

Active thrusting in the inner forearc of an erosive convergent margin, Pacific coast, Costa Rica

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[1] Structural and geomorphic analyses of the Fila Costeña thrust belt in southwest Costa Rica indicate active thrusting within the inner forearc. The Fila Costeña exposes three major thrust faults that imbricate the late Tertiary forearc basin sequence of the Térraba basin. The frontal thrust of the Fila Costeña marks the boundary between an uplifting inner forearc and a subsiding outer forearc, with only local uplift astride the indenting Cocos Ridge. On the basis of surface constraints a cross section across the thrust belt suggests that all three thrusts flatten into parallelism with a low-angle décollement horizon near the contact between the basement and the cover sequence of the Térraba basin. This décollement lies at a depth of ~ 4 km. The minimum shortening recorded by restoration of fault-related folds is 17 km, or 45%. Observations of late Tertiary marine sediments, tilted and faulted late Quaternary fluvial terraces, and raised Holocene marine terraces indicate that Fila Costeña uplift was likely initiated in the Quaternary and is ongoing. Given that the coastal mountains are separated from the Talamanca Range by a valley, the décollement that delaminates the forearc basin from the underthrusting forearc must continue as a flat beneath the valley but must link with the plate boundary along a crustal-scale ramp system, a structural geometry that has resulted in uplift of the Talamanca Range, the highest peaks in Central America. The dichotomy between uplift in the inner forearc and subsidence in the outer forearc is explained in terms of the response of an arcward thickening wedge to rough, subducting crust.

INDEX TERMS: 8150 Tectonophysics: Plate boundary—general (3040); 8123 Tectonophysics: Dynamics, seismotectonics; 8010 Structural Geology: Fractures and faults; 8015 Structural Geology: Local crustal structure; 8005 Structural Geology: Folds and folding;

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1. Introduction

[2] In the last two decades it has become increasingly apparent that a significant percentage of Earth's convergent plate boundaries are not actively accreting material but are experiencing erosion of the forearc [*von Huene and Scholl*, 1991]. In these cases the term “erosive margin” is appropriate for the outer part of the forearc where there is evidence for widespread lower slope scarring by seamounts, upper slope subsidence, and arcward trench migration. However, along plate boundaries such as the Japan Trench or the Middle America Trench in Costa Rica (i.e., cases where the outer forearc shows evidence for erosion [*Lallemand et al.*, 1992; *Ranero and von Huene*, 2000; *Vannucchi et al.*, 2001]), the innermost forearc is exposed, with the greatest uplift and unroofing in regions landward of bathymetric highs on the subducting plate [*Gardner et al.*, 1992; *Fisher et al.*, 1998].

[3] The dichotomy between these fundamental differences in vertical tectonism within the inner and outer forearc requires either seaward tilting of forearc slope sediments at the regional scale or development of a thrust fault or system of faults near the coastlines of erosive margins. Along accreting convergent margins (e.g., Barbados [*Westbrook et al.*, 1988] and Nankai [*Park et al.*, 2000]), on-land fold-and-thrust belts (e.g., Cape Smith [*Lucas*, 1989], the Pyrenees [*Meigs and Burbank*, 1997], and Taiwan [*Kao and Chen*, 2000]), and in analog experiments [*Gutscher et al.*, 1998], such faults have been described as “out of sequence” on the basis of the occurrence of slip on faults well arcward of the active deformation front or the frontal thrust. In these cases the faulting can be modeled and interpreted in terms of critical taper wedge mechanics [e.g., *Davis et al.*, 1983]. Thrusts are generated within the interior of the thrust belt from either work hardening of the basal shear zone [*Lucas*, 1989] or understeepening of the wedge shape with respect to critical taper [*Meigs and Burbank*, 1997]. The basal shear zone may experience work hardening if sediments experience dewatering and porosity loss [*Fisher and Byrne*, 1987] or if the thermal relaxation due to thrust stacking in the interior of continental collisions leads to metamorphism and grain growth [*Lucas*, 1989]. The process of frontal accretion in the absence of under-

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plating leads to lateral growth without vertical thickening in the rear of the wedge, a process that could cause understeepening.

[4] The development of out-of-sequence thrusts along nonaccreting or erosional convergent margins requires a different explanation in the sense that faulting at the rear of the wedge is accompanied by wedge thinning and arcward trench migration at the front of the wedge. Active thrusting in the inner forearc along these margins could allow for loading and underthrusting of the outer forearc and could account for at least some of the arcward trench migration and subsidence of the upper slope. The Fila Costeña thrust belt in southwest Costa Rica has characteristics that are consistent with such a system. These coastal mountains lie along a system of laterally continuous thrust faults that cut and deform the Tertiary forearc basin sequence. The frontal thrust of the Fila Costeña marks the boundary between the uplifted inner forearc and portions of both the subsiding outer forearc and the uplifting outer forearc astride the axis of the subducting Cocos Ridge.

[5] In this paper, we present the results of structural analysis of a central transect across the Fila Costeña, including a balanced cross section that constrains the depth to the décollement and the minimum shortening. A Holocene marine terrace along the thrust front and late Quaternary fluvial terraces within the transverse gorge and along the thrust front constrain timing and rates of thrusting. Finally, the dichotomy between the behavior of the offshore and onshore portions of the forearc is interpreted in terms of the mechanical response of an arcward thickening forearc to seamounts and ridges on the subducting Cocos plate.

2. Geologic and Tectonic Setting

2.1. Regional Tectonics

[6] The Pacific coast of Costa Rica is part of the Central American forearc and magmatic arc created by northeastward subduction of the Cocos plate beneath the Caribbean plate and Panama block at the Middle America Trench (Figure 1a). From the Nicoya Peninsula on the northwest to the Burica Peninsula (Figure 1b) on the southeast, a distance of ~400 km, subduction at the Middle America Trench becomes progressively more complex in response to Cocos plate bathymetry and trench indentation by the Cocos Ridge

(Figure 1b), upper crustal thickening (Figure 1c), shallowing of the Wadati-Beniof zone (Figure 1d), and intersection with the Panama fracture zone [Corrigan *et al.*, 1990; Protti, 1991; Gardner *et al.*, 1992; Protti *et al.*, 1994, 1995].

[7] The Quesada Sharp Contortion is a tear or sharp bend in the seismic slab at depths greater than 70 km that marks a significant southeastward shallowing of the Wadati-Beniof zone under central Costa Rica [Protti, 1991; Protti *et al.*, 1994]. To the northwest of this contortion (Figure 1d, section A-A'), deeper portions of the seismically active slab dip at about 80° and reach depths ranging from 200 km near the Nicaraguan border to 135 km at the tear. To the southeast of the contortion the seismically defined slab dips at about 60° (Figure 1d, section B-B') and decreases to near 0° (Figure 1d, section C-C') landward of the incoming Cocos Ridge [Protti *et al.*, 1995]. Our study of the Fila Costeña lies within thrust sheets of the upper plate in the region with a shallowly dipping Wadati-Beniof zone.

[8] The updip projection of the Quesada Sharp Contortion coincides with a change in bathymetry, roughness, and age of the Cocos plate (Figure 1b). The bathymetric change, the northern rough-smooth boundary of Hey [1977], roughly coincides with the transition from smooth Cocos lithosphere created at the East Pacific Rise to rough Cocos lithosphere created at the Cocos-Nazca Spreading Center. Variations in the age of oceanic crust in this region can be explained in terms of the position of the fossil trace of the Cocos-Nazca-Pacific triple junction and a small region of crust associated with an early Miocene (19.5 Ma) ridge jump [Barkhausen *et al.*, 2001].

[9] Southeast from the southern tip of the Nicoya Peninsula, the rough bathymetric domain of the Cocos plate subducts northeastward beneath the Caribbean plate and the Panama block at the Middle America Trench (Figures 1b and 1c). Prominent bathymetric features on the rough domain include the aseismic Cocos Ridge, the Quepos Plateau, and the Fisher seamount group (Figure 1). Subduction of these bathymetric features has a profound effect on seismicity [Protti *et al.*, 1995], trench-slope morphology [von Huene and Scholl, 1991; von Huene and Flüß, 1994; Dominquez *et al.*, 1998; von Huene *et al.*, 2000], and the style of both short-wavelength [Marshall and Anderson, 1995; Fisher *et al.*, 1998; Gardner *et al.*, 2001] and long-wavelength [Gardner *et al.*, 1992] forearc deformation. Furthermore, the incoming seamount domain has caused

Figure 1. (a) Major plate tectonic features of Central America. Principal plates are bold and underlined. SOAM, South America; CARIB, Caribbean; SPDB, South Panama deformed belt; NPDB, North Panama deformed belt; CCRDB, central Costa Rica deformed belt; PPB, paleoplate boundary; RJ, ridge jump. Box shows location of Figure 1b. Major tectonic features are compiled from Mackay and Moore [1990], Silver *et al.* [1990], Meschede *et al.* [1998], Barckhausen *et al.* [2001], and Marshall *et al.* [2000]. Modern plate motion relative to the Caribbean is from DeMets *et al.* [1990] and Protti *et al.* [1995]. Small numbers give age of Cocos seafloor. (b) Generalized geologic map of Costa Rica and bathymetry of Cocos plate. Bathymetric data are from von Huene and Flüß [1994]. Contour interval is 200 m, bold, and labeled every 1 km. Magnetic anomalies are from Barckhausen *et al.* [2001]. FSG, Fisher seamount group; QP, Quepos Plateau; CR, Cocos Ridge; CCRDB, central Costa Rica deformed belt. Geologic regions are represented as follows: Nicoya Complex, dashed; active arc volcanics, dot-dashed; extinct arc volcanics, stippled; Cretaceous and Cenozoic fluvial to marine clastic and volcanoclastic rocks, horizontal lined. Box shows approximate location of study area in the Fila Costeña. (c) Crustal structure along line X-X' in Figure 1b (modified from von Huene *et al.* [2000]). (d) Location of historic earthquakes across the Wadati-Beniof zone along sections A-A', B-B', and C-C' in Figure 1b (modified from Protti *et al.* [2001]).

considerable tectonic erosion in the offshore part of the forearc. The trench slope is scalloped and indented around seamounts, and margin-perpendicular scarps on the outer slope record net subsidence in the wake of subducted seamounts (Figure 1b) [von Huene et al., 1995]. As these features indicate, the outer forearc kinematics are characterized by initial uplift in response to the underthrusting of a

seamount followed by subsidence in the wake of the seamount [e.g., von Huene and Lallemand, 1991; von Huene and Scholl, 1991; von Huene et al., 1995]. This deformation history leaves a characteristic signature within the lower slope of a scarp-bounded depression where a seamount has passed beneath the margin [von Huene and Scholl, 1991; Dominguez et al., 1998].

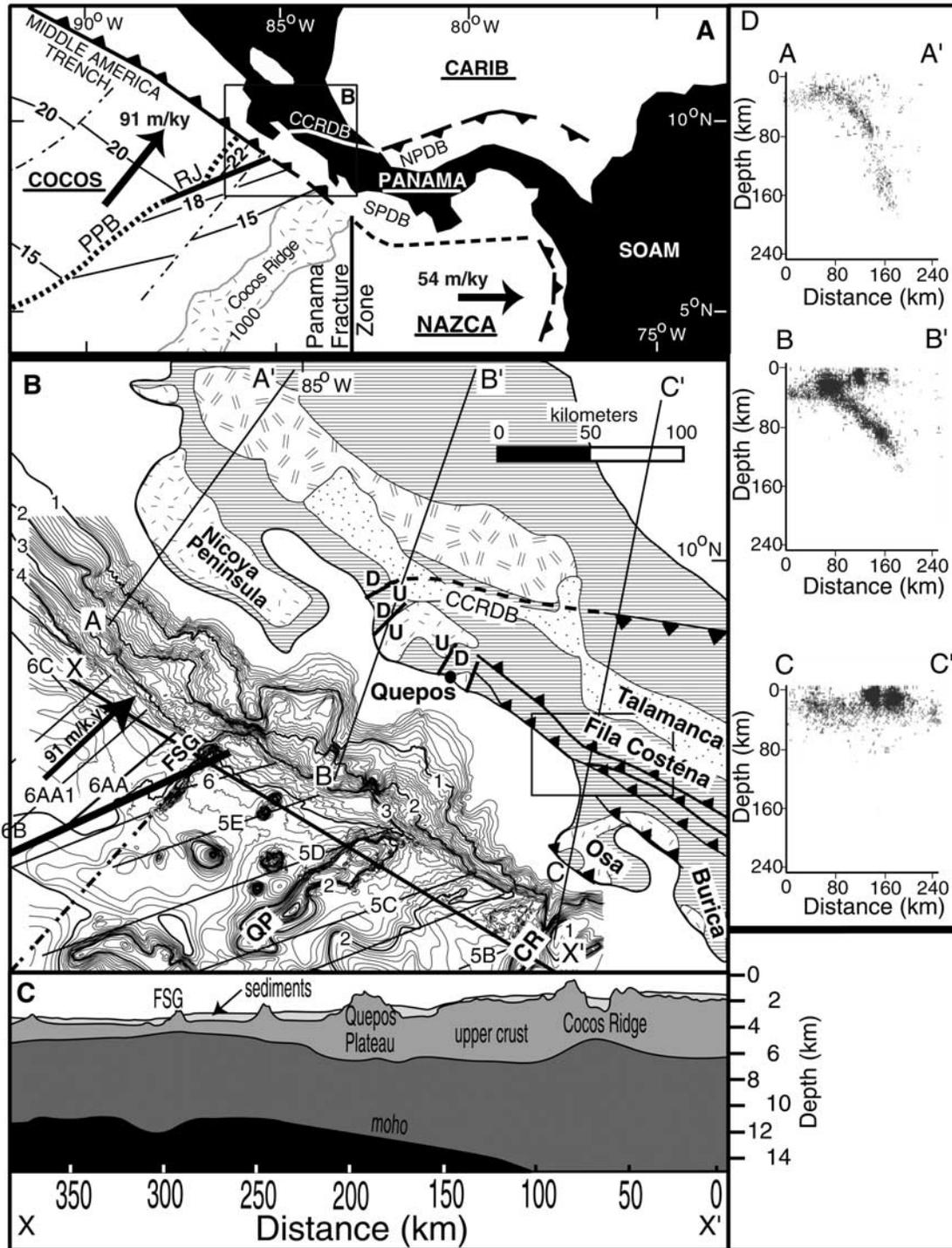


Figure 1.

[10] The upper slope in the offshore portion of the Costa Rican forearc displays a rough, irregular reflector in seismic reflection profiles that marks the boundary between the low-velocity slope cover and a higher-velocity margin wedge consisting of Caribbean basement or old accreted material [Shipley *et al.*, 1992; McIntosh *et al.*, 1993; von Huene and Flüh, 1994; Hinz *et al.*, 1996; Ranero and von Huene, 2000, von Huene *et al.*, 2000]. This reflector has been interpreted as an unconformity buried by a deepening upward sedimentary sequence that records late Tertiary subsidence of the outer forearc and arcward migration of the trench axis [Vannucchi *et al.*, 2001].

[11] The inner forearc along the Costa Rican margin records a history of net uplift [Gardner *et al.*, 1992; Fisher *et al.*, 1998; Gardner *et al.*, 2001] that contrasts sharply with the net subsidence in the outer forearc [Vannucchi *et al.*, 2001]. In the subducting seamount region the inner forearc displays a system of active faults oriented at high angles to the margin that segments the forearc thrust belt into blocks with significantly different uplift rates [Fisher *et al.*, 1998; Marshall *et al.*, 2000]. Trunk segments of fluvial systems that drain the Pacific slope typically follow these margin-perpendicular faults. Correlation of marine and fluvial terraces between drainages and across faults along the Costa Rican Pacific coast reveals a pattern of spatially variable forearc uplift that roughly coincides with the distribution of incoming seamounts. Uplift rates range from 1 to 8 m/kyr, with the greatest uplift landward of the Cocos Ridge and related seamount chains [Gardner *et al.*, 1992; Fisher *et al.*, 1998; Gardner *et al.*, 2001; Sak *et al.*, 2004a]. The magnitude and distribution of Quaternary uplift along the Pacific coast of Costa Rica suggest that despite a thin incoming sediment pile the inner forearc has thickened, a characteristic that can only be due to thrusting and/or underplating of seamounts.

[12] Our study describes a series of thrust sheets in the Fila Costeña that lie at the toe of the Talamanca Range, the uplifted and exhumed plutons of the extinct arc (Figures 1 and 2). The Fila Costeña is a 250-km-long and 20-km-wide mountain range with peak elevations of 1000–1500 m. Thrusts of the Fila Costeña are oriented parallel to the Middle American Trench, and the thrust belt is bounded by the Pacific coast and the Valle General-Rio Coto Brus Valley to the northeast (Figure 2). The Fila Costeña is bisected perpendicularly by the Rio Térraba (formed by the convergence of the Rio General and Rio Coto Brus). On the basis of previously mapped geomorphic, structural, and geologic features in the region, Protti *et al.* [1994] suggested that the genesis and evolution of the Fila Costeña and the General-Coto Brus Valley are strongly related to the same process: episodic crustal deformation induced by the repetitive coseismic slip along a buried fault system. The Longitudinal Fault defines the thrust front for the Fila Costeña and extends for over 220 km from beyond Punta Dominical to the Panama fracture zone (Figure 2). This fault has been interpreted as a reverse fault [Mora, 1979], a dextral fault [Bérrange, 1987; Bérrange and Thorpe, 1988], or a sinistral fault [Corrigan *et al.*, 1990; Mann and Corrigan, 1990; Bullard, 1995]. Here we present structural

data throughout the Fila Costeña that can be used to address the existence of strike-slip motion along and within the mountain front.

2.2. Local Stratigraphy

[13] Four formations exposed within the Fila Costeña thrust belt represent a slope basin sequence (i.e., the Térraba basin) that is 2–4 km thick: the Brito Formation, the Térraba Formation, the Curré Formation, and the Paso Real Formation (Figure 3) [Mora, 1979; Lowery, 1982; Phillips, 1983; de Boer *et al.*, 1995]. The Brito Formation is a middle to upper Eocene shallow water sequence composed of bioclastic limestones and mixed bioclastic and volcanoclastic turbidites. Carbonate units equivalent to the Brito occur discontinuously along the Pacific coast of Costa Rica and Nicaragua, suggesting that there was a regional shoaling that may correspond to the development of the Costa Rica-Panama arc complex [Lundberg, 1982; Phillips, 1983]. The contact between the Brito and the underlying basement is poorly exposed in the Fila Costeña but is most likely an unconformity or a faulted unconformity. In general, the Brito Formation is exposed in rugged, narrow ridges parallel to the thrust front.

[14] The Térraba Formation overlies the Brito Formation along a gradational contact and consists of ~1000 m of Oligocene to lower middle Miocene volcanoclastic turbidite fan deposits [Phillips, 1983]. In general, the Térraba Formation records a coarsening upward sequence, with mudstone and marl facies near the base grading into sandstone and conglomerate facies near the top. Gabbroic intrusions within the Brito and Térraba Formations have been dated by K-Ar at 15–11 Ma [de Boer *et al.*, 1995]. The Térraba Formation is conformably overlain by the middle to upper Miocene Curré Formation. The Curré Formation consists of ~800 m (in the Térraba Valley) of conglomerates, sandstones, and mudstones deposited in a shallow marine or fluvial/lacustrine environment [Lowery, 1982]. Thus the stratigraphy of the Térraba basin consists of a coarsening upward, transgressive-regressive cycle from the middle Eocene through the upper Miocene.

[15] The existence of an unnamed Pliocene marine mudstone overlying the Curré Formation [Kesel, 1983] is one indicator that thrusting within the Térraba basin did not begin until after the late Pliocene. The Pliocene mudstone is unconformably overlain by the Pliocene Paso Real Formation, a volcanic deposit of lahars, pyroclastics, and lava flows ranging in age from 1 to 4 Ma [de Boer *et al.*, 1995].

[16] Significant Quaternary deposits within the Fila thrust belt are limited to discontinuous fluvial terraces of the Rio Térraba and isolated marine terraces (Figure 2). A flight of four fluvial terraces up to 100 m above river level occurs along the transverse gorge and along the thrust front near the mouth of the Rio Térraba [Bullard, 1995]. Thrust faulting along the mountain front locally offsets the uppermost terrace [Bullard, 1995]. Ages of the fluvial terraces are poorly constrained to the Pleistocene (20 to >40 ka) and Holocene by correlation of soil profiles to radiocarbon-dated deposits on the Osa Peninsula [Bullard, 1995]. Here we use a well-preserved, Holocene marine terrace along the mountain

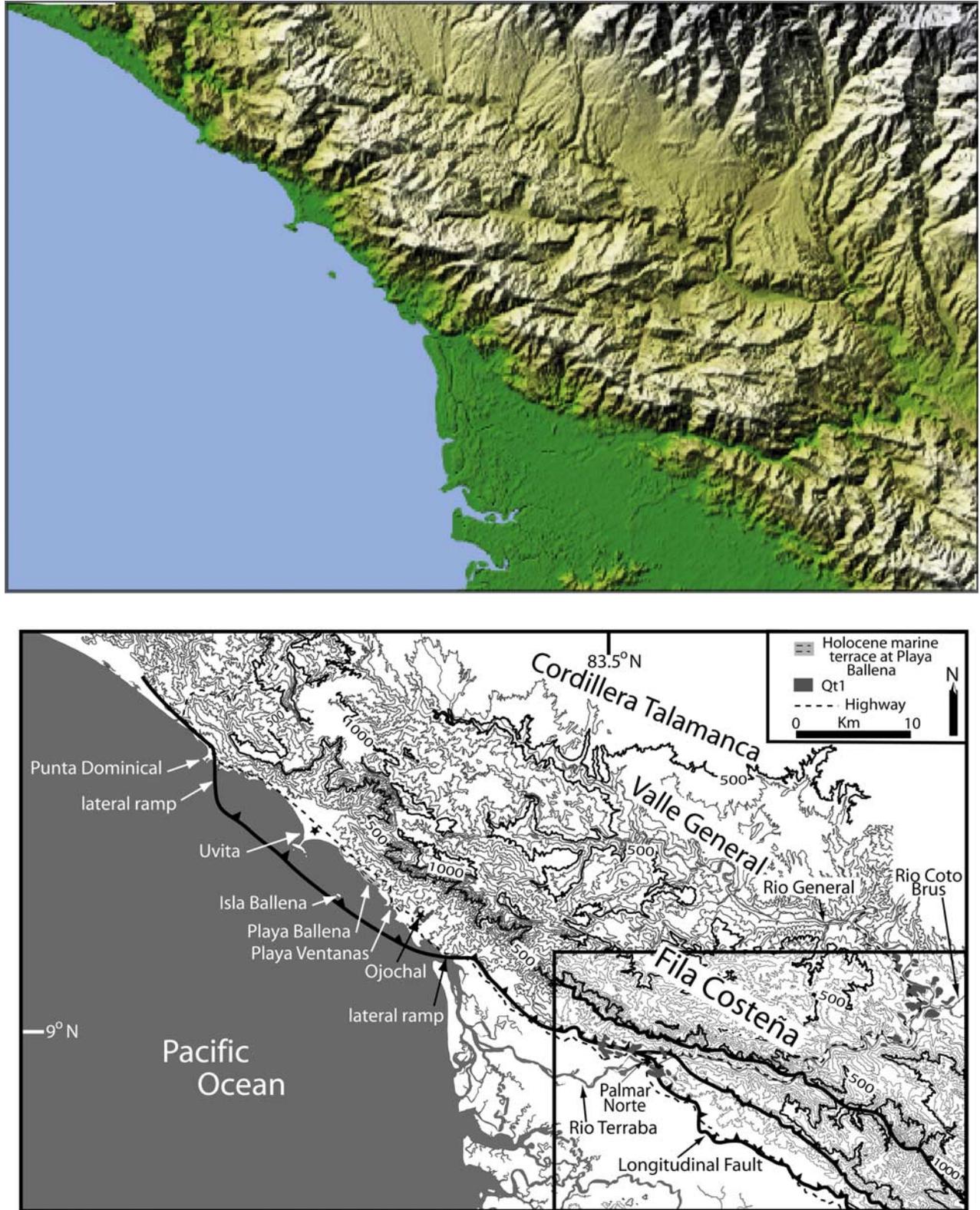


Figure 2. Shuttle Radar Tomography Mission series colored height and shaded relief image of the Fila Costeña (available at <http://photojournal.jpl.nasa.gov>). Physiographic features in the central Fila Costeña show the trace of thrust faults and distribution of high-level Quaternary fluvial terraces along the Rio Terraba and Holocene marine terrace along the Pacific coast at Playa Ballena. Contour interval is 100 m with 500- and 1000-m contours in bold.

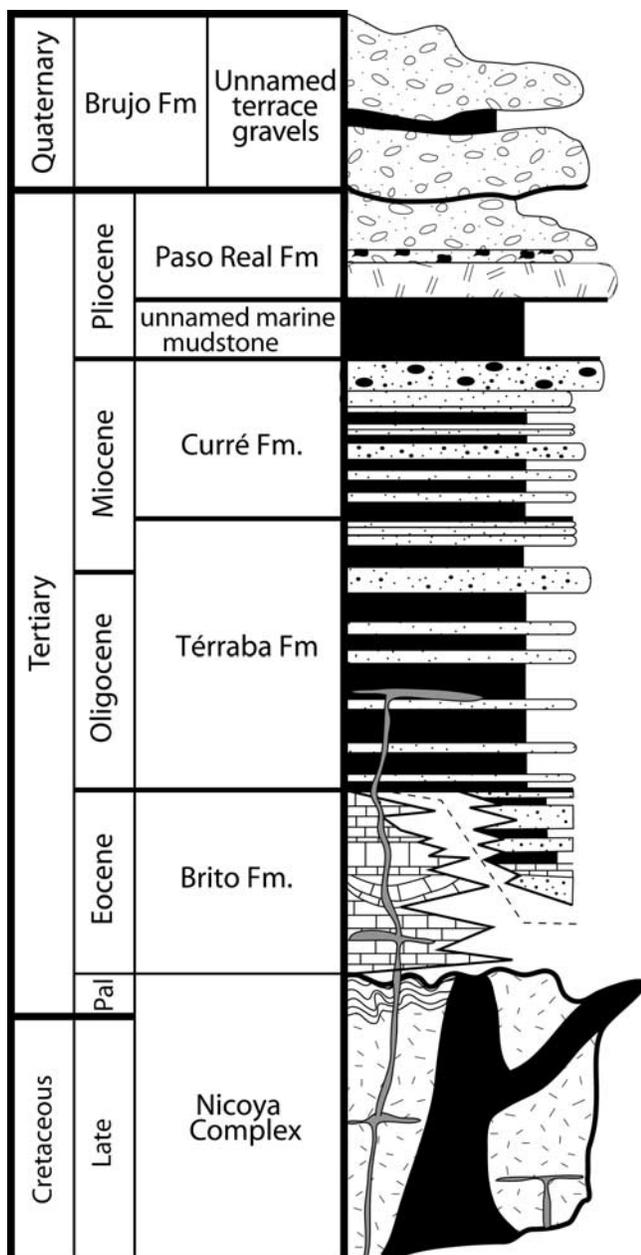


Figure 3. Stratigraphic column for the central Fila Costeña (modified from Phillips [1983]).

front (Figure 2) to estimate uplift rate of the frontal thrust of the Fila Costeña. Holocene uplift rates and regional terrace correlation that are based on weathering rind thickness of basaltic gravel clasts further constrain ages of the fluvial terraces of the Rio Térraba and the age of thrusting.

3. Structural Geology and Quaternary Deformation of the Fila Costeña

[17] The Fila Costeña is ~20 km wide in the area of the Térraba gorge and exposes three major thrust faults that

imbricate the late Tertiary forearc basin sequence of the Térraba basin (Figures 4 and 5, thrust sheets 1, 2, and 3). In general, strata strike parallel to thrusts (WNW-ESE) and dip ~15°–30° in the front 15 km of the thrust belt. Strata dip more gently or are subhorizontal in the rear 7 km of the thrust belt (e.g., bordering the Valle General). In the Térraba gorge each of the three thrust faults carries the limestones of the Brito Formation in the hanging wall. Thus the oldest units of the basin are exposed in imbricates near the thrust front, and younger units (e.g., the Curré Formation) are restricted to the gently dipping rocks along the trailing edge of thrust sheet 3 (Figure 5). Thrust sheets 1 and 3 display hanging wall cutoffs and fault-related folds. However, on thrust 2 the hanging wall cutoffs are eroded. Moreover, the basal fault of thrust sheet 2 is a hanging wall flat near the base of the Brito Formation.

[18] To the northwest of the Térraba gorge the thrust front steps seaward along a lateral ramp and then jogs back landward along a lateral ramp farther to the northwest near Punta Dominical. This trenchward step in the thrust front leads to exposure of dipping strata on Isla Ballena and preservation of the shallow marine terrace between Isla Ballena and Playa Ballena (Figure 2).

[19] A balanced cross section can be constructed from structural data on Tertiary and Quaternary deposits exposed in the Térraba gorge. The presence of Brito limestones at the base of each thrust sheet, combined with the observation of a hanging wall flat near the base of the Brito Formation in thrust sheet 2, suggests that all three thrusts flatten into parallelism with a low-angle décollement horizon in the vicinity of the contact between the basement and the cover sequence of the Térraba basin. This décollement lies at a depth of ~4 km. On the basis of the cross section the minimum slip is 5.2, 6.1, and 6 km for faults 1, 2, and 3, respectively, with a minimum shortening of ~17 km, or 45%.

[20] Estimates for fault slip are based on the assumption that strike-slip deformation within the belt is small. Limited strike-slip motion is generally consistent with the geometry of fault-related folds in the area. The anticline in the hanging wall of fault 1 has a fold axis with a trend of 325° and a 7° plunge, and the anticlinal fold axis related to fault 3 has a trend of 307° and a plunge of 8°. Fold axes are near horizontal parallel to the local strike of fault surfaces and near orthogonal to the cross-section line. Both anticlines have steep, NE dipping axial surfaces and interlimb angles of 90°–100°. The geometry of these map-scale folds is similar to the geometry of rare, mesoscale folds in the area. All folds verge to the SW throughout the Térraba gorge area. The only evidence for NE vergent thrusting is a mesoscale duplex developed within the coarse facies of the Térraba Formation that are exposed on both sides of Playa Ventanas (Figure 6). At Playa Ventanas, strata dip to the NE, but at least seven imbricates are observed with slickenlines that depict downdip thrusting to the northeast. The duplex records a minimum cumulative slip of 15 m. We interpret this structure as evidence for a passive roof near the thrust front [e.g., Banks and Warburton, 1986; Vann *et al.*, 1986] that was backthrust over the southwest advancing

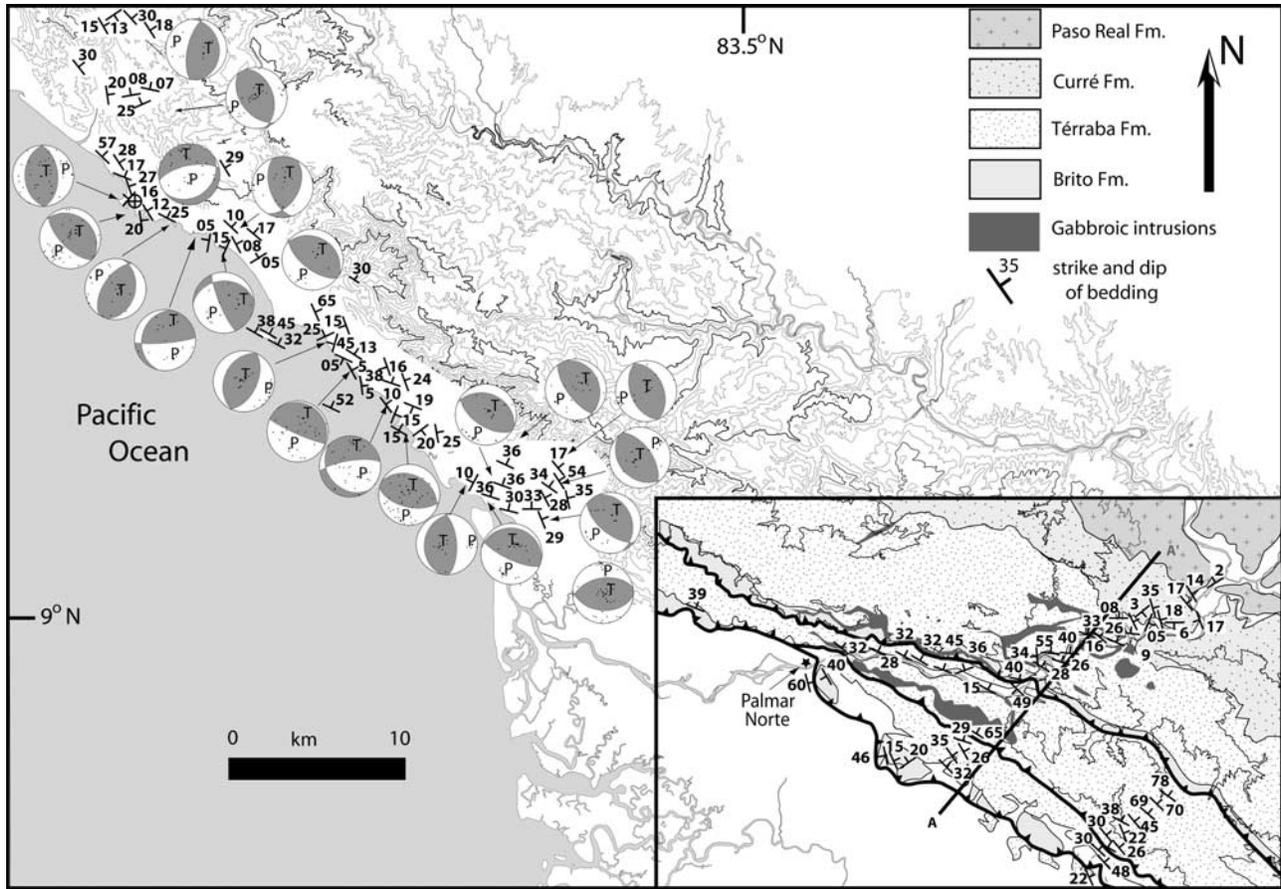


Figure 4. Map of the Fila Costeña from Punta Dominical to the Rio Térraba gorge with fault plane solutions that depict the shortening and extension axes on the basis of mesoscale fault populations [e.g., Marrett and Allmendinger, 1990]. See Figure 2 for contour labels.

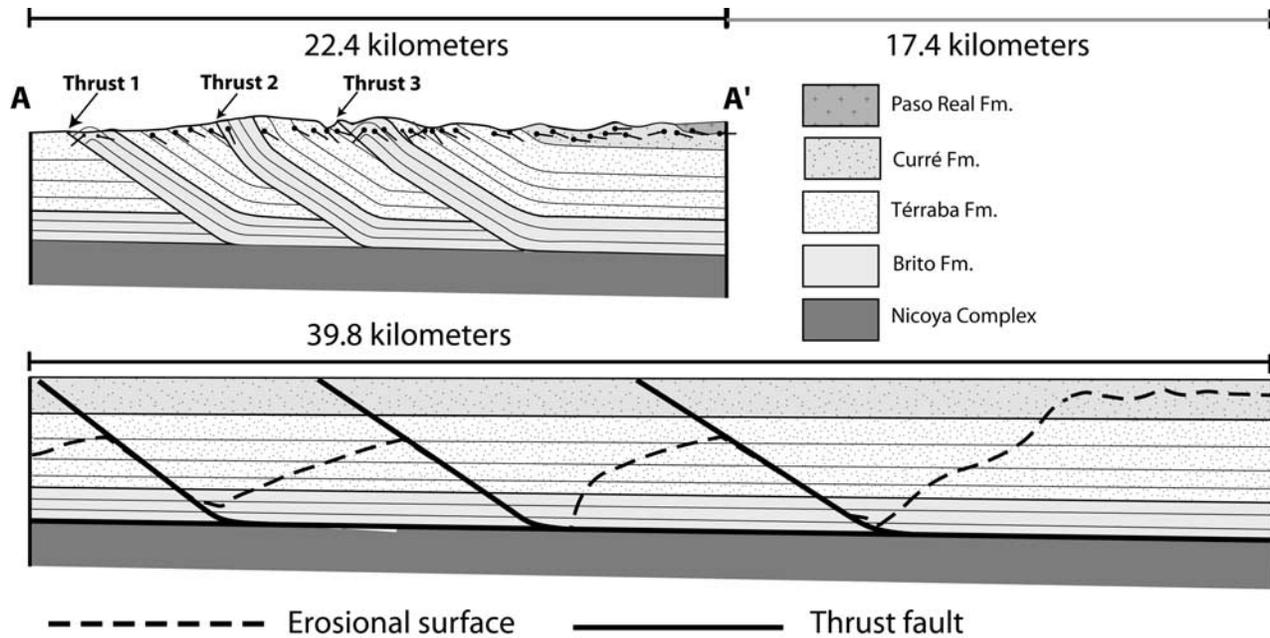


Figure 5. Cross section along line A-A' from Figure 4 insert.

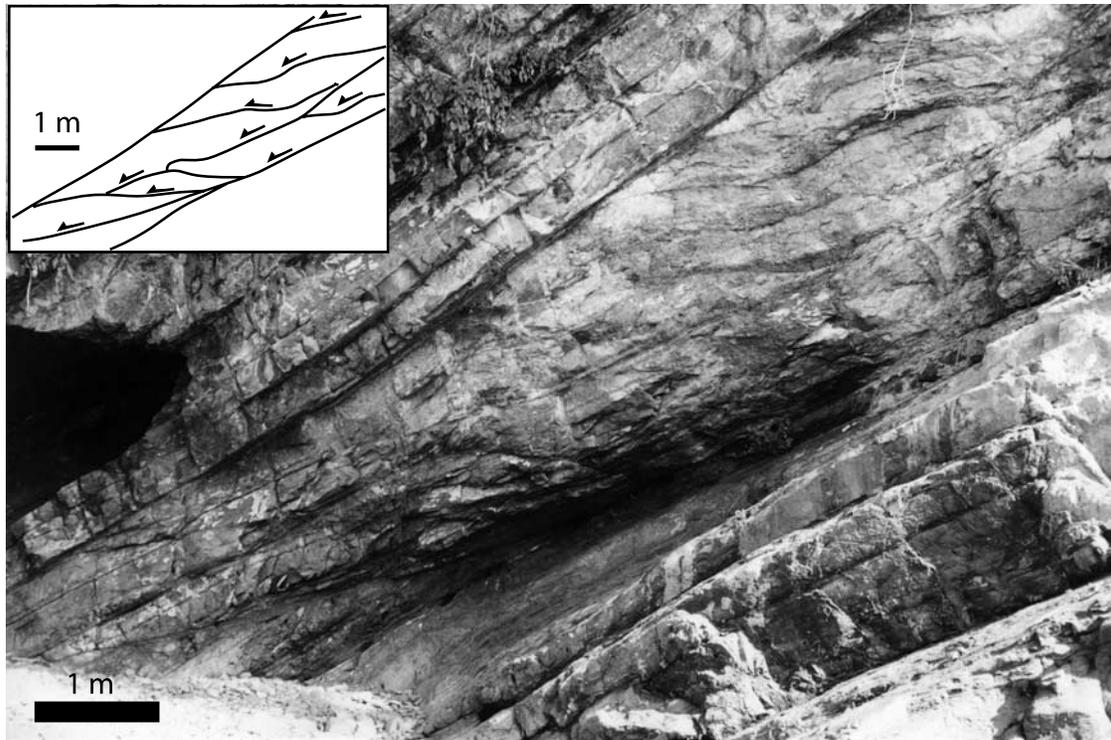


Figure 6. Field photograph of mesoscale imbricates along the coast that form a duplex recording backthrusting prior to the landward tilting of the package.

imbricates deeper in the section, with later northeast tilting of the package after development of a more seaward thrust that cut through the roof sequence.

[21] Another argument against strike-slip motion is that populations of mesoscale faults throughout the Fila Costeña generally depict top-to-the-southwest slip with almost no strike-slip component (Figure 4). Mesoscale faults are typically bedding-parallel or steeper northeast dipping than bedding. The strike-slip faults that we have observed are related to the two lateral ramps, particularly in the region around Punta Dominical.

[22] The presence of an uplifted Holocene marine terrace and faulted late Quaternary fluvial terraces indicates that deformation of the Fila Costeña is active and ongoing. A Holocene marine terrace is preserved along the coast at Playa Ballena (Figures 2 and 7c) [Sanchez, 2002]. Two raised beach ridges occur for several kilometers along the marine terrace (Figure 7c). The higher and older beach ridge has a radiocarbon age of 5.54 ± 0.07 ka from a carbonized log excavated from an auger hole.

[23] Uplift rate of the marine terrace can be calculated using the following equation (modified from Gardner *et al.* [1992]):

$$\text{uplift rate(m/kyr)} = \frac{\text{modern elevation (m)} - \text{depositional elevation(m)} - \text{paleosea level elevation (m)}}{\text{age(ka)}}$$

Elevation is measured relative to modern mean sea level and is positive upward. The modern elevation, -0.435 m, is that

of the carbonized log in the auger hole. Error range for the modern elevation is the sum of two factors: ± 0.015 m from the accuracy of the transit survey and ± 0.3 m from the average thickness of debris at the high tide debris line, the original transit survey datum in the field. The depositional elevation is reconstructed from modern environments. The log was most likely deposited in a backswamp environment near mean sea level. This yields a depositional elevation of 0 m with an associated error of ± 0.6 m because this environment is identifiable in the field to within approximately half a tidal cycle. Paleosea level elevation was calculated using the most recently compiled Holocene sea level curve of Fleming *et al.* [1998]. Paleosea level elevation at the time of deposition of the dated log was 2.3 m. The error term in the sea level calculation is ± 1 m. The calculated Holocene uplift rate for the marine terrace at Playa Ballena is 0.34 ± 0.4 m/kyr.

[24] At Playa Ballena the marine terrace is developed exclusively where strata exhibit dips of $\sim 5^\circ$ (Figure 4), suggesting that the terrace is preferentially developed above a flat in the underlying fault surface. However, along rocky headlands adjacent to Playa Ballena, inner edge notches that formed concurrently with the Holocene beach ridge are well preserved and indicate significantly higher uplift rates. Inner edges along rocky headlands form at mean high water, and the modern inner edge coincides with the modern high tide debris line. Given a tidal range of ~ 2.4 m at Playa Ballena, the modern inner edge occurs at ~ 1.2 m above mean sea level. However, the Holocene inner edges at Playa Piñuela, 3 km to the southeast, occur at elevations ranging from 2.5

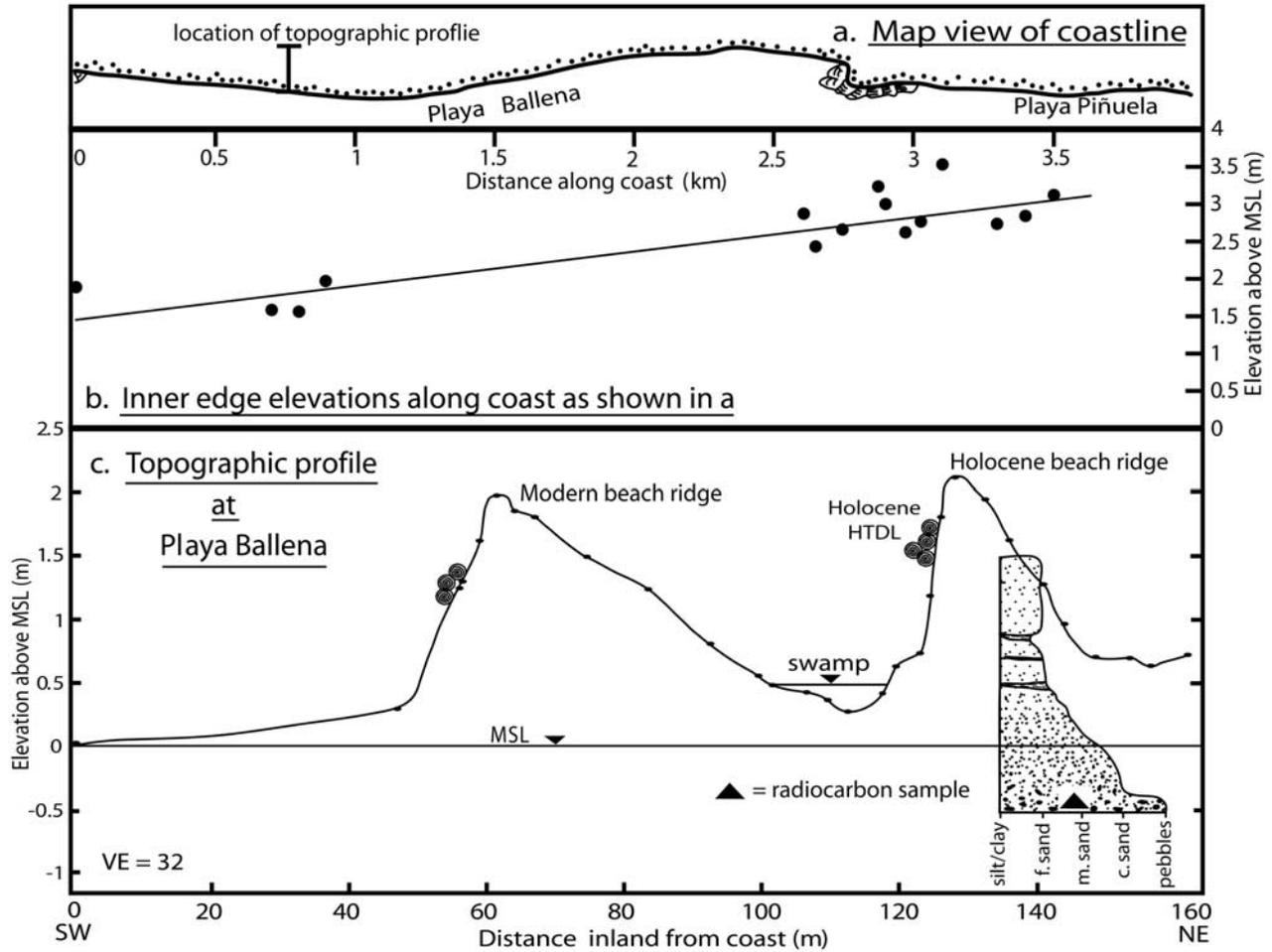


Figure 7. (a) Map of coastal section at Playa Ballena (see Figure 2 for location). (b) Plot of Holocene inner edge elevations with distance along coast. (c) Topographic profile across the marine terrace at Playa Ballena. Profile depicts the double beach ridge sequence and the stratigraphic column of an auger hole in the Holocene beach ridge. MSL, mean sea level; HTDL, high tide debris line, which marks the seasonal average high tide line.

to 3.5 m, indicating uplift rates at least twice as large as at Playa Ballena, ~ 0.7 m/kyr. Interestingly, dip of strata increases from $\sim 5^\circ$ at Playa Ballena to 10° – 15° along the rocky headlands near Playa Piñuela. Rates of uplift along the rocky headlands that lie above frontal ramps in the frontal thrust with dips $>30^\circ$ (Figure 4) are likely to be significantly more rapid than in the area of the raised marine terrace, with rates possibly greater than 1.5 m/kyr. This is one of the few places around the world where an active frontal thrust is in the zone of maximum marine wave abrasion with potentially high rates of mass removal and unloading further promoting movement along the frontal thrust.

[25] The highest fluvial terraces of the Rio T erraba and tributaries, Qt1 of Bullard [1995], are faulted along the mountain front (Figure 8), but previously, the age of that terrace has only been constrained to >40 ka. The Holocene uplift rate at Playa Ballena can be used to further constrain the age of faulted fluvial terraces along the mountain front.

A terrace at Ojochal (see Figure 2 for location; Qt1 of Bullard [1995]), 5 km to the southeast of Playa Ballena, occurs at 90 m. Extrapolation of the Holocene uplift rate from Playa Ballena suggests that the Qt1 terrace is most likely oxygen isotope stage (OIS) 5e (~ 120 ka) but is possibly as old as OIS 7 (~ 200 ka) (Figure 8).

[26] Weathering rind thicknesses on basalt clasts from Bullard’s [1995] Qt1 terrace range from 2.5 cm on well-preserved, undisturbed terraces to 0.8 cm on poorly preserved, colluviated, and bioturbated terraces [Murphy, 2002]. Correlation with dated terraces along the coast [Sak et al., 2004b] using rind thickness indicates an age for Qt1 of ~ 120 ka (OIS 5e). This age overlaps the Holocene extrapolation at OIS 5e. At the coast, where the Rio T erraba is cutting down through unconsolidated fluvial-deltaic deposits, river incision likely keeps pace with rock uplift. Incision rates for these fluvial terraces range from 0.5 to 1.5 m/kyr, somewhat less than the 3.0 m/kyr maximum incision rate inferred by Bullard [1995] but close to the rock

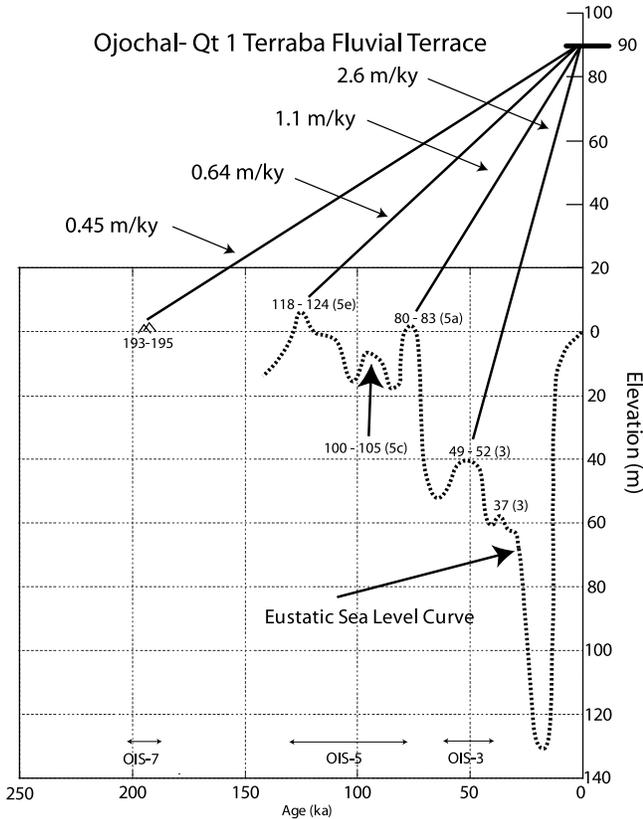


Figure 8. Range of possible average uplift rates for the Qt1 fluvial terrace that occur at an elevation of 90 m along the Pacific coast at Ojochal (see Figure 2 for location). Calculated uplift rates, assuming a long-term uniform uplift rate and terrace formation at sea level highstands, are shown only for oxygen isotope stages (OIS) 3, 5a, 5c, 5e, and 7. Uplift rates most closely match the Holocene uplift rate of 0.34 ± 0.4 m/kyr for a Qt1 terrace formed at OIS 5e or 7. Eustatic sea level curve is from Chappell and Shackleton [1986] and Lambeck and Chappell [2001].

uplift rate estimated for frontal thrust ramps with $\sim 20^\circ$ dips. These data indicate that the frontal thrust of the Fila Costeña is active during the late Quaternary, with rock uplift rates ranging from slightly less than 0.4 m/kyr to a maximum of 1.5 m/kyr. Given the correlation between uplift rate and the dip of strata, the uplift rate likely reflects the dip on underlying footwall ramps. The estimated slip rate on a frontal thrust with a dip of 20° and an uplift rate of 1.5 m/kyr is 4.4 m/kyr.

4. Discussion

[27] This study documents a minimum of 17 km of shortening related to three major thrust faults near the front of the mountainous Fila Costeña of Costa Rica. The cross section indicates that this shortening involves decoupling of the forearc basin sequence from a stronger margin wedge, with regional-scale underthrusting of the outer forearc

beneath an inner forearc thrust system. There are several observations indicating that this observed thrusting did not begin until the Quaternary and is likely ongoing. The unnamed Pliocene marine deposit that caps the Terraba basin sequence [Kesel, 1983] indicates that uplift above sea level did not occur until after the Pliocene. Late Quaternary terraces exposed in the gorge are tilted parallel to underlying strata of the Miocene Curré Formation, and the Rio Terraba is deeply incised into them. Along the mountain front the late Quaternary terraces are offset along thrust faults. Incision rates calculated from the Qt1 terrace range from 0.5 to 1.5 m/kyr. Finally, the Holocene marine terrace at Playa Ballena indicates an uplift rate of 0.34 ± 0.4 m/kyr along a gently dipping segment on the underlying fault surface, with rates perhaps as high as 1.5 m/kyr along frontal ramps. If a minimum of 17 km of underthrusting accumulated since the deposition of the Pliocene marine mudstone (since 2–5 Ma), the minimum rate of underthrusting would be 3.4–10 m/kyr. In other words, at least 4–10% of the total Cocos-Caribbean convergence (i.e., ~ 90 m/kyr) has shifted from the Middle America Trench to the interior of the forearc. In the absence of accretion this shift can account for a significant portion of the total arcward trench migration since the middle Tertiary as measured to the southwest [e.g., Vannucchi et al., 2001], although the total erosion and hence trench migration is likely greater landward of the Cocos Ridge. The loading associated with crustal thickening and southward directed thrusting can explain the landward thickening slope apron sequence observed offshore [Hinze et al., 1996]. An initiation of thrusting in the Quaternary is consistent with the timing for eastward migration of the Panama fracture zone across this part of the margin and the arrival of thickened crust related to the Cocos Ridge [Gardner et al., 1992; MacMillan et al., 2004]. If the axis of the Cocos Ridge arrived along the margin at ~ 1 Ma as suggested by plate tectonic models [Gardner et al., 1992] or 2–3 Ma as suggested from age and distribution of volcanic units [MacMillan et al., 2004], then the rate of shortening in the Fila Costeña thrust belt could be >17 m/kyr.

[28] In previous studies, the 11–15 Ma gabbros are mapped cutting major faults within the region of the Terraba gorge [Mora, 1979; Kolarsky et al., 1995]. However, these gabbros predate upper Miocene and Pliocene marine deposition in the Terraba basin, and we have not been able to document any cases of a gabbro cutting a thrust fault in the area. Consequently, we favor the interpretation that the gabbro sills were imbricated during the shortening recorded within the Fila Costeña.

[29] More than 17 km of slip has occurred at the contact between the margin wedge and the overlying forearc basin sequence in the latest Neogene to Present, so ultimately this slip must be rooted with the larger subduction boundary. Figure 9 shows the Fila Costeña in the context of a cross section from offshore Osa across the Talamanca Range to the north Panama thrust belt along the Caribbean coast. The top of the subducting slab is constrained by hypocenter locations [Protti et al., 1995], and the position of the leading edge of the subducting slab is based on projection of the

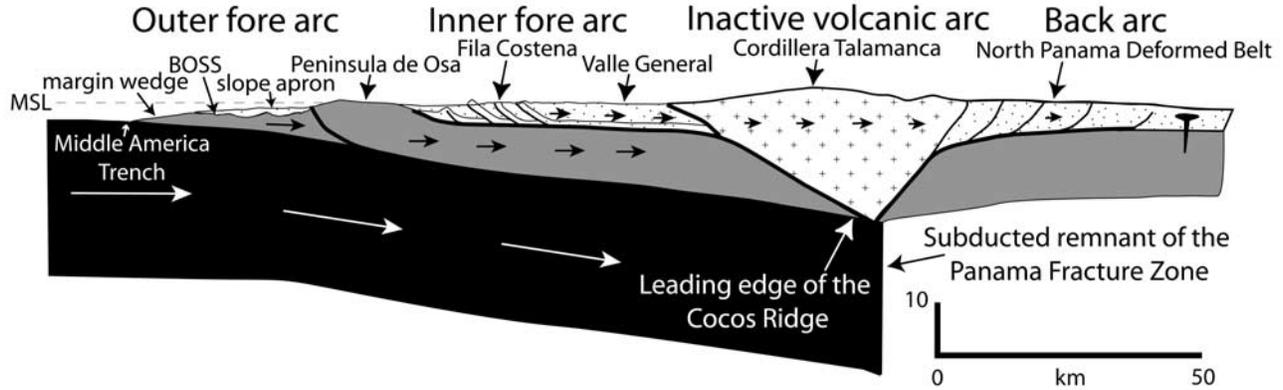


Figure 9. Regional cross section from the Middle America Trench to the north Panama deformed belt. Small arrows show velocity relative to the pin line on the Caribbean plate. BOSS, base of slope sediments reflector.

north Panama fracture zone northward to where it intersects the line of section. Since the coastal mountains are separated from the Talamanca Range by a valley (i.e., the Valle General), we suggest that the décollement that delaminates the forearc basin from the underthrusting forearc continues as a flat beneath the valley but links with the crustal-scale ramp system that has resulted in uplift of the Talamanca Range, the highest peaks in Central America. An outcome of this model is that all the upper plate rocks of the cross section from the Middle America Trench to the frontal thrust of the north Panama thrust belt (Figure 9) are converging with the Caribbean plate at a velocity that decreases to the northeast [see Cloos and Shreve, 1988]. These upper crustal rocks compose the Panama block of Fisher et al. [1994] and Marshall et al. [2000].

[30] Another issue highlighted by these results is the apparent dichotomy between the behavior of the external and internal portions of the forearc. The outer forearc experiences extension, arcward retreat of the trench, crustal thinning, and long-term subsidence of the upper slope, whereas the inner forearc simultaneously records crustal thickening, seaward advance of the thrust front, and rock uplift. This dichotomy between the inner and outer forearc may be a fundamental attribute of erosive margins. For example, along the Japan Trench, active uplift of the two major exposed inner forearc metamorphic massifs (i.e., the Abakuma and Kitakami massifs) occurs directly inboard of areas of outer forearc subsidence and scarring by incoming seamounts [e.g., Yamazaki and Okamura, 1989; Lallemand et al., 1992; Koike and Machida, 2001].

[31] This arcward change in behavior may reflect the response of an arcward thickening wedge to the subduction of seamounts and ridges. In the outer forearc, where the thickness of the upper plate is several kilometers or less, the margin wedge must deform in response to the subduction of bathymetric features with 1 km or more of relief [Sak et al., 2004a]. If seamounts composed of thick basaltic crust behave rigidly or are stronger than the leading edge of the upper plate, then the upper plate must experience leading-edge-up shear as the edge of the seamount subducts (Figure 10). In

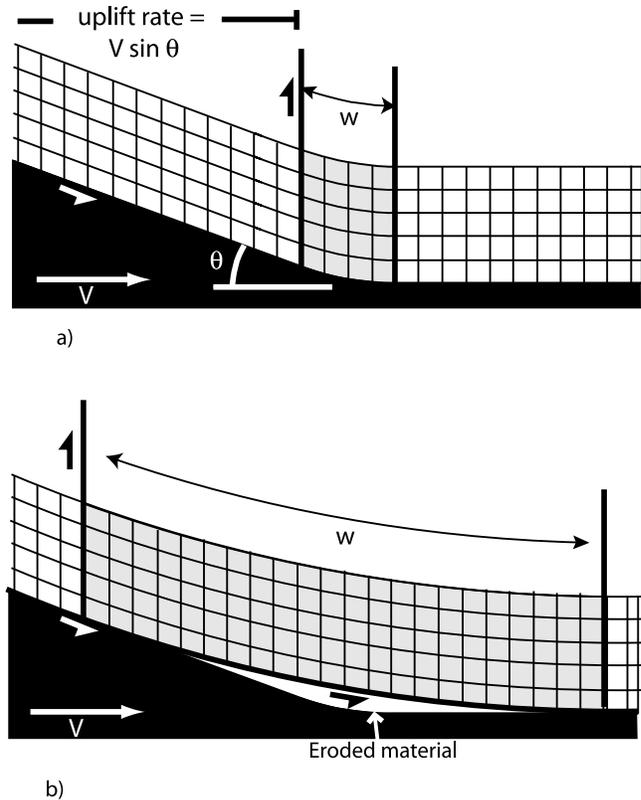


Figure 10. (a) Cartoon showing deformation of the upper plate in response to a large, rigid, bathymetric feature (in black) subducting at the convergence velocity, V [after Knipe, 1985]. This deformation involves leading-edge-up shear in the upper plate, with deformation restricted to a region (shaded) where the incoming plate has curvature. The width of this deforming region parallel to the relative plate motion vector is the straining distance, w . The surface uplift over the incoming feature depends on the slope and convergence velocity but could also reflect global shortening or upward dissipation of the effects of the asperity. (b) Hanging wall shortcut that increases the straining distance for a given ramp angle. Upper plate material detaches at the base of the wedge and couples with the subducting plate.

such a system the strain rate in the upper plate is a function of the slope of the incoming feature and the duration of the strain accumulation. This duration depends on the convergence rate and straining distance [e.g., *Knipe*, 1985], a distance that is a function of the curvature at the base of large incoming bathymetric features.

[32] Thus upper plate rocks must deform as they adapt to changes in slope on a rigid, incoming plate. If the slope of the seamount is too steep or the curvature is too angular for flow of the upper plate through the bend, then the hanging wall could fail pervasively or develop a hanging wall shortcut that increases the straining distance and decreases the required strain rate [*Knipe*, 1985]. Formation of a hanging wall shortcut in such a system is a mechanism of subduction erosion that detaches material from the hanging wall and couples it to the downgoing plate along the leading and trailing edges of subducting asperities. Along accretionary margins a ramp in the plate boundary can lead to footwall collapse of the weak overpressured subducting pile and accretion of duplexes [*Sample and Fisher*, 1986]. In contrast, the subducting plate along erosive margins is typically stronger than the upper plate, so incoming asperities lead to hanging wall collapse and basal erosion of the forearc. Thus the occurrence of accretion or erosion along a plate boundary may depend on the relative strength of the upper and lower plates in the vicinity of ramps or irregularities in the plate boundary geometry.

[33] The passage of seamounts beneath the outer forearc wedge causes the zone of leading-edge-up shear to advance arcward. As this zone encounters the base of the inner forearc, the upper plate is thicker and potentially stronger (e.g., the edge of the Cretaceous Caribbean basement), and there is more work done in uplifting and deforming a thicker plate to allow subduction of asperities. For a given wavelength and amplitude of asperities on the lower plate, there may be a threshold wedge thickness at which point the lower plate fails and seamounts are sheared off (i.e., underplating) or the upper plate fails and new faults propagate to the surface. Both processes ultimately result in down-dip smoothing of the plate boundary.

[34] The strength of asperities on the downgoing plate likely depends on the wavelength of roughness parallel and

perpendicular to the relative plate motion vector. Thus large ridges oriented perpendicular to the margin (e.g., the Cocos Ridge) continue to penetrate beneath the arc and to produce upper plate shortening in the inner forearc and uplift of the outer forearc. Smaller features such as conical seamounts can be detached from the downgoing plate and can be accreted arcward of the zone of outer forearc erosion to produce local centers of uplift and crustal thickening. Underplating of seamounts would explain the corrugated uplift pattern in Costa Rica to the west of the Fila Costeña, where discrete, uplifted blocks that expose forearc basement lie inboard of subducting seamount chains [*Fisher et al.*, 1998; *Gardner et al.*, 2001].

5. Conclusions

[35] The Fila Costeña is an inner forearc thrust belt that records a minimum of 17 km of shortening in the inner forearc landward of an outer forearc that, with the exception of Osa Peninsula, generally shows evidence of tectonic erosion and subsidence. The structure of the belt is characterized by a series of southwest directed thrust faults that strike parallel to the coastline and uplift the Tertiary forearc basin sequence of the Térraba basin. Cross sections depict a décollement at ~4 km near the contact between the Eocene Brito Formation and the underlying basement of the margin wedge. Thus the shortening of the Fila Costeña has allowed for underthrusting of the outer forearc. Observations of marine Pliocene strata, Quaternary fluvial terraces, and a Holocene marine terrace indicate that the thrusting began in the latest Neogene-Quaternary and is ongoing at shortening rates between 4 and 10 m/kyr and uplift rates of up to 1.5 m/kyr. On the basis of these data a cross section of the margin suggests that the décollement of the Fila Costeña joins the deeper subduction system along a ramp or system of ramps at the southwest edge of the uplifting mountains of the Cordillera Talamanca.

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