous in time and space along the Nicoya Peninsula. Other margins lacking evidence for frontal accretion, and where drilling reached basement rock beneath a BOSS-type horizon, have been affected by subsidence instead of uplift, as, for example, the Japan [Von Huene et al., 1982], Mariana [Hussong and Uyeda, 1981], Peru [Von Huene and Scholl, 1991], and Tonga forearc [Clift and MacLeod, 1999].

Drilling of the Guatemala margin on Legs 67 and 84 ~500 km northwest of Leg 170 sites also recovered no evidence of net accretionary thickening or uplift since at least the beginning of the Eocene, if not since the pre-Campanian [Von Huene et al., 1985]. Accretionary thickening is also missing from the frontal part of the Costa Rica margin. The amount of underplating occurring beneath the BOSS surface remains, however, a matter of debate [Meschede et al., 1999]. However, the recovery of shallow water sediment at Site 1042 documents margin-wide subsidence during the past 16-17 Myr, the opposite observation of that predicted by models of massive underplating of oceanic sediment beneath the BOSS horizon, a process that would have thickened and thus uplifted the margin across most of its width. However, consistent with an accretionary underplating model is the lack of °Be in young Costa Rica eruptive rocks. The occurrence of °Be in arc magmas [Brown et al., 1982; Tera et al., 1986] provides quantitative information on the amount of sediment that is both accreted to and subducted beneath the margin to mantle depths (see review by J.D. Morris in Scholl et al., 1994). °Be is a cosmogenic, relatively short-lived isotope (half-life is 1.5 Myr) that is strongly adsorbed by fine-grained sediment entering the subduction zone. For this reason, °Be enrichments are commonly observed in modern arc volcanic rocks, but not in the eruptives of other tectonic settings [Tera et al., 1986; Morris and Tera, 1989; Morris et al., 1990].

North of Costa Rica, efficient subduction of °Be-enriched sediment and thus the lack of any significant measure of frontal and underplating accretion, is documented by the high content of °Be in arc volcanic rocks of Guatemala and Nicaragua [Carr and Rose, 1987; Carr et al., 1990]. But the Guatemala and Nicaragua °Be peak dramatically decreases to the southeast in Costa Rica volcanic units [Carr and Rose, 1987; Carr et al., 1990], despite the fact that the °Be content of Cocos plate sediment entering the Costa Rica subduction zone is sufficiently high to produce a strong arc signal [Valentine et al., 1997]. Thus the absence of °Be in young Costa Rican eruptive rocks seemingly greatly limits the volume of young oceanic sediment that is subducted to mantle depths. Because Leg 170 drilling established that frontal accretion is, effectively, not taking place, a rational explanation for the lack of a prominent °Be signal in Costa Rica eruptive rocks is that young underthrusting sediment is sequestered, at depth, by underplating beneath the base of the margin. This conjecture does not square with evidence for as much as ~4 km of margin subsidence near the trench floor in the face of, at the most, ~100 m of coastal uplift. The answer to the enigma of the absent °Be signal is unknown to us, other than to suggest that dilution caused by the flushing of subduction erosion debris to the mantle might be involved. Alternatively, °Be-bearing fluids and sediment might be transferred to the mantle at depths too shallow and cold for the generation of arc magmas [Leeman and Carr, 1995].

ODP sites 1040 and 1043 show that at the Costa Rica margin, effectively all the incoming oceanic deposits are subducted below its high-velocity rock framework [Kimura et al., 1997]. Bathymetric and seismic reflection data also
Figure 14. Tectonic evolution of the Pacific margin of Costa Rica off Nicoya Peninsula since early Miocene.
document that subducted seamounts thin the upper plate of the Costa Rica margin as they cut tectonic grooves through the base of the landward trench slope [von Huene et al., 1995, 2000]. Although tracts of seamounts are not now entering the sector of the Nicoya margin drilled by Leg 170, they may have done so in the past to, in part, account for the high rate of crustal thinning recorded by the shallow water limestone breccia recovered at Site 1042. More importantly, Ranero and von Huene [2000] infer from seismic reflection images that along the Costa Rica margin upper plate material is transferred to the lower plate by the removal of large rock lenses (as much as 1-2 km thick and tens of kilometers in length) from the base of the margin. Seamount subduction does not need to be involved in this process of basal subduction erosion.

8. Summary and Conclusions

The Cenozoic evolution of the Costa Rica margin may be interpreted from the combined implication of Leg 170 results, regional seismic reflection and refraction data, and the regional geology of the Nicoya Peninsula (Figure 14). These data form a coherent picture from the early Miocene to the present, but the Mesozoic history is based only on the rock record exposed in the Nicoya Peninsula where the stacked and shortened mafic rocks of the Nicoya Complex document the convergence of the Farallon and Caribbean plates.

By the early Tertiary, convergent processes uplifted and carried the Nicoya Complex and overlying deep water beds of the Garza Supergroup to shallow water depths and subaerial elevations (Figure 14). Subsidence of this regional erosion surface below wave base began at least by late Paleogene time, resulting in its burial by the calcareous and rich fossiliferous beds of the largely Neogene and younger Mal Pais Supergroup. Subsidence of the Mal Pais unconformity in Neogene and Quaternary time to trench depths requires the landward migration of the trench axis and coastline, relative to a fixed interior Costa Rica, by at least 40-50 km (Figure 14). The shift is about the same as the landward migration of the volcanic chain [Alvarado, 1993]. The Mal Pais erosion surface can thus be laterally tied to the BOSS horizon at Site 1042, which, near the base of the landward trench slope, separates Garza and Mal Pais depositional units.

As recorded at Site 1042, and at least since the top of the early Miocene (~16-17 Ma), the seaward edge part of the submergent forearc has subsided to a depth of 3900 m below the sea level—(Figure 14). Subsidence of the Mal Pais unconformity is inferred to record crustal thinning caused by processes of frontal and basal subduction erosion. The removal of upper plate material may be linked to the missing 10Be signal in the Costa Rica arc lavas and with the arcs’ landward shift. The recovery of shallow water carbonate beds at ~4000 m below seal level at the base of the forearc implies ~50-60 km of forearc truncation during the past 16-17 Myr. The trench axis landward migration is thus ~3 km Myr\(^{-1}\). The volume loss during subduction erosion is ~560-600 km\(^3\) per kilometer along the trench, equivalent to a rate of 34-36 km\(^3\) Myr\(^{-1}\).

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