STRUCTURE AND DEFORMATION HISTORY OF THE NORTHERN RANGE OF TRINIDAD AND ADJACENT AREAS

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Abstract. Conflicting models have been proposed for both the evolution of northern South America and the neotectonics of the south Caribbean plate boundary zone. The Trinidadian portion of the margin is particularly controversial, but surprisingly it has been little studied. We present a structural analysis of Trinidad’s Northern Range, pertinent updates of the island’s stratigraphy and sedimentology, and new zircon fission track age determinations, and use them to constrain Trinidad’s geologic history, and to better understand the controlling tectonic processes. In our interpretation Trinidad’s three E-ENE striking ranges, which are separated by late Neogene-Recent depocenters, expose (1) the Northern Range Group, generally greenschist-metamorphosed Upper Jurassic to Cretaceous north facing continental slope sediments of the Northern Range, deposited on the northern South American passive margin 200-400 km to the WNW, and (2) the Trinidad Group, Cretaceous-Paleogene shelf slope sediments of the central and southern Trinidad deposited less than 100 km WNW of their present location. A small allochthon composing the Sans Souci Group Cretaceous tholeiitic volcanoclastic, basaltic, and gabbroic rocks (Sans Souci Formation) and sediments (Toco Formation) now in the northeastern Northern Range, has been transported hundreds of kilometers from the west with the Caribbean Plate. Despite earlier references to Cretaceous orogenesis, all deformation in Trinidad is of Cenozoic age. The first deformation in the Northern Range (D1) formed north vergent nappes and induced greenschist metamorphism, probably in the Late Eocene or Oligocene. The nappes developed either by the underthrusting of the Proto-Caribbean crust beneath South America due to convergence between North and South America, or as gravity slides caused by oversteepening induced by this convergence and/or the passage of the Caribbean Plate’s peripheral bulge and arrival of its foredeep. Northern Range D2 deformation is south vergent and represents the incorporation of Northern Range metasediments into the Caribbean accretionary prism. The transition to D3 brittle transpressive right-lateral strike-slip faulting is interpreted to be due to the uplift and east-southeastward transpressive emplacement of Northern Range/Cretaceous prism rocks onto the South American stepped shelf. This emplacement formed the Miocene transpressive thrust belts and foreland basin in central and southern Trinidad. In the final phase of Northern Range deformation (D4) -E-W normal faults and shear zones and conjugate NNW-SSE and NE-SW normal faults developed, and displacement on preexisting -E-W right-lateral strike-slip faults continued. The 11 Ma Northern Range zircon fission track ages suggest rapid uplift from the Late Miocene to Recent. Late Miocene subsidence of the Tobago platform immediately to the north of the Northern Range, and greater than 3 km of normal, down to the north, displacement indicated for the North Coast Fault Zone separating the Northern Range and Tobago platform, leads us to postulate that the rapid uplift of the Northern Range was in response to the northward detachment of the Tobago platform from above the Northern Range, along the north-dipping transtensional North Coast Fault Zone. This Late Miocene change in deformation style can be explained by a change from Caribbean/South American right-lateral transpression to right-lateral strike-slip generally striking 080°. This has generally induced a component of extension on pre-existing faults striking at greater than 080°, and a component of compression on faults striking at less than 080°.

INTRODUCTION

Northern South America comprises a Jurassic-Cretaceous passive margin section overlain by northward thickening foreland basin deposits which have been overthrust by allochthons of metamorphic rock at debated times [Maresh, 1974; Gonzalez de Juana et al., 1980; Pindell et al., 1988]. One type of model [e.g., Bellizzi, 1972; Maresh, 1974; Boets et al., 1984; Chevaller et al., 1988] considers that arc collision and metamorphism occurred along northern South America in the Cretaceous and was followed by Cenozoic mainly right-lateral relative displacement of the Caribbean and South American plates. Another model [e.g., Duncan and Hargraves, 1984; Robertson and Burke, 1989; Erlich and Barret, 1990] considers that Cretaceous arc collision and metamorphism occurred along the western margin of South America and that fragments of that orogen were subsequently displaced and emplaced by mainly strike-slip motions across northern South America in the Cenozoic. A third model [e.g., Mattson, 1984; Pindell and Barret, 1990; Snake, 1991] produces the metamorphism through arc-polarity reversal in the Pacific and then translates the metamorphic terranes with the Caribbean Plate and obducts them, with only partial reheating, onto the northern South American margin in Cenozoic times. Only the second and third types of models fit modern regional kinematic frameworks [Pindell et al., 1988]. Each model is testable at specific locations along northern South America.

The paleotectonic development of Trinidad (Figure 1) has been considered to encompass tectonism in the Late Cretaceous, Paleocene, and Eocene [Kugler, 1953; Barr, 1963; Tyson et al., 1991], followed by Neogene folding, thrusting and right-lateral strike-slip [Kugler, 1953; Farfield and Bally, 1991; Payne, 1991; Tyson et al., 1991; Robertson and Burke [1989] and Erlich and Barret [1990] assumed Cretaceous metamorphism in the Northern Range but translated that block from far to the west, with no Cretaceous tectonism in the remainder of Trinidad. Algar and Pindell [1991] doubted the evidence for Cretaceous deformation in the Northern Range, postulating that it was deposited closer to present-day Trinidad because of the absence of any indication of Caribbean plate interaction until the Cenozoic. Pindell et al. [1988, 1991] and Pindell and Barret [1990] showed, from sediment accumulation and marginal subsidence history, that a large tectonic load, that is, arc continent collision and foredeep
development, did not occur in Trinidad until the Miocene. Pindell et al. [1991] suggested that Paleogene tectonism may have been due to the onset of convergence between North and South America, when the Caribbean Plate was still far to the west.

In this paper, we present new data, particularly from the Northern Range, that allow us to assess the geological history of Trinidad. We present detailed structural analyses and consider them in conjunction with our recent advances in the understanding of the island's stratigraphy and sedimentology, new zircon fission track dates, and new views on some significant regional considerations. A proper understanding of the geological history of Trinidad is important because it further constrains the evolution of the eastern south Caribbean plate boundary zone (SCPBZ), and northern South America.

**STRATIGRAPHY**

Mapping of the Northern Range [Kugler, 1961; Potter, 1974; Algar, 1993] along with new sedimentological observations and interpretations in southern Trinidad [Algar, 1993] have distinguished three stratigraphies (Figures 1, 2, and 3a):

**Northern Range Group**

The Northern Range Group incorporates Jurassic and Cretaceous passive margin sedimentary rocks of northern South America, metamorphosed to greenschist facies [Hutchison, 1939; Barr, 1963; Saunders, 1972; Potter, 1974; Frey et al., 1988; Algar, 1993] (Figures 3, and 4). Though the field-scale deformation is complex, lithostratigraphic formations have a simple map pattern (Figure 3b). Formation contacts are rarely visible, but the units have consistent thicknesses (Figure 3b), and lateral variations in rocks type are due to either sedimentary or metamorphic facies changes (Figure 3a). Structural reconstructions of these formations (Figure 3b) suggest the same age sequence as that indicated from the fauna, indicating that the Northern Range Group was once a cohesive sediment package [Kugler, 1961; Barr, 1963; Saunders, 1972]. Sandstones, slates, metaquartzites, and phyllites are the most common rock types, but limestones, cherts, and conglomerates also occur. Evaporites outcrop locally along the southern flank of the Northern Range [Kugler, 1953; 1961] and have been cored in the Gulf of Paria [Bray and Eva, 1983], where they intrude heavily fractured Lower Cretaceous limestones, possibly indicating a diapiric
and/or tectonic upwelling from an unexposed Jurassic and/or Lower Cretaceous unit [Algar, 1993] (Figure 2).

The Upper Jurassic [Hutchinson, 1939] Maraval Formation limestones are generally well bedded, with phyllic (micaceous) layers, and rare beds of calcareous slates. Massive limestone beds, up to 100 m thick, wedge out laterally over a few kilometers. The limestones locally contain algal and oolitic remains [Potter, 1968], but the metamorphism has recrystallized the limestone, obliterating sedimentary structures. The depositional environment is thus uncertain, and the formation may represent shallow water platform limestones and/or calciturbidites [Potter, 1974; Algar, 1993].

Overlying the Maraval, the interbedded quartzites and slates of the Maracas Formation grade laterally into slates, siltstones, and conglomerates of the Guayarama Formation. The Maracas is interpreted as a sequence of mid fan overlapping depositional lobes on the slope apron, while the Guayarama is more characteristic of the upper mid fan, where the channel fill gravels in a fan lobe surrounded by interbedded sandstones and slates representing levee deposits [Algar, 1993]. The Rio Seco, Lopinot and Laventille Formations are lateral facies equivalents of one another (Figure 3a). The coarse sandstones of the Rio Seco Formation interfinger with slates and are interpreted as meandering channels on the mid fan [Algar, 1993]. To the west in the Lopinot Formation the sandstones are finer, and siltstones are more abundant than in the Rio Seco. The Laventille Formation comprises relatively unmetamorphosed shales with a deepwater fauna, surrounding variably oriented shallow water limestone blocks, interpreted to have slumped into the shales on the continental slope [Potter, 1974; Algar, 1993]. Though sedimentation appears to have been continuous throughout the Cretaceous, the Aptian through Campanian is represented by a relatively thin stratigraphic sequence. This has been observed elsewhere on the northern South American passive margin [e.g., Ghosh, 1984] and may be a condensed section caused by the Middle Cretaceous eustatic sea level high. The slight sea level lowering that occurred in the Maastrichtian was probably responsible for the sudden influx of upper fan clastics of the Galera Formation.

From the above discussion, and a consideration of sedimentation rates and regional geology [Algar, 1993], we conclude that the Upper Jurassic through Upper Cretaceous Northern Range Group stratigraphy was deposited on a passive margin. The quartz, mica, and feldspar-rich mineralogy, along with the lack of arc volcanic debris imply that the depocenter was along the northern South American margin, removed from the Andes and Caribbean, less than 700 km WNW of present-day Trinidad, and probably between 200 and 400 km WNW because of stratigraphic similarities to eastern Venezuela’s Serrania del Interior [Algar, 1993].

Sans Souci Group

The Sans Souci Group comprises the Sans Souci and Toco formations and forms an allochthonous terrane distinct from the Northern Range Group [Algar and Pindell, 1991] (Figures 1, 2, 3a, and 3b). Three main aspects distinguish the two groups. First, the Sans Souci Formation contains basaltic, brecciated basalts, dolerites and gabbros of mid-ocean ridge basalt (MORB) chemistry [Wadge and Macdonald, 1985]. Turbiditic shales, quartzo-feldspathic sandstones, and conglomerates of the Toco Formation are interbedded with the igneous units, Zircons (fission track age = c. 100 Ma) from the Toco Formation (sample T4, Figure 3b) are interpreted to be derived from, and thus date, at least part of the crystallization of the Sans Souci volcanic rocks as Albanian [Algar, 1993]. South America's Proto-Caribbean passive margin formed in the Jurassic [Pindell, 1983]; therefore, the Sans Souci Group probably derives from either a younger (between South America and Yucatan), now subducted part of the Proto-Caribbean seafloor, or the far-traveled Caribbean Plate. In the
former case, the SANS SOUCI Group may be an elevated crustal fragment originally accreted to the Caribbean accretionary prism. With the exception of rare tholeiitic volcanic beds in the Maracas Formation [Potter, 1974; Jackson et al., 1991], no igneous units have been found elsewhere in Trinidad. The igneous rocks of the nearby island of Tobago are quite unlike those of the SANS SOUCI Group, being island arc in origin [Wadge and Macdonald, 1985; Frost and Snake, 1989; Jackson et al., 1991]. Second, Toco Formation shales have no metamorphic cleavage, unlike the slates and phyllites in the Northern Range Group, indicating a different deformation history. Third, metamorphic grade in the SANS SOUCI Group (prehnite/pumpellyite) is significantly lower than that in coeval rocks of the Northern Range Group (greenschist) [Frey et al., 1988; Algar, 1993].

The boundary between the SANS SOUCI and Northern Range groups is a broad fault zone (Toco-Grande Riviere fault zone, discussed below) which Algar and Pindell [1991] interpreted as a splay from the North Coast Fault Zone, 2 km to the north [Robertson and Burke, 1989] (Figure 1). Similar allochthonous oceanic rocks (El Copey metavolcanics and parts of Uquire Formation) occur along the western continuation of the North Coast Fault Zone, the Coche Fault Zone, along the north coast of the Araya-Paria Peninsula (Figure 1) [Gonzalez de Juana and Munoz, 1971].

**Trinidad Group of the Central and Southern ranges**

In the Central and Southern range’s structural highs (Figures 1 and 2), Cretaceous through Miocene rocks are juxtaposed against Miocene to Recent deposits of the surrounding lowlands [Kugler, 1961]. These two ranges are traditionally viewed as thrust fronts in an SSE vergent foreland fold, and thrust belt active mainly during the Miocene [Kugler, 1961; Speed, 1985; Tyson et al., 1991]. High angle faults coincident with, and possibly crosscutting, these thrust fronts,
Fig. 3b. Geological map of the Northern Range of Trinidad after Algar [1993]. Unit positioning in the western Northern Range modified from Potter [1974]. Fission track sample sites are those of Algar [1993]. A-D are north-south cross sections through the entire Northern Range showing dominant D1 north vergent nappes, whilst E-F are detailed cross sections across the fault contacts of the Maracas and Lopinot Formations in the southern Northern Range, showing D2 backthrust and D4 normal faulting.
have been variably interpreted as reverse faults and right-lateral strike-slip faults [Kugler, 1961; Payne, 1991] (Figures 1, and 2). The existence of presently nonquantifiable strike-slip offsets on these faults through the Central and Southern ranges has potentially altered the pre-Late Miocene paleogeography in and out of north-south cross section. On the basis of estimates of strike-slip offset at the Venezuelan portion of the El Pilar fault [e.g., Vierbuchen, 1984] and offsets through the Gulf of Paria pull-apart basin and faults within Trinidad (see below), it becomes apparent that elements of the Central and Southern ranges may have been laterally translated from the west by a total of 40 to 70 km.

Separate lithostratigraphies were developed for the Northern Range and central/southern Trinidad (Northern Range versus Trinidad Groups) because only the Northern Range rocks are discernibly metamorphosed [Kugler, 1953, 1961]. Both stratigraphies are dominated by slope turbidites and interchannel shales and hemipelagites, and may have been deposited on the same passive margin. The Cuche Formation is lithologically and faunally identical to the Rio Seco and Lepota Formations [Algar, 1993], and includes limestone blocks like those in the Lavenfille [Barr, 1952]. The organic rich siliceous mudstones of the Naparima Hill Formation were probably deposited on the shelf by upwelling currents during the same sea level high that slowed deposition further offshore in the Northern Range depocenter [Algar, 1993]. From the Maastrichtian through the Late Eocene, mudstones are dominant with periodic influxes of sandstone. The first evidence for a significant change in depositional depth is in the Late Eocene/Early Oligocene when shallow water limestones and conglomerates of the San Fernando Formation developed locally, in association with a slowing of deposition [Kugler, 1953; Algar, 1993]. These deposits are interpreted to have resulted from the passage of the Caribbean flexural forebulge, and/or margin buckling due to convergence of North and South America, possibly aided by eustacy [Pindell et al., 1991; Algar, 1993]. In the Late Oligocene, initial deposition of a thick sequence of shales and siltstones, including the Pointe-a-Pierre [Algar, 1993], Nariva and Cipero Formations, marks the onset of Neogene foredeep subsidence [Pindell et al., 1991]. We postulate that an eastward forebulge/foredeep migration ahead of the Caribbean may have controlled Robertson and Burke’s [1989] eastward shift of deltaic sediments supplied by the Orinoco from the west. Since the Late Miocene the Orinoco delta (Figure 1) has supplied much of southern Trinidad and the surrounding area with up to 7 km of deltaic and nearshore mudstones, siltstones, and sandstones.

NORTHERN RANGE STRUCTURAL STYLES

Phase 1 (D1)

Most Northern Range rocks possess a fabric parallel pressure solution cleavage. This cleavage is almost always folded by the first generation folds, but there are very rare examples in which this does not occur, and the first cleavage is axial planar to these folds. To accommodate this feature, we assign bedding S0, first folds F1, axial cleavage to F1 folds S1, and the bedding parallel cleavage folded by F1 folds S0. Prehnite-pumpellyite and greenschist facies metamorphism [Frey et al., 1988] occurred during the development of S0 and S1, since both are slaty, phyllitic or schistose.

F1 folds are tight to isoclinal, asymmetric, Ramsay type Ic, with a 0.1 to 2 m wavelength, and 0.1 to 1 m amplitude (Figure 4a). Despite their tightness, S1 axial cleavage is very rare. Intrafolial folds are common in the Maraval Formation. Stretching lineations often parallel F1 axes (Figure 5).

F1 vergence is northward in upright, south facing folds (25 of 170 F1 axes), and southward in overturned, north facing folds (140 of 170 F1 axes). This F1 vergence and the stratigraphic overturning of the center of the Maraval Formation (Figure 3b) imply that the minor folds are parasitic on an antcline overturned to the north. This structure is also indicated by the stratigraphic repetition of units either side of the Maraval Formation [Potter, 1973], and the nonvergence of F1 folds within the Maraval Formation near the core. The greater thickness of Maracas and Guayaramara Formation north of the anticlinal axis (Figure 3b) is interpreted as the overturned limb of a second, contemporaneous, anticlinal nappe decapitated by the main antclinal nappe to the south. Nappe hinges plunge gently (<15°) but frequently switch plunge direction from east to west (Figure 3b).

F1 axes in the main Northern Range (Figure 5a) broadly mimic the E-W trend of the D1 nappes (Figure 3b). However, in the eastern Northern Range domain (ENRD), and southern Northern Range domain (SNRD), F1 axes trend NE-SW, respectively, 45° and 15°, counterclockwise of the D1 nappe hinge trend in each region. This variation is also shown by the S0 strike in the SNRD and main body. F2 axes are distributed similarly to F1 axes, but S2 strikes consistently ENE in all areas. These data indicate that the ENRD and SNRD were rotated counterclockwise relative to the main body by up to 45°, at some time during D1 and/or D2. It is probable that this occurred during the D2 convergence discussed below, as the main Northern Range body was thrust and rotated under the SNRD and ENRD (Figure 2). In addition to the ENE trend in the ENRD, the D1 nappe hinge trend varies from its E-W course in the western Northern Range, where it steps southward and vertically down (Figure 3b), resulting in the loss of the Maraval Formation outcrop.

Incorporation of the Maastrichtian Galera Formation into all D1 deformation, including nappe formation (Figure 3b), requires that D1 be post-Cretaceous [Algar and Pindell, 1991]. The northward vergence of the Northern Range nappes is contrary to the southward vergence of the main compressive deformation in northeastern South America (Caribbean overthrusting South America), e.g., Maresch, 1974; Speed, 1985; Stephen, 1985; Pindell and Barrett, 1990]. As will be shown below, these D1 nappes are overprinted by D2 south vergent structures. Thus they are unlikely to be synchronous backthrust-related structures and probably formed prior to the south vergent deformation. *Ar* / *Ar* mica dates [Speed and Foland, 1990; Foland et al., 1992] and zircon fission track ages (Figure 3b; Algar, 1993) both suggest two separate periods of cooling following metamorphism (Oligocene to Early Miocene and Middle-Late Miocene), which probably correspond to D1 and D2.

We visualize two possible mechanisms by which the northward vergent nappes could have formed: (1) northward vergent thrusting caused by the Middle Eocene onset of plate convergence between North and South America [Pindell et al., 1988; Pindell et al., 1991], and (2) large-scale gravity sliding of oversteepened continental slope sediments after the passage of the Caribbean Plate's peripheral bulge and arrival of its foredeep/trench depression.
From the Middle Eocene to the Recent, as much as 150 km of convergence between North and South America may have been accommodated in the region of Trinidad. Farther west, away from the pole of plate rotation, as much as 300 km may have been accommodated [Pindell et al., 1988]. Pindell et al. [1991] postulated that this convergence caused buckling of the northern South American margin, and eventually the subduction of the Proto-Caribbean crust beneath northern South America. Subduction initiation may have been aided by the load placed on the Proto-Caribbean from the Caribbean Plate approaching from the west. Pindell et al. [1991] noted that the shallowing observed throughout northern South America around this time may have, at least partially, resulted from intraplate shortening, or buckling prior to subduction, and also cited tomographic images [Van der Hilst, 1990] that implied the presence of a subducted slab beneath northern South America. Subduction remained amagmatic because rates of convergence averaged only 6-7 mm/yr. The Northern Range northward vergent nappes may have formed within the accretionary prism of this subduction complex after the Middle Eocene, but before the arrival of the Caribbean Plate to which it was accreted.

Alternatively, given that D1 nappes formed from sediments on the relatively steep, north facing passive margin slope, gravity sliding may have created the nappes. Many of the characteristics of the D1 nappes described above have been documented in gravity slides: (1) the lack of fold axial planar cleavage in most F1 isoclinal folds; (2) the sudden offsets of, and discordance of minor folds from, the main nappe axis; (3) intrafolial folds; and (4) the severing of the upper limb of underlying nappes [Helwig, 1970; Gawthorpe and Clemmey, 1984; Elliott and Williams, 1988; Webb and Cooper, 1988]. Because the F1 minor folds fold a pressure solution cleavage (S0i), the sediments must have been well lithified prior to their deformation. This is not consistent with the classical "soft sediment" aspects of slumping, but it is known from modern gravity slides, some of which show a high degree of internal cohesion in seismic sections, for example the Mexican Ridges of western Gulf of Mexico [Buffler et al., 1979] and the western margin of Africa [N. Zitellini, personal communication to JLP, 1991]. Gravity sliding may have been triggered by slope oversteepening after passage of the Caribbean peripheral bulge and progressive arrival of the foredeep basin ahead of the ESE-migrating Caribbean Plate.

Fig. 4. Structural styles in the Northern Range. (a) F1 tight asymmetric minor fold folding S0i, foliation. North to the right, in overturned limb of D1 anticline. Lens cap is 55 mm diameter. (b) Open, symmetric F2 fold with gently dipping fold axial plane refolding tight asymmetric F1 fold. North to the right, on upright limb of anticline. Hammer is 32 cm long. (c) S2 cleavage crenulating S0i cleavage in the Galera Formation slates. North to the right. (d) Subvertical D3 brittle fault zone dissecting S0i, foliation planes and tight, F1 fold. Buckle folding adjacent to fault is contemporaneous with D3 faulting. Right-lateral strike-slip fault displacement implied from subhorizontal slickenside lineations on nearby faults. Hammer is 32 cm long. (e) D4 extensional high-angle shear band.
(Eocene-Early Miocene depending upon the location of the Northern Range depocenter) [Pindell and Barrett, 1990], or by the margin buckling from the Middle Eocene to Recent convergence of North and South America noted above. Sliding may have been facilitated by an evaporitic horizon which has been postulated to underlie the Northern Range sequence [Algar, 1993].

At present we are unable to distinguish which of the two methods proposed above best explains the Northern Range northward vergent nappes, though the lower heat flow associated with underthrusting might hinder heating to the observed grade of metamorphism. It is possible that they acted together. Many of the features that favor the gravity slide interpretation may have been caused by the underthrusting. Inclusion of the Maastrichtian Galera Formation in the D1 nappes, together with the zircon fission track [Algar, 1993] and 40Ar/39Ar cooling ages [Foland et al., 1992] broadly confine D1 to Paleocene through Early Miocene. Because the S0 cleavage occurs throughout the Maastrichtian Galera Formation as well as in the older units, the Galera was once buried beneath several kilometers of sediment prior to D1, implying that D1 occurred well into the Paleocene-Early Miocene time span, probably in the Late Eocene or Oligocene. Assuming that Caribbean interaction with the Northern Range sediments (accrretion of slope sediments to prism) should have produced a recognizable deformation, the Eocene relative position of South America and the Caribbean (e.g., reconstruction of Pindell et al. [1988]) can be used to define the maximum westward position of the depocenter. Using this logic, we suggest that the original Northern Range Group depocenter was no farther than 600-700 km to the west of Trinidad today. Comparing the Northern Range Group to northern South America’s passive margin section, the best correlations are made with eastern Venezuela’s Serranias del Interior [Algar, 1993]. Thus, it seems likely that the Northern Range Group was deposited about 200-400 km to the west of its present location.

Phase 2 (D2)

D1 structures are deformed by open folds and a crenulation cleavage assigned to the same, D2, phase of deformation, because they are both deformed by all later structures. F2 folds are symmetric, open and similar, quite unlike the tight to isoclinal, asymmetric F1 minor folds that they refold (Figure 4b). This style of folding occurs throughout the Northern Range but is particularly well developed along the north coast. In addition to their larger apical angle (>70°-120°), F2 folds are distinguished from F1 folds by their larger wavelength (2-10 m versus 0.1-2 m), larger amplitude (1-3 m versus <1 m), and the presence of an axial planar cleavage that often dips to the north (Figure 5a). An S2 crenulation cleavage was commonly observed without associated F2 folds and is plotted separately from F2 fold axial planar cleavage but has a similar orientation (Figure 5a). Both dip predominantly north, oblique to the south dipping S0 (Figures 4c, and 5a). This trend shows no significant growth of metamorphic minerals along S2, implying that the majority of metamorphic minerals (mainly white micas) grew parallel to S0 during D1. However, both zircon fission track and 40Ar/39Ar methods suggest a second period of heating and cooling after D1, from ~350°C at ~15 Ma [Speed and Foland, 1990] through 200°-250°C at 11 Ma [Algar, 1993]. D2, the last fully ductile deformation, therefore likely occurred around 15 Ma.

The northward dip of the cleavages, the SSE vergent asymmetry of some F2 folds, and the relative abundance of D2 structures along the north coast of the Northern Range are consistent with a SSE vergent overthrust onto the Northern Range.

Phase 3 (D3)

Faults and gentle folds are assigned to the third phase of deformation (D3) because they deform D2 structures but are themselves deformed by later (D4) structures. The faults are the most pervasive structure, severely disrupting D1 and D2 structures, especially in the eastern and southern Northern Range (Figures 4d, and 5b). Fault surfaces are generally sharp and well defined by a thin (<5 mm) fault gouge or, less commonly, slickenside lineations. Faults are frequently steep (>70°), and tend to anastomose within the outcrop in both strike (±30°) and dip (±20°). Block rotation about subvertical axes from shear between two faults, asymmetric drag folding due to shear along a single fault, and symmetric folding due to compression between fault planes is common, and together with subhorizontal slickenside lineations and calcite shear vein lineations imply a major component of right-lateral transpression along and between the faults. It is not possible to quantify these fault displacements because of a lack of measurable lineations. Though there is a broad spread in the orientation of individual faults (Figure 5b), NE to ENE strikes predominate. The spread is likely due mainly to fault planes anastomosing, whilst the similarity of D3 fault and S0 foliation strikes implies a strong dependence on the preexisting structural grain. D3 fault orientation is thus unlikely to be indicative of a paleostress field.

The magnitude and relative proportion of compression and strike-slip are difficult to ascertain for this transpressional phase of deformation because the lack of marker horizons and small outcrop size hinders recognition of fault offsets greater than a few meters. The greater intensity of faulting and block rotation in the eastern and southern Northern Range (Figure 5b) implies that these regions are either more strongly or more pervasively affected by the transpression than the rest of the Northern Range. This may be due to the relatively less competent argilaceous (versus arenaceous) dominated rheology of the southern and eastern Northern Range?

Some larger outcrops show gentle folding of the S0 foliation with a wavelength of 5-20 m. The large wavelength precludes recognition in the majority of outcrops; therefore the data set in Figure 5 may not be a true representation of the relative pervasiveness of these folds. The 170°±20° azimuth average for the fold axes (Figure 5b) is inconsistent with the similarly striking normal faults described below and may indicate a change in regional stresses. L- and LS-teectonites are well developed along the north coast, where D2/D3 deformation is most intense, and the LS fabric crenulates S0. This region’s LS-teectonite fabric is thus related to D2 and/or D3 deformation. In relation to the stretching lineations observed elsewhere in the Northern Range as parallel to F1 minor folds, is uncertain.

Phase 4 (D4)

The latest Northern Range deformation formed steep (>60°), brittle normal faults and high angle shear bands (Figure 4c).
FAULT ZONES OF THE EASTERN SCPBZ

The relative roles of compressional, extensional, and strike-slip structures, and their implications for Caribbean-South American motion in the eastern SCPBZ has been the subject of numerous conflicting studies [Rod, 1956; Vierbuchen, 1984; Speed, 1985; Robertson and Burke, 1989; 1991; Erlich and Barrett, 1990; Speed et al., 1991]. Part of the problem revolves around incomplete exposure of important structural zones. We maintain, however, that the main problem is that all models to date assume and attempt to verify a single kinematic mode for the development of all structures. The four structural phases of the Northern Range described above are an indirect indicator of how the regional deformation history may have changed through time. These can be considered in conjunction with aspects of some of the larger-scale fault zones and adjacent areas in and around Trinidad to further elucidate the evolution of the eastern SCPBZ. It will be shown that transpression was replaced by right-lateral strike-slip, with the change occurring around the end of the Middle Miocene (~10 Ma). We suggest that this is why no single kinematic regime has satisfactorily explained all the structures of the area.

El Pilar Fault

The El Pilar fault is well documented as possessing significant (several tens of kilometers) right-lateral strike-slip (RLSS) in eastern Venezuela [e.g., Vierbuchen, 1984] (Figure 1). Although heavily debated, the traditional view that it also

Fig. 5a. Lower hemisphere equal area stereograms of some relevant structural elements of the Northern Range. Data are divided into structural domains that best show the variations in the orientation and relative abundance of structures. See Fig. 5b for key to structural domains.

Fig. 5b. Lower hemisphere equal area stereograms of some relevant structural elements of the Northern Range. Arrows on D4 faults/extensional planes indicate slip direction on hangingwall determined from slickenside lineations, or ductile extension lineation. Data is divided into structural domains that best show the variations in the orientation and relative abundance of structures.
continues into Trinidad between the Northern Range and the Caroni Basin as a major (tens of kilometers) fault of reverse and/or RLSS displacement is supported by (1) the relatively sudden topographic transition from the Northern Range mountains to the low-lying Caroni swamps (1000 m over ~10 km); (2) the metamorphic/nonmetamorphic transition from Northern Range rocks to those in the south; (3) the interpretation that thrust, normal, and strike-slip faults observed within Pleistocene deposits immediately south of the Northern Range indicate RLSS motion along an underlying El Pilar fault [Robertson and Burke, 1989]; and (4) the observation of flower structures in seismic lines off the east coast of Trinidad [Erlich and Barrett, 1990]. We, too, have found evidence for RLSS along the southern flank of the Northern Range in the form of the D3 right-lateral transpressive faulting, described earlier, and the outcrop pattern of the Galera Formation along the southern flank of the Northern Range, which resembles that of a RLSS duplex (Figure 3b). However, after full consideration of all the data, we conclude that the offset and regional significance of the high-angle RLSS El Pilar fault known from eastern Venezuela is unlikely to be present to anywhere near the same extent in Trinidad. First, although slip could be aseismic or locked, the zone in Trinidad is historically (70 years) seismically inactive [Peraza and Aggarwal, 1981; Speed et al., 1991; Russo et al., 1992] and shows no offset of Holocene alluvial fans or disturbance of Holocene sediments [Robertson and Burke, 1989]. Second, our study of recently shot seismic lines to the east of Trinidad (courtesy Trinidad Ministry of Energy) shows clearly that the flower structure fault zone claimed by Erlich and Barrett [1990] to be a continuation of the El Pilar is more likely to be continuous with the NE-SW striking faults that extend from the Central Range (Figure 1). Figure 6 shows an onland north-south seismic line across the Caroni Basin, with this same flower structure, that it is clearly related to the Central Range. Furthermore, the El Pilar zone itself has no reverse displacement but instead is a zone of significant normal, down to the south displacement of Cretaceous units. Strike-slip displacement cannot be ruled out, but the minor amount of disruption of Neogene sediments implies that it was a relatively small component in at least the Neogene. Thus although the normal faulting may have accentuated it, the topographic transition between the Northern Range and Caroni Basin is primarily a result of sedimentary onlap.

In summary, only in Venezuela is the El Pilar a major fault with tens of kilometers of RLSS from the Late Miocene to Recent. In Trinidad, it appears to have been a normal fault zone since the Late Miocene with only minor right-lateral displacement. The juxtaposition of the Northern Range against central and southern Trinidad was transpressive, and displacement was in our opinion most likely accommodated along a blind thrust beneath the Caroni Basin, possibly rooted on Jurassic/Lower Cretaceous evaporites (Figure 2), and at a later stage, along the Arima Fault (see below).

Arima Fault

The Arima fault (Figures 1, 2, and 3) anastomoses along the southern flank of the Northern Range, placing the Lopinot and Rio Seco Formations above the Maracas and Guayarama Formations. The rivers and roads that traverse across the boundary show overturning of the otherwise monotonously southward dipping and facing Maracas Formation within 1 km of the contact, in the style of a footwall syncline to a north directed thrust (Figure 3b, E-E', F-F'). North vergent thrusting along the Arima Fault is also implied by imbrication of the Rio Seco and Guayarama Formations in the eastern Northern Range (Figure 3b, D-D'). However, because the Lopinot

![Figure 6. Seismic reflection profile (two-way travel time, migrated) through the Caroni Basin showing northward onlapping of Neogene sediments onto the Northern Range. The El Pilar fault has a normal displacement as well as an indeterminate, but probably minor, right-lateral strike-slip component. Key: LK, Lower Cretaceous; MZ, Manzanilla Fm. (Upper Miocene-Middle Pliocene); SPV, Springvale Fm. (Middle Pliocene); LT, Lower Talparo Fm. (Upper Pliocene); and UT, Upper Talparo Fm. (Lower Pleistocene). Section and interpretation after Payne (1991).](image-url)
Formation is both younger and of slightly lower metamorphic grade than the Maracas Formation [Algar, 1993], a normal, down to the south, reactivation may constitute a more significant proportion of the displacement along the Arima Fault. Some element of lateral motion is also likely because D3 transpressive faults are common to the south of the fault. Because the S0, foliation and F1 folds are folded by the Arima fault's fault propagation folds, it is implied to have formed as a backthrust. It may have been induced by the D2 blind thrusting of the Northern Range into the Cretaceous rocks of central and southern Trinidad (Figure 2).

North Coast Fault Zone (NCFZ)

The North Coast Fault Zone (NCFZ) coincides with the boundary between South American and allochthonous Caribbean rocks (Figure 1) and has been shown from seismic reflection profiles to have considerable normal and right-lateral offsets [Robinson and Burke, 1989]. Onestrand alone of the NCFZ has accumulated 10 km of right-lateral offset since 1.6 Ma [Robinson and Burke, 1989], or roughly a third of the total Caribbean/South American E-W relative motion (assuming full rate = 15-20 mm/yr) [Pintell et al., 1988, 1991]. Because the NCFZ comprises a number of such strands, it may have been the dominant site of Plio-Pleistocene RLSS in the eastern SCPBZ. Historically, however, the near-surface NCFZ identified in these seismic reflection profiles is aseismic (last 70 years) [Peretz and Aggarwal, 1981; Speed et al., 1991; Russo et al., 1992].

A number of observations indicate that the normal, down to the north, component of offset is such that basement along the north flank of the NCFZ (Tobago terrane of Speed and Walker [1991]) was situated structurally above the Northern Range in the Middle Miocene: (1) The Sans Souci Group is interpreted to be an allochthon juxtaposed against the Northern Range Group along a strand of the NCFZ, the Toco-Grande Riviere Fault Zone [Algar and Pindell, 1991]; (2) An outcrop of dolerite 10 km south of the Sans Souci allochthon may constitute a klippen from the same allochthonous body [Barr, 1963]; (3) A subaerial erosional hiatus separates several kilometers of Late Miocene-Recent clastic rocks from Cretaceous basement in the southern, but not the northern, part of the Tobago terrane [Robertson and Burke, 1989]. This indicates that the southern part of Tobago terrane was uplifted and exposed above sea level prior to the Late Miocene; (4) Northern Range D2/D3 structures are best interpreted as the result of oblique overthrusting; (5) Middle-?Late Miocene Cunapo Formation proximal alluvial-fluvial conglomerates in the Caroni Basin (Figure 3a) contain cobbles and pebbles of basic igneous rock (Guaico #1 well) [Algar, 1993]. Likely source terrains for these are the Sans Souci Group, Tobago terrane, or equivalents. If so, then the Caribbean allochthons must have been higher than, and probably on, the intervening Northern Range at the end of the Middle Miocene.

If the Tobago terrane/Sans Souci Group (Caribbean accretionary prism/forearc?) was once subaerially situated above the Northern Range, a tectonic mechanism (in addition to erosion) is required to explain the relative subsidence of the erosional unconformity by 3-4 km to its present depth below sea level (Figure 2). The normal throw on the northward dipping NCFZ was probably primarily responsible for the northward withdrawal of the allochthons from above the Northern Range, in the process stranding the Sans Souci klippen. As Robertson and Burke [1989] noted, the E-W trend of the Northern Range puts it into transension in a zone of strike slip that appears from the structures in the NCFZ to trend 070° ± 15°. If this unloading occurred quickly, the Northern Range may have rebounded relatively rapidly to maintain isostatic equilibrium. Such a model is consistent with zircon fission track data [Algar, 1993] and 40Ar/39Ar mica data [Speed and Poland, 1990; Poland et al., 1992] that indicate rapid uplift of the Northern Range (~1.0 mm/yr), beginning in the Middle Miocene (15 Ma), synchronous with subsidence of the Tobago terrane immediately north of the Northern Range.

Toco-Grande Riviere fault zone and Sans Souci allochthon

The boundary between the Galera and Toco formations of northeastern Trinidad was regarded by Barr [1963] as an erosional unconformity with a basal lag breccia in the lower Galera, indicating to him a period of Middle Cretaceous orogenesis (Figure 3). We interpret the boundary as a major fault zone, with brecciation induced by motion along the boundary forming a tectonic mélangé [Algar and Pindell, 1991]. The eastern strand of the fault zone is well exposed along the coast, where blocks of Toco and Galera rocks up to 100 m across are juxtaposed in a multitude of orientations. Separating the blocks are cataclastic fault breccias up to 5 m thick, containing millimeter- to centimeter-sized clasts of sandstone, limestone, or chert, of both formations, enclosed in a matrix of mudstone with scaly cleavage (Figure 7). The foliation is commonly subvertical, and strikes between 045° and 100°, with a mean around 060°, subparallel to the mapped boundary between the two formations. Strain indicators such as clast rotation, slickenline lineations, vein offsets, and clast

Fig. 7. Typical subvertical cataclastic fabric of the tectonic mélangé in the Toco-Grande Riviere Fault Zone. Looking west at vertical plane. Lens cap is 55 mm in diameter.
long-axis orientation imply a dominance of lateral motion, of indeterminate sense, parallel to the fabric. D4 normal faults crosscut the melange fabric.

Similar structures occur to the west in the Grande Riviere river section, where the Sans Souci and Toco formations are juxtaposed against the Galena and older formations (Figure 3b). Lack of exposure prevents tracing of the fault zone between these two areas of outcrop, but since the only observed boundaries are faulted and, as noted above, the boundary separates two very different rock units, it seems likely that a continuous fault zone bounds the Sans Souci and Northern Range groups.

Because (1) D1 structures are absent from the Sans Souci allochthon but are deformed within the fault zone, and (2) D4 normal faults crosscut the melange fabric, juxtaposition of the Sans Souci and Northern Range groups was post-D1 and pre-D4 (i.e., D2 and/or D3). Zircon fission track analysis on a sandstone from the center of the eastern strand of the fault zone found what were interpreted as unannealed zircons dated at 33 ± 8 Ma (sample TFZ-1 in Figure 3b) [Algar, 1993]. This represents the maximum age of the sandstone (Oligocene), implying that the Toco-Grande Riviere fault zone was active, and D2/D3 occurred during and/or after the Oligocene.

Because the Toco-Grande Riviere fault zone is only 2 km to the south of the NCFZ and separates rocks similar to those separated by the NCFZ, it is likely to be a splay from the latter. As such, the Toco-Grande Riviere fault zone may once have formed a significant part of the eastern SCPBZ, accommodating relatively large amounts of RLSS and/or convergence between the Caribbean Sans Souci allochthon and the South American Northern Range Group, in accord with the very different early deformation histories of each area.

**Other SCPBZ deformation in and around Trinidad**

Rapid Middle Miocene (15 Ma) shoaling of the Central Range culminating in deposition of the Tamana Formation shallow water limestones, the identification of the Middle Miocene Rio Claro flysch, and the interpretation of imbricate structural styles under central and southern Trinidad,
oversteepening whose Oligocene Miocene), assumes re-entrants Interior. South sands 700 sediment similarities Pierre 8. reworked American Formation 8. Barbados's likely to the margin marked by (1) cooling of Northern Range by erosion and/or overthrusting onto cooler section (40Ar/39Ar age 15 Ma), [Speed and Folkand, 1990], and (2) thrusting of the Central Range to sea level as recorded by Tamana shallow water limestones [Erlich et al., 1993]. Strike-slip component of deformation recorded in Northern Range by D3; (f) Change in azimuth of Caribbean motion initiates E-W shear zone in this region of the SCPBZ. Structures throughout the region trending less than or greater than ~80° generally are compressional or extensional respectively, with variable RLSS components. The transtension on the NCFZ (Northern Range D4) allowed isostatic exhumation and further cooling as indicated by 11 Ma zircon fission track ages [Algar, 1993], and rapid subsidence and burial of previously exposed Tobago shelf (northward normal detachment off of Northern Range?); (g) Southward stepping of the EI Pilar RLSS fault zone along the NW-SE strike of a Jurassic marginal offset to E-W and ENE-WSW RLSS faults in central and southern Trinidad develops Caroni (CA) and Gulf of Paria (GP) pull-apart basins, similar to Late Miocene to Recent formation of Caraco basin to the west. Under this regime significant clockwise rotations are expected. Northern Range continues its rapid exhumation. Inset summarizes the relative motion history of the Caribbean (represented by Tobago) relative to northern South America.

Fig. 8. Model of tectonic evolution of the southeastern Caribbean in the vicinity of Trinidad and northeastern Venezuela incorporating the results of this study. Northern South America is divided into four tectonostratigraphic units whose deformations are independently estimated. Construction assumes 16 km/Ma Caribbean/South American relative motion at angles shown, 90 km of NW-SE shortening in Serranía del Interior [Passalacqua et al., 1991], and 50 km of shortening across southern Trinidad [Rohr, 1991]. (a) Jurassic to Eocene passive margin deposition. Though the northern South American margin trends broadly E-W, salients and re-entrants are suggested based on geometry of Tertiary thrust belt [Pindell and Erikson, 1993]. The Northern Range sediment depocenter was along this margin, somewhere up to 700 km to the west of the present position of the Northern Range, but likely 200-400 km away because of sedimentary similarities to the Serranía del Interior of eastern Venezuela [Algar, 1993]; (b) Passage of flexural forebulge ahead of the Caribbean Plate and/or underthrusting of the Proto-Caribbean beneath northern South America, raises the shelf. This induces D1 north vergent nappes either as gravity slides by oversteepening of the slope, or as tectonic nappes within an accretionary prism. This deformation and metamorphism probably occurred between the late Middle Eocene and Late Oligocene (40Ar/39Ar cooling ages Late Oligocene-Early Miocene), [Foland et al., 1992]; (c) Encroachment of the Caribbean Plate initiates SSE directed thrusting in Serranía del Interior. Barbados’s Scotland Formation probably comprises sediments reworked from infilled portions of the foredeep and Guyana Shield (regressive, prograding). Trinidad’s Pointe-a-Pierre Formation may record the onset of marine conditions after the passage of the forebulge (transgressive)?; (d) Onset of the major period of fold and thrust belt formation in Serranía, Central Range (CR) and Southern Range (SR), as the Caribbean Plate/prism overthrusts and pushes the Northern Range-Paria body up the continental slope (Northern Range D2); (e) Oblique obduction of Caribbean prism and Northern Range rocks onto the margin marked by (1) cooling of Northern Range by erosion and/or overthrusting onto cooler section (40Ar/39Ar age 15 Ma), [Speed and Folkand, 1990], and (2) thrusting of the Central Range to sea level as recorded by Tamana shallow water limestones [Erlich et al., 1993]. Strike-slip component of deformation recorded in Northern Range by D3; (f) Change in azimuth of Caribbean motion initiates E-W shear zone in this region of the SCPBZ. Structures throughout the region trending less than or greater than ~80° generally are compressional or extensional respectively, with variable RLSS components. The transtension on the NCFZ (Northern Range D4) allowed isostatic exhumation and further cooling as indicated by 11 Ma zircon fission track ages [Algar, 1993], and rapid subsidence and burial of previously exposed Tobago shelf (northward normal detachment off of Northern Range?); (g) Southward stepping of the EI Pilar RLSS fault zone along the NW-SE strike of a Jurassic marginal offset to E-W and ENE-WSW RLSS faults in central and southern Trinidad develops Caroni (CA) and Gulf of Paria (GP) pull-apart basins, similar to Late Miocene to Recent formation of Caraco basin to the west. Under this regime significant clockwise rotations are expected. Northern Range continues its rapid exhumation. Inset summarizes the relative motion history of the Caribbean (represented by Tobago) relative to northern South America.
have caused the Central and Southern ranges to be considered as south vergent imbricated thrust fronts (equivalent to Northern Range D2 dated by "Ar/Ar" of Speed and Poland [1990] at 15 Ma), mainly active in the Miocene [Kugler, 1961; Tyson et al., 1991], and accommodating as much as 50 km of NNW-SSE convergence (Rohr [1991], based on cross-section balancing). The Central Range has, however, also been cited as a center of Paleocene and Cretaceous tectonics because of the interpretation of the St. Joseph's conglomerate boulder bed in the Chaudefour Formation (Figure 3a) as a synorogenic conglomerate, and of Paleocene and Cretaceous melanges as wildflysch [Kugler, 1953]. However, these "wildflysch" are now thought to be tectonic and/or diapiric melanges of latter Neogene age, and the boulder bed is most probably a localized submarine erosional product [Algar, 1993]. The Late Eocene-Oligocene San Fernando Formation, discussed above, is the only unit produced by pre-Miocene deformation, and is attributed to flexural extension associated with the passage of a flexural forebulge ahead of the Caribbean Plate.

Recent seismic reflection surveys have identified a number of ENE striking high-angle faults in the Central and Southern ranges, with continuations offshore (Figures 1, and 2) [e.g., Payne, 1991; Farfan and Bally, 1991]. A component of strike-slip is indicated for these faults by subhorizontal slickenline lineations, and by the flower structure seen in at least the north flank of the Central Range (e.g., Figure 6). There is little control on the magnitude of total strike-slip offset, but strain is broadly distributed on numerous faults throughout the ranges. Judging from sedimentary units involved, this phase of deformation appears to have been active from the Pliocene or Late Miocene to Recent (7), after the Middle Miocene culmination of SSS thrusting.

Seismic sections through the Gulf of Paria and Caroni basins [Payne, 1991], (e.g., Figure 6) show normal faults striking in a variety of orientations, but mainly NW-SE and E-W (Figure 1). The geometry and timing of activity of these faults and the age and positioning of basin fill are best explained as a series of Late Miocene to Recent pull-apart half grabens resulting from a right step in the El Pilar fault trace across the Gulf of Paria to at least two main centers of displacement, the Central and Southern ranges (Figure 8). Though much of the motion along the El Pilar in Venezuela was transferred to the south, forming these pull-apart structures, some may have been transferred to the NCFZ across the Araya-Paria peninsula.

DISCUSSION

E-W and ESE-WNW faults with large normal displacements, active since but not before the Late Miocene, occur in the Gulf of Paria/Caroni Basins, in the Southern Basin, along the flanks of the Northern Range, and throughout the allochthonous Tobago terrane north of Trinidad and Araya-Paria (Figures 1, and 2) [Wilson, 1968; Case and Holcombe, 1980; Robertson and Burke, 1989; Payne, 1991]. The strike of these faults and the rapid subsidence associated with them (see also NCFZ, above) are inconsistent with models of continued right-lateral oblique collision [Speed, 1985; Speed et al., 1991]. This implies that the transpression that formed the D2/3 structures in the Northern Range, developed the fold and thrust belt in central and southern Trinidad and the Venezuelan Serranía did not continue after the Late Miocene.

Payne [1991] showed that faults striking at greater than 080° are presently transtensional, for example, Los Bajos and El Soldado faults, and North Coast Fault Zone, and only structures striking at ~070° or less are compressional, for example segments of Central and Southern Range faults (Figure 1). This implies that the strike-slip plane in Trinidad since the Late Miocene was ~080°. This orientation is slightly oblique to the 090° trend of the northern South American margin, perhaps because of the proximity to the transition between the 090° SCPBZ transform boundary and the 000° Lesser Antilles subduction zone.

TECTONIC EVOLUTION

Figure 8 summarizes our model of the Cenozoic tectonic development of the Trinidad region. This model differs from previous models, first, by our invalidation of arguments for Cretaceous orogenesis, which are (1) the observation that the Maastrichtian Galera Formation is involved in all phases of deformation affecting the Northern Range; (2) the recognition of Barr's [1963] supposed mid-Cretaceous angular unconformity between the Toco and Galera Formations as a cataclasitic fault zone (Toco-Grande Riviere Fault Zone); (3) the identification of Kugler's [1953] Cretaceous wildflysch in the Central Range as occurrences of Neogene tectonic and/or diapiric melange [Algar, 1993]; and (4) the reinterpretation of the Paleocene St. Joseph's boulder bed as a submarine erosion product [Algar, 1993], rather than as a synorogenic deposit [Kugler, 1953]. Second, most authors [e.g., Speed, 1985; Robertson and Burke, 1989] have employed either continued right-oblique convergence or E-W striking RLS to generate the structures in the eastern SCPBZ. However, the structural styles of the Northern Range and environs cannot be explained by a simple, ongoing mode of deformation. They require, instead, that the deformation in the region of the eastern SCPBZ has changed from transtension (Early to Middle Miocene) to right-lateral strike-slip at 080° (Late Miocene to Recent). We perceive the general model proposed by Speed [1985], in which oblique collision is accommodated by strain partitioning on ENE thrusts and E-W right-lateral faults, to be essentially correct for the Early to Middle Miocene only. Likewise, the simple-shear model of Robertson and Burke [1989] best suits the Late Miocene to Recent phase of development and is consistent with regional seismicity, including the continued overthrusting of the northeast striking Proto-Caribbean slab [Perez and Aggarwal, 1981; Russo and Speed, 1992], though the azimuth of net relative motion probably strikes around 080°.

CONCLUSIONS

Trinidad's stratigraphy is split into three groups, the Northern Range, Sans Souci, and Trinidad Groups. The Sans Souci Group, an allochthon in the northeastern Northern Range, probably derives from the Caribbean accretionary prism. The Northern Range and Trinidad Groups were both deposited on the northern South American passive margin from the Jurassic through the Eocene. The Northern Range Group may have been deposited up to 700 km to the WNW of its present location but probably was between 200 and 400 km WNW. There is no evidence for Cretaceous deformation in either the Trinidad or Northern Range Groups. The first deformation folded the Northern Range sediments into
northward vergent nappes between the Middle Eocene and Late Oligocene. The nappes formed either by thrusting within an accretionary prism (formed by the underthrusting of the Proto-Caribbean crust beneath northern South America), or by gravity sliding (induced from the oversteepening of the margin by the encroachment of the Caribbean Plate from the west, and/or the convergence of North and South America). Passage of the flexural forebulge ahead of the Caribbean Plate and/or convergence of North and South America may have been responsible for the uplift of Trinidad Group in the latest Eocene and earliest Oligocene. Middle to Late Oligocene foredeep subsidence heralded the Early to Middle Miocene transpression between the Caribbean Plate and northern South America in the region of Trinidad. The Northern Range was incorporated within the Caribbean accretionary prism, and then thrust to the southeast, creating a foreland fold and thrust belt in central and southern Trinidad. From the Late Miocene the azimuth of relative plate motion changed, and deformation in the right-lateral SCBPZ is typified by transpression and transpression at structures trending greater than, or less than, 080°, respectively. The transtension on the NCFZ allowed isostatic exhumation and further cooling as indicated by 11 Ma zircon fission track ages [Algar, 1993], and rapid subsidence and burial of the previously exposed Togo shelf (northern normal detachment off of Northern Range?). This strictly Cenozoic deformation history accords with, and helps to verify, kinematic models of the southern Caribbean [Pindell et al., 1988]. In relation to the rest of the SCBPZ the Northern Range of Trinidad must sit to the south of a still poorly defined interthrust boundary between Caribbean metamorphic rocks and Proto-Caribbean/South American marginal sediments.

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