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Tectonic reconstruction of the northern Andean blocks: Oblique convergence and rotations derived from the kinematics of the Piedras–Girardot area, Colombia

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Abstract

A detailed kinematic study in the Piedras-Girardot area reveals that approximately 32 km of ENE-WSW oblique convergence is accommodated within a northeast-trending transpressional shear zone with a shear strain of 0.8 and a convergence factor of 2. Early Campanian deformation is marked by the incipient propagation of northeast-trending faults that uplifted gentle domes where the accumulation of sandy units did not take place. Maastrichtian unroofing of a metamorphic terrane to the west is documented by a conglomerate that was deformed shortly after deposition developing a conspicuous intragranular fabric of microscopic veins that accommodates less than 5% extension. This extensional fabric, distortion of fossil molds, and a moderate cleavage accommodating less than 5% contraction, developed concurrently, but before large-scale faulting and folding. Paleogene folding and southwestward thrust sheet propagation are recorded by syntectonic strata. Neogene deformation took place only in the western flank of this foldbelt. The amount, direction, and timing of deformation documented here contradict current tectonic models for the Cordillera Oriental and demand a new tectonic framework to approach the study of the structure of the northern Andes. Thus, an alternative model was constructed by defining three continental blocks: the Maracaibo, Cordillera Central, and Cordillera Oriental blocks. Oblique deformation imposed by the relative eastward and northeastward motion of the Caribbean Plate was modeled as rigid-body rotation and translation for rigid blocks (derived from published paleomagnetic and kinematic data), and as internal distortion and dilation for weak blocks (derived from the Piedras-Girardot area). This model explains not only coeval dextral and sinistral transpression and transtension, but also large clockwise rotation documented by paleomagnetic studies in the Caribbean-northern Andean region. © 2005 Elsevier B.V. All rights reserved.

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1. Introduction

Most agree that the latest Mesozoic and Cenozoic evolution of the northern Andes was dominated by the relative northeastward, and then eastward advance of the buoyant Caribbean plate with respect to stable South America (Case et al., 1971; Malfait and Dinkelman, 1972; Pindell and Dewey, 1982; Burke et al., 1984; McCourt et al., 1984; Laubscher, 1987; Avé Lallemant, 1997; Villamil and Pindell, 1998; James, 2000). The relative eastward advance of this buoyant plate favored the development of transcurrent plate boundaries to its north (Rosencrantz et al., 1988) and south (Kafka and Weidner, 1981; Pennington, 1981). Unlike the sharp northern boundary, however, the southern Caribbean plate boundary is a collection of continental fragments that resisted the advance of the buoyant Caribbean Plate, and led to the progressive dextral transpressional distortion, further dismembering, rigid-body translation, and clockwise rotation of the continental fragments that make up the northwestern Andes.

This paper documents the structural style and timing of deformation in the Piedras-Girardot area, a small portion of the northern Andes that is characterized by dextral transpression (Montes, 2001; Montes et al., 2003). The kinematics and timing of deformation derived from the observations made in this foldbelt directly contradict current tectonic models for this part of the northern Andes. Consequently, a new regional tectonic model, that satisfies kinematic compatibility criteria, was developed to explain some of the puzzling problems of this margin, such as the relationship between the regional sinistral and dextral strike-slip faults. This speculative model integrates previous kinematic, paleomagnetic, and paleogeographic observations and hypotheses from the southern Caribbean plate margin in northern Venezuela and the Mérida Andes with the tectonics of the Cordilleras Oriental and Central in Colombia, two traditionally unconnected topics in the geologic literature. The Piedras-Girardot area afforded an excellent opportunity to establish an anchor point to study the interaction between the Caribbean Plate and the northern Andes because this is the only place where the Cordilleras Central and Oriental overlap, its moderate size, adequate exposure, and availability of oil industry subsurface data.

Detailed geologic mapping and collection of fabric and stratigraphic data in the field constitute the core of this study. Field studies concentrated on collecting structural data and mapping lithologic contacts and stratigraphic units at 1:25,000 scale to map changes in thickness and stratigraphic pinchouts. Mesoscopic fabric elements such as cleavage, veins, systematic joints, folds, fault surfaces, and slickenlines were measured in the field, also recording lithology, surface morphology, and a visual estimate of spacing. Shattered pebbles in conglomerate and deformed ammonite molds in black shale were employed to obtain the orientation of strain axes and their geometric relationship to more widespread fabric elements such as cleavage and veins. With the exception of the Ibagué and Cambao faults, and the Guaduas and Gualanday synclines, the nomenclature of faults and folds for the Piedras-Girardot area presented here is new.

2. Regional setting

Regional kinematic reconstructions indicate that the Caribbean Plate is in many respects an abnormal oceanic plate that apparently resists subduction due to its origin as a thick and buoyant oceanic plateau in the Pacific Plate (Malfait and Dinkelman, 1972; Pindell and Barrett, 1990; Montgomery et al., 1994; Kerr et al., 1998). Seismic refraction, reflection, and gravity studies (Edgar et al., 1971; Zeil, 1979; Bowland and Rosencrantz, 1988; Westbrook, 1990) reveal that the Caribbean oceanic crust is abnormally thick (12 to 15 km) and 1-2 km shallower than predicted by its minimum age (Early Cretaceous). This abnormally thick, shallow plate also shows signs of internal deformation that may have resulted from its relative eastward insertion through a bottleneck (Fig. 1) between the Los Muertos and the southern Caribbean deformed belts (Burke et al., 1978).

The relative eastward drift of the buoyant Caribbean Plate with respect to the American plates (Case et al., 1971; Ladd, 1976) necessarily imposes large strike-slip components along the southern and northern Caribbean plate margins. The Cayman trough (Fig. 1), along the sharp northern Caribbean plate boundary contains evidence for more than 1100 km of sinistral motion since Eocene times (Rosencrantz et



Fig. 1. General tectonic map of the Caribbean region; black indicates elevations higher than 500 m above sea level. AA: Antilles arc; CC: Cordillera Central; CO: Cordillera Oriental; CT: Cayman trough; ChB: Chortis block; FB: Falcón basin; LMDB: Los Muertos deformed belt; MB: Maracaibo basin; OD: Orinoco delta; PA: Panamá arc; SCDB: Southern Caribbean deformed belt; SM: Sierra Nevada de Santa Marta (modified from: Burke et al., 1978, 1984).

al., 1988). Similarly, the eastern segment of the southern Caribbean plate margin along northern Venezuela contains paleontologic evidence (Diaz de Gamero, 1996) that documents more than 1000 km of east-west, dextral strike-slip motions, in agreement with younging metamorphism to the east (Pindell, 1993). In contrast with these sharp strike-slip boundaries, the southwestern segment of the southern Caribbean plate boundary in Colombia is characterized by a broad and diffuse zone of oblique deformation (Kafka and Weidner, 1981; Pennington, 1981; Audemard, 2001) with the Cordillera Oriental forming its eastern border (Mann et al., 1990). Dextral strike-slip is indeed recorded along faults in the Cordillera Central, and Mérida Andes (Campbell, 1968; Barrero et al., 1969; Feininger, 1970; Schubert and Sifontes, 1970; Barrero and Vesga, 1976; Schubert, 1981; Schubert, 1983; Pindell et al., 1998), but has been deemed insignificant in the Cordillera Oriental or Magdalena Valley in regional two-dimensional reconstructions that have chosen to ignore this component of deformation (Colletta et al., 1990; Dengo and Covey, 1993; Roeder and Chamberlain, 1995). The Piedras-Girardot area, however, contains evidence of ENE tectonic transport relative to

stable South America, which is oblique to the general northeast structural grain of the northern Andes (Fig. 2), and indicates that strike-slip is significant and must be accounted for in regional reconstructions. This paper documents this direction of tectonic transport, discusses the implications of these findings, and frames the results in a new regional tectonic model that satisfies kinematic compatibility criteria.

3. Piedras-Girardot area

The rugged hills of the Piedras–Girardot area interrupt the otherwise flat and wide Magdalena Valley in central Colombia. Topographic relief in this area locally exceeds 500 m with maximum elevation reaching some 900 m above sea level. This geomorphic province constitutes the only barrier encountered by the Magdalena River in its northward journey to the Caribbean Sea. The study area has natural geographic boundaries to the east, in the western foothills of the Cordillera Oriental where elevation exceeds 1000 m, and to the northwest, along the steep topographic front of the Cordillera Central. Quaternary volcaniclastic deposits cover the



Fig. 2. Tectonic map of the northern Andes of Colombia and Venezuela. BSF: Bucaramanga–Santa Marta fault system; CF: Cambao fault; CiF: Cimitarra fault; EF: El Pilar fault; MF: Morón fault; OF: Oca fault; PF: Palestina fault; RF: Romeral fault system; SF: Salinas fault; VF: Valera fault (modified from: Feininger, 1970; Schubert and Sifontes, 1970; Tschanz et al., 1974; Skerlec and Hargraves, 1980; Schamel, 1991).

southern portion of this area. The structure of this geologic province was previously explained as the result of fold interference patterns (Tellez and Navas, 1962) or gravity gliding tectonics (Kammer and Mojica, 1995).

Four regional-scale structures of the northern Andes terminate in the Piedras–Girardot area: the Ibagué, Cambao and Alto del Trigo faults and the Guaduas syncline (Fig. 3). This relatively small area (approximately 500 km²) exhibits a wide variety of structures, deformation styles, and trends—a hint of its complex structural evolution. It exhibits an array of dextral strike-slip faults, northwest- and southeast-verging thrust faults, a positive doubly vergent structure, northeast-trending tight folds, and north-dipping normal faults. Structural trends swing from east–west to north–south in a sigmoidal sinistral stepover with faults verging outwardly in opposite directions, and diverging southward from the southern termination of the Guaduas syncline (Fig. 4). Because of the dramatic changes in structural trends aforementioned, the tectonic transport direction for structures of the Piedras–Girardot area cannot simply be assumed to be perpendicular to structural trends.



Fig. 3. Tectonic map of part of the northern Andes (modified after Schamel, 1991). Stippled patterns indicate elevations greater than 500 m above sea level. Notice the narrowing of the Magdalena Valley in the Piedras–Girardot area, and termination of major structures of the Cordilleras Central and Oriental. AF: Ataco fault; AtF: Alto del Trigo fault; CaF: Calarma fault; CF: Cambao fault; GS: Guaduas syncline; HF: Honda fault; IF: Ibagué fault; MF: Magdalena fault; PF: Palestina fault; RFZ: Romeral fault zone.

A kinematic analysis (Montes, 2001; Montes et al., 2003) indicates that this foldbelt is a dextral transpressional system where approximately 52% ENE shortening (about 32 km) is accommodated, in agreement with the minimum displacement of the Ibagué fault, and other estimates of shortening in the western margin of the Cordillera Oriental (between 16 and 30 km, Namson et al., 1994). This direction of tectonic transport was derived from three independent sources: (1) asymmetry of syntectonic strata; (2) a map-scale stratigraphic piercing point; and (3) a three-dimensionally validated palinspastic reconstruction. The first two criteria will be discussed in this paper, while the last one (Montes et al., 2003) is only briefly summarized below.

The palinspastic reconstruction was constrained by a map-scale piercing point that resulted from the discontinuous deposition of sandstone units during Campanian times. Such marker recorded an 8-km right-lateral offset across a fault, and was instrumental in determining the structural style dominant in this area. This and other structures were projected to depth in a suite of eight cross-sections using Geosec 2D[®] (two of which are shown in Fig. 5A). Each thrust sheet on each cross-section was then unfolded to obtain a map view of its undeformed shape and extent. The resulting thrust sheets were transported along the strike to achieve a geometric fit, like the pieces of a puzzle, in agreement with the observation of non-plane strain and the stratigraphic piercing point. The results yield a quantitative palinspastic restoration (Fig. 5B) indicating that this foldbelt accommodates contraction along north- and northwest-trending segments of the Cotomal and Camaito faults, dextral strike-slip along their northeast-trending segments, and extension along the north-trending Luní fault. In total, approximately 32 km of ENE-WSW contraction is recorded in this foldbelt, 17 km is accommodated to the WSW in the Cambao fault, approximately 7 km in the Camaito fault, and approximately 8 km to the ENE in the Cotomal fault. The Cambao fault represents the master fault in this foldbelt transporting the entire Mesozoic sequence and basement along a winding ramp-flatramp geometry. The northwest-verging Camaito thrust sheet obliquely transported the entire Cretaceous sequence and a portion of basement to the southwest over an irregularly shaped ramp-flat-ramp surface of the Cambao fault. The emplacement of the Camaito thrust sheet was accommodated by two elements: tectonic wedging with the southeast-verging Cotomal fault as a roof thrust that has significant dextral offset; and development of a positive flower structure between the Camaito and Santuario faults at the tip of the wedge (Fig. 5A).

The approximately 32 km of ENE–WSW contraction agrees with the minimum displacement of the Ibagué fault. This fault traverses the Cordillera Central into the study area, where the northernmost exposures of the Ibagué batholith granodiorite, and the north-trending topographic front of the Cordillera Central (marked by the 1000-m contour interval in Fig. 3) are separated approximately 30



Fig. 4. Simplified geologic map of the Piedras–Girardot area (Montes, 2001). White circles indicate location of paleontologic material. Black circles 1 to 4: Stratigraphic pinchout of the sandy member of the Nivel Intermedio; 5: Growth strata in Paleocene unit; 6: Split fold axis; 7: Folded growth strata; 8: Post-Miocene fault; 9: Pre-Miocene fold; 10–11: Quaternary activity (Base map from Raasveldt, 1956; Tellez and Navas, 1962).



Fig. 5. Structure of the Piedras–Girardot area (Montes et al., 2003). (A) Cross-sections with corresponding failed area-restoration attempts; see Fig. 4 for location. (B) Simplified palinspastic reconstruction showing direction of tectonic transport, shear strain, and convergence values for the Piedras–Girardot area.

km. Because the Ibagué batholith granodiorite is not exposed further east beyond the Piedras-Girardot area, 30 km may be considered a firstorder approximation to the minimum horizontal displacement of the Ibagué fault. The Ibagué fault separates provinces with markedly different structural styles. The northern block of the Ibagué fault has apparently behaved as a rigid block because it contains undeformed strata in a region where rocks the same age are ubiquitously folded and faulted. South of the Ibagué fault, in contrast, the same crystalline rocks are thrust upon folded and faulted Mesozoic and Cenozoic strata (Butler and Schamel, 1988; Schamel, 1991). The Ibagué fault also marks a significant change along the eastern margin of the Magdalena Valley, as large north-trending, westverging thrust faults mark the topographic front of the Cordillera Oriental north of the Ibagué fault. South of it, the doubly-vergent structure of the Piedras-Girardot area breaks this trend, and the topographic and deformation front of the Cordillera Oriental recedes eastward several tens of kilometers (Fig. 3).

The at least 30 km of ENE dextral, rigid-body translation of the northern block of the Ibagué fault was accommodated in different ways depending on the relative position of cover rocks with respect to this rigid block. The segment of the western flank of the Cordillera Oriental immediately north of the Piedras-Girardot area is directly in front of this ENE advancing block, and apparently accommodates this contraction along the major faults of Bituima and Cambao, and folding in the Guaduas syncline. The amount of east-west shortening calculated for these structures was between 16 and 26 km (southernmost two sections of Namson et al., 1994). Thus, the 32 km of ENE-WSW contraction independently calculated for the Piedras-Girardot area is compatible with the amount of contraction calculated for the western flank of the Cordillera Oriental, and the minimum displacement of the Ibagué fault (about 30 km). Northtrending folds and faults along the western flank of the Cordillera Oriental, and a diverging transpressional foldbelt in the Piedras-Girardot area are the response to the ENE advance of the northern block of the Ibagué fault. Both systems record similar amounts of contraction, and both are compatible with the amount of contraction imposed by the rigid indenter.

3.1. General stratigraphy of the Piedras–Girardot area

Stratigraphic relationships assembled during geologic mapping were analyzed to deduce the temporal and spatial distribution of deformation. First, basic stratigraphic relationships helped decipher maximum and minimum age of deformation of specific structures. For instance, the age of the oldest unit fossilizing a given fault, or not affected by folding, brackets the time of deformation to sometime between the accumulation of oldest undeformed strata, and the voungest deformed strata. Second, pronounced thickness changes and depositional pinchouts record syntectonic accumulation directly dating the deformation age and duration around the locus of accumulation. These two criteria are fundamentally different: one provides minimum and/or maximum age of deformation along specific structures, while the other dates the duration of deformation without a direct link to a given structure. Third, the composition of coarsegrained syntectonic deposits keeps a record of the composition of the source, providing supporting evidence to interpret other stratigraphic features that record local or regional deformation. Fourth, growth folds contain information about both the time, and locus of deformation, and if accurate absolute ages are available within the growth strata, rates of faulting and folding can be derived (Suppe et al., 1992). In addition, if fold axial patterns are preserved, the direction of tectonic transport can be deduced.

A major caveat when trying to decipher absolute timing of Cenozoic deformation in the Piedras-Girardot area is the exceedingly complex chronostratigraphic framework of the Tertiary sedimentary sequence (Van Houten and Travis, 1968; Wellman, 1970; Van der Wiel and Van den Bergh, 1992; Van der Wiel et al., 1992), and the absence of suitable material for age determinations. In contrast, abundant fossils yield precise ages in most of the Cretaceous sedimentary sequence (Fig. 6), and allowed accurate determinations for timing of late Mesozoic deformation. The main contribution of the stratigraphic observations consists of mapping previously unrecognized stratigraphic pinchouts (Fig. 4). A reference Cretaceous stratigraphic column was established along La Tabla Ridge, where good exposures and simple structure allowed better observation of the

Period	Age	Column	Maximum Thickness	Stratigraphic unit name
Quaternary		DDDDDDDDDDDDDDDDDDDDDDDDDDDDDDDDDDDDDDD	Ť	Alluvial deposits and colluvium
		a = p	ess	Ibagué, and Espinal fans
Tertiary	Miocene			Honda Formation
	Oligocene		thi	La Cira Formation
		°°°°°°°°		🚊 👌 Doima Formation
			Tiat	Potrerillo Formation
	Late Eocene	0 0 0 0	Van	Chicoral Formation
	Paleocene		<u> </u>	Guaduas Formation
	Maastrichtian		~ 100 m	La Tabla Formation, conglomerate member
			0 - 100 m	La Tabla Formation, sandy member
	Campanian	€ 11	~ 200 m	Nivel de lutitas y arenas, sandy member.
			~ 200 m	Nivel de lutitas y arenas, silty member.
			50 m	Lidita Superior Formation
iceous			0 - 220 m	$\frac{9}{2}$ Nivel Intermedio, sandy member (El Cobre Formation)
reta	Santonian	e 10	~ 100 m	Nivel Intermedio, silty
	Conjugion		45 m	Lidita Inferior Formation
	Turonian	e 5,6 3,4,7,8 2	3,9 ~ 300 m	(Loma Gorda Formation E La Frontera Formation?)
	Cenomanian	@_]	~100 m	(Lower Hondita Formation?).
	Albian	s de la companya de l	~100 m	Tetuán limestone.
		sod	~100 m	Caballos Formation
		tex	~150 m	El Ocal Formation
	Aptian	-No	~100 m	Alpujarra Formation
Tricacio			~50 m	Yaví Formation
Jurassic				Ibagué Batholith.

Fig. 6. Stratigraphic column of the PGFB. Compiled from early studies (Bürgl and Dumit, 1954; De Porta, 1965), and modern regional stratigraphic work (Etayo Serna, 1994; Florez and Carrillo, 1994; Gómez and Pedraza, 1994) modified only to include measured thicknesses, and stratigraphic pinchouts encountered while mapping. Numbers in column indicate the location of paleontologic material in the geologic map (Fig. 3) and column (Etayo Serna, pers. comm.). 1: *Wrightoceras ralphimlayi* (Etayo Serna); 2: *Coilopoceras* sp. ind.; 3: *Peroniceras correai* (Etayo Serna); 4: *Subprionotropis columbianus* (Basse); 5: *Prionocycloceras guayabanum* (Steinmann); 6: *Niceforoceras boyacaense* (Etayo Serna), *Peroniceras correai* (Etayo Serna); 7: *Barroisiceras loboi* (Etayo Serna), *Barroisiceras subtuberculatum* (Gerhardt), *Peroniceras guerrai* (Etayo Serna); 8: *Niceforoceras boyacaense* (Etayo Serna); 9: *Protexanites cucaitaense* (Etayo Serna); 10: *Reginaites* sp. aff. leei (Reside); 11: *Exiteloceras* cf. jenneyi (Whitfield).

sequence and accurate estimation of stratigraphic thicknesses. Stratigraphic thicknesses were calculated by correcting outcrop width with dips measured on the map.

Two very distinctive stratigraphic packages are exposed in the Piedras–Girardot area: an Upper Cretaceous marine carbonate and siliciclastic sequence, and a Tertiary nonmarine, coarse grained, and mostly red siliciclastic sequence. These sequences rest on a Triassic–Jurassic volcaniclastic, and plutonic basement. The reader is referred to published works for Paleozoic (Forero, 1990; Restrepo-Pace et al., 1997); Triassic–Jurassic (Cediel, 1969; Cediel et al., 1981; Bayona et al., 1994); Cretaceous (Bürgl and Dumit, 1954; De Porta, 1965; Etayo Serna, 1994; Villamil, 1998; Villamil et al., 1999); and Tertiary stratigraphy (Van Houten and Travis, 1968; Wellman, 1970; Anderson, 1972; Caicedo and Roncancio, 1994).

3.1.1. Syntectonic sedimentation, Late Cretaceous

Stratigraphic pinchouts occur throughout the study area in Upper Cretaceous strata across and within fault blocks. Two thick, cliff-forming, sandstone units (Fig. 6) are present south of Piedras: the sandy member of the Nivel de Lutitas y Arenas, (originally reported by De Porta, 1965), and the sandy member of the Nivel Intermedio. These Campanian units are absent to the south near Girardot (Bürgl and Dumit, 1954; Cortés, 1994) and disappear or thin most notably across the Camaito and Cotomal faults (Fig. 4). These two sandy units up to 200 m thick consist of similar monotonous fine-grained, lithic sandstone commonly occurring in planar beds up to 50 cm thick, with abundant shelly material and calcareous cement; (for further details on the stratigraphy of these units, the reader is referred to: Bürgl and Dumit, 1954; De Porta, 1965; Guerrero et al., 2000). To the north, a third sandy member, present locally along La Tabla Ridge, is the sandy member of the La Tabla Formation of Maastrichtian age. Unlike the other two sandy members, this unit is a clean, medium-grained, well-sorted, permeable sandstone that occurs in massive beds up to 10 m thick immediately below the conglomerate beds of the La Tabla Formation.

The sandy member of the Nivel Intermedio in the Oliní Group (Fig. 6), also known as El Cobre Formation (Guerrero et al., 2000), disappears across the northeast-trending, northern segments of the Camaito and Cotomal faults, and along the El Guaco anticline (labeled 1 to 4 in Fig. 4). This unit thins to zero in the northernmost segment of the hanging wall of the Cotomal fault and in the footwall of the same fault approximately 8 km to the southwest (labeled 2 and 3 in Fig. 4). Published stratigraphic sections indicate that this unit is also missing in the core of the El Guaco anticline (labeled 1 in Fig. 4) immediately north of Girardot (Bürgl and Dumit, 1954; Cortés, 1994). Despite these stratigraphic truncations and hiatuses, the unit immediately above (Lidita Superior Formation) is present throughout this foldbelt, resting conformably on units below and maintaining a constant thickness. The sandy member of the Nivel de Lutitas y Arenas also thins from approximately 200 m along La Tabla Ridge to zero along the southern segment of the El Guaco anticline near Girardot. Changes in thickness across the Cotomal and Camaito faults, however, cannot be evaluated because this unit has been eroded from these hanging walls.

3.1.2. Syntectonic sedimentation, Paleogene

The massive mudstone and claystone of the Paleocene Guaduas Formation exhibits remarkable thickness changes within the Gualanday syncline. This unit is seldom exposed, thus most observations and interpretations are based on seismic reflection data (Fig. 7), and analysis of regional map patterns (Fig. 4).

In cross-section, the Gualanday syncline is an asymmetric, almost box fold with steeply dipping limbs, and a very wide, nearly horizontal hinge zone. Gualanday Group strata within this wide hinge zone dip gently to the west, while Cretaceous strata immediately below dip gently to the east (Fig. 7). The geometry of this syncline is directly related to the pronounced thickness changes that take place within the intervening Guaduas Formation. Both the geologic map (Fig. 4) and a seismic reflection profile (Fig. 7) show that the Guaduas Formation significantly thickens eastward, from the axis of the Doima anticline toward the axis of the Gualanday syncline where it reaches its maximum thickness.

The Guaduas Formation apparently rests conformably on Upper Cretaceous strata. In contrast, marked angularity exists between the bold reflectors at the base of the Gualanday Group and faint reflectors at the top of the Guaduas Formation (Fig. 7). This angularity, however, disappears in the west-dipping, eastern limb of the Gualanday syncline where reflectors of the Guaduas Formation and the Gualanday Group become parallel. The fold axial trace of the Gualanday syncline in the time profile is complex because it diverges downward from the top of the Guaduas Formation (Fig. 7). A similar divergent pattern is evident in the geologic map where the wide, south-plunging, northernmost part of the Gualanday syncline consists of three relatively uniform dip slopes: a NNE trending, southeastdipping western limb, and a NNE- and NNW trending eastern limb that define a converging axial trace (labeled 6 in Fig. 4).



Fig. 7. Seismic section across the Gualanday syncline. See Fig. 4 for location. Note thickening of folded Paleogene strata towards the axis of the Gualanday syncline, complex fold axial trace, and angularity between reflectors at the base of the Gualanday Group west of the eastern limb of the fold. Part of line GT-90-1625. 1: Axial surfaces; 2: Gualanday Group; 3: Growth strata (Guaduas Fm.); 4) Pre-growth strata (Upper Cretaceous); 5) Pre-Cretaceous strata; 6) Cambao fault; 7) Hanging wall cutoff.

Thickness variations define a complex fold axial trace morphology that splits, with the axial traces delineating a roughly triangular zone that widens downward (Fig. 7), and northward (labeled 6 in Fig. 4). The apex of this triangular zone is at the base of the Gualanday Group, marking the end of growth strata. Thickness changes and the morphology of the fold axial trace of the Gualanday syncline (both in map and cross-section view) reveal a time of syntectonic sedimentation, fold growth and propagation (Paleocene Guaduas Formation), followed by a time of conglomerate accumulation and folding (Gualanday Group). Whether or not this later folding took place as the conglomerate of the Gualanday Group was accumulating cannot be tested because erosion has removed higher stratigraphic levels in the Piedras-Girardot area.

Thickening of the growth strata to the east, the location of the hanging wall cutoff (Fig. 7) east of the axis of the Gualanday syncline, and preservation of the eastern, but not of the western growth axial traces, indicate that the Cambao thrust sheet was moving westward to southwestward relative to stable South America when growth strata accumulated.

3.1.3. Syntectonic sedimentation, Neogene

The distinctive volcaniclastic sandstone of the Late Oligocene to Miocene Honda Formation provides additional constraints to construct a deformation timeline. This unit rests unconformably and overlaps upper Cretaceous folded strata near Girardot (Raasveldt, 1956). Near Piedras, in contrast, the same unit is folded, and in faulted contact with Cretaceous strata along the Cambao fault (labeled 8 and 9 in Fig. 4). These relationships simply indicate that while deformation took place along the Cambao fault after the Miocene, it did not take place to the south, near Girardot. In addition, this demonstrates that folding in the southern part of the Piedras–Girardot area is pre-Miocene.

The Quaternary Ibagué fan blankets the western half of the study area with Pleistocene volcaniclastic material 50 to 300 m thick (Vergara, 1989) derived from the axis of the Cordillera Central (Thouret and Laforge, 1994; Thouret et al., 1995). This fan, however, is tilted, uplifted, and offset near Piedras, and along the trace of the Ibagué fault in aligned en échelon domes with axes oriented obliquely to the main trace of the fault (labeled 10 and 11 in Fig. 4), constraining the latest activity of the Ibagué fault to Holocene times (Vergara, 1989). All other structures on the western side of the Piedras–Girardot area (Cambao and Camaito faults, Gualanday syncline and Doima anticline) are fossilized by these young deposits.

3.2. Paleogeographic interpretation

Stratigraphic pinchouts in fine-grained Cretaceous strata are interpreted to record early Campanian mild uplift (Fig. 8a) along the northern, northeast-trending, segments of the Cotomal and Camaito faults, and along the easternmost, also northeast-trending, structure of this foldbelt (El Guaco anticline). Late Campanian rejuvenation may have occurred because the sandy member of the Nivel de Lutitas y Arenas (Fig. 6) is also missing in the El Guaco anticline.

Altogether, these coarse-grained marine clastic units represent early and late Campanian (~84 and ~74 Ma) gentle deformation near the northeasttrending, northern segments of the Cotomal and Camaito faults. Although gentle uplift may be responsible for these pinchouts and truncations, no major unconformities were developed at this time, as overlying units rest apparently conformably with syntectonic units below. Since bedding geometry remained mostly parallel, it is unlikely that the Cotomal and Camaito faults propagated to the surface at this time. These stratigraphic pinchouts could also be interpreted as the result of eustatic sea level changes (Guerrero et al., 2000). Nonetheless, Late Cretaceous arrival of the leading edge of the Caribbean Plate (Pindell and Dewey, 1982; Lugo and Mann, 1995) at this latitude provides the initial driving force for deformation in this part of the Andes. Even though a eustatic signature must be overprinted, the tectonic component is probably dominant beginning in Late Cretaceous times. Later, during Maastrichtian times, a nearly continuous blanket of conglomerate and sand conformably covered underlying units, recording a time of local quiescence, and regional unroofing to the west (Fig. 8a). Paleogeographic analysis of the uppermost Cretaceous Cimarrona Formation further north offers a model where a series of fan deltas dominated by braided rivers prograded from west to east as they drained the eastern flank of the Cordillera Central (Gómez and Pedraza, 1994) covering a previously established,



Fig. 8. Palinspastic paleogeographic reconstruction of the ENE translation of the rigid indenter of the Cordillera Central (showing its minimum extent) along the Ibagué fault. (a) Latest Cretaceous–Early Paleogene: incipient propagation of faults. (b) Paleogene: propagation of thrust sheets, and accumulation of syntectonic molasses. (c) Miocene to recent: accumulation of the Honda Fm.

albeit gentle, relief. The Cordillera Central nearby is a good candidate as source because it contains quartzite and phyllite from the metamorphic basement, and was likely covered by a Cretaceous sedimentary veneer of black chert and mudstone (Villamil, 1999). The widespread distribution, lateral continuity, and overall uniformity of this unit in the Piedras–Girardot area indicate that latest Cretaceous uplift and deformation were minor, less than 400 m, estimated from the combined thicknesses of syntectonic units.

The Paleogene depositional setting of the Piedras-Girardot area began to be partitioned by a rising northand northeast-trending elongate plateau that prevented accumulation of sediments between the Camaito fault and El Guaco anticline (Fig. 8b). As this plateau rose, molasse sedimentation took place along its southwestern and northeastern flanks, in the Guaduas and Gualanday synclines. Growth strata in the Gualanday syncline (Fig. 7) records continuous early Paleogene uplift and movement of the Cambao thrust sheet to the west or southwest relative to stable South America. Tertiary sedimentation occurred only west of the Camaito fault, north of the Luní fault and east of the El Guaco anticline (Fig. 8b and c). This is because even in very low structural positions such as the hinge of the La Vega syncline, the very conspicuous Tertiary sedimentary sequence is absent. Even though it is possible that the Paleogene sequence had accumulated throughout, it is unlikely because syntectonic deposits clearly indicate that large-scale thrust sheet movement and surface uplift were taking place at this time between the Camaito fault and El Guaco anticline.

Changes in clast composition between the base of the Gualanday Group (Chicoral Formation) and its top (Doima Formation) have been interpreted as a result of changing from a metamorphic to a sedimentary source (Van Houten and Travis, 1968). This agrees with the paleogeographic interpretation presented here because the metamorphic rocks of the Cordillera Central were already exposed and actively contributing clasts to Maastrichtian sedimentary deposits, most notably the La Tabla Formation. Clast composition in the Gualanday Group records the shift from western to local sources because the contribution of metamorphic rock clasts decreases from the Chicoral to the Doima Formation (Van Houten and Travis, 1968). The source of the Doima Formation must be local because Paleogene uplift clearly took place in the Piedras-Girardot area at this time (Fig. 7). The Cordillera Oriental to the east can be ruled out because paleoflora collected on the crest of the Cordillera Oriental record uplift from about 500 m to about 3000 m only until Pleistocene times (Van der Hammen et al., 1973). Early uplift rates are very low (0.03-0.05 mm/year), compared to Pliocene rates (0.6-3.0 mm/year, Gregory Wodzicki, 2000). Therefore, the provenance for this clastic material must be local, with early contributions from the Cordillera Central and the rising Piedras–Girardot area providing most of the sediments for the Paleogene molasse deposits.

3.3. Microscopic and mesoscopic strain

This section presents an analysis of mesoscopic and microscopic fabric elements of the Piedras-Girardot area to evaluate the amount and directions of nonrigid finite deformation. Three elements are investigated: (1) deformed fossils; (2) cleavage; and (3) microscopic and mesoscopic veins. Even though deformed fossils faithfully record orientations of finite strain axes, they alone cannot be used to estimate the magnitude of strain because useful fossils are uncommon and restricted to a few stratigraphic horizons. Cleavage, while ubiquitous in some lithologic types, is absent in others, and can be used to estimate strain magnitude where present. Microscopic intraclastic veins record amounts and direction of extension, but they are stratigraphically restricted to a small part of the column. Together, these elements are used to construct a summary of finite strain, and the contribution of nonrigid-body deformation in the Piedras-Girardot area.

3.3.1. Deformed fossils

Deformed external molds of ammonite shells preserved in black shale in the lower Villeta Group were used as strain markers. Fossilization completely eliminated the original shell, leaving behind only the shell imprint on a bedding surface so there is no ductility contrast between the rock and the strain marker. The friability of the black shale where these molds are found prevented sample collection for lab study so only field measurements could be made. Although better preserved macro- and microfossils (ammonites, bivalves, forams, and gastropods) are present sometimes ubiquitously, either their plane of symmetry was not properly aligned with respect to bedding, or they were replaced shells that could not properly record finite strain due to a ductility contrast between the specimen and the matrix.

While determining the strain modification of the spiral angle of ammonites would have been preferred for strain measurement, difficulty in measuring angles on frail imprints in weak shales precluded this approach. Instead, because most gastropod shells have typical spiral angles of more than 80° (Tan, 1973), which approaches a circle, the ammonite molds were treated as deformed circles for which long and short axial lengths and orientations were measured. Field measurements of deformed external molds of ammonite shells are presented in Fig. 9 (insets 1–4). Each station represents a summary of measurements

made along stratigraphic intervals of a few centimeters where molds were abundant. All four stations (insets 1 to 4 in Fig. 9) are located in Villeta Group black shale, except for one specimen measured within the lowermost part of the Oliní Group; the axial ratio measured in these localities ranges between 1.0 and 2.3, with an average of 1.5. Station 4 (inset 4 in Fig. 9) contains slightly deformed molds where the difference in axial length is very small. Axial orientations were



Fig. 9. Map of mesoscopic structures in the Piedras–Girardot area. Insets 1 to 4 contain two-dimensional strain ellipses estimated from deformed ammonite impressions in black shale. Rose Diagrams of: (a) Long axes of deformed ammonite molds after dip of bedding has been removed; (b) Mesoscopic fold axes; (c) Mesoscopic fault planes; (d) Slickenlines. (e) Lower hemisphere, equal-area Kamb contour diagram of poles to cleavage; (f) Poles to cleavage after tilt of bedding has been removed.

recorded in the field, and then later rotated to remove the dip of bedding assuming horizontal fold axial lines.

The long axes of deformed molds generally parallel the trend of cleavage (Fig. 9) ranging in orientation between ENE approximately 5 km away from the Ibagué fault to NNE trends less than one km away from the Ibagué fault. Magnitude of axial ratios is similarly related to distance to faults as high ellipticity values are found less than 400 m from a fault, whereas low- to nearly undeformed molds are found more than 1000 m from a fault. The true distance to the fault surface in all cases must be less than the horizontal distance measured on the map because the stations are in all cases located in the hanging wall of a fault. These observations indicate that the magnitude of strain, in some cases relatively large, is not pervasive through the entire stratigraphic column but is restricted to highly deformed stratigraphic intervals near faults. The orientation of the long axes of deformed molds defines the changing orientation of a maximum horizontal finite strain that becomes more northerly as distance to the Ibagué fault decreases.

3.3.2. Cleavage

Cleavage is the most pervasive, uniform, and prominent outcrop-scale structure found in finegrained Cretaceous strata of the Piedras-Girardot area. Cleavage is commonly anastomosing (Powell, 1979), bedding-normal, and regularly spaced even in the hinges of map-scale and mesoscopic folds; it stands out in weathered exposures, where the intersection between wavy domains and lamination in shale defines pencil structures. Cleavage domain surfaces are commonly smooth in hand sample, and microlithons contain no detectable deformation structures. Spacing between cleavage domains is independent of distance to faults or fold axial traces, but is dependent on lithology. Siliceous mudstone, siltstone, and calcareous shale contain strong to moderate (e.g., Engelder and Marshak, 1985) non-sutured, wavy cleavage. Fine-grained sandstone commonly develops a weak planar cleavage. Coarse-grained sandstone beds lack cleavage, but have widely spaced nonsystematic fractures. Chert, also, displays no cleavage.

Microscopic examination of cleavage domains in siliceous siltstone reveals surfaces with a sutured

morphology (Engelder and Marshak, 1985), where thin films (approximately 0.1 mm) of insoluble material are concentrated indicating pressure solution (Fig. 10). Each cleavage domain consists of a few (three or four) discontinuous overlapping, and osculating surfaces that are deflected around large particles (Fig. 10). The microlithons show no evidence of dissolution, and the rock is pristine. Assuming that an extreme amount of dissolution (for instance 50%) took place along each film of insoluble material, and that each cleavage domain consists of 5 overlapping selvages, each one 0.1 mm thick, the amount of shortening accommodated by a rock with a 2-cm domain spacing is only 5%. Shortening values, however, are likely to be much less since volume loss along each film is likely to represent less extreme values, and because microscopic observations do not support extreme shortening values.

Cleavage commonly trends ENE, almost parallel to the Ibagué fault, and to fault traces in the northern third of the study area (northern segments of Cotomal, Camaito and Santuario faults). As these faults turn to the NNE in the middle third of the area mapped, cleavage trends remain unchanged, now oblique to the southern, north- and NNE trending segments of the Camaito and Santuario faults, and nearly perpendicular to the southern half of the north-trending segment of the Cotomal fault. Sampling density decreased in the southern third of the map and prevents further observations. Cleavage traces are also oblique to the traces of most folds in the study area. A Kamb equalarea stereographic plot shows a simple unimodal



Fig. 10. Photomicrograph of cleavage domains taken looking onto bedding in a siliceous mudstone. Note cleavage deflected around microfossils. Width of photograph is 18 mm.

distribution of poles to cleavage (Fig. 9e) that coincides with the observation that cleavage traces are nearly independent of changes in trend of mapscale structures such as faults and folds. These changes in the orientation of cleavage, like the long axes of deformed ammonite molds, define a systematic ENE orientation.

Removing the dip of bedding due to folding and faulting from the attitude of cleavage surfaces results in an unimodal distribution of poles to cleavage, slightly tighter and with a steeper mean pole (Fig. 9f) than the original that is still 14° from the vertical. Field observations of cleavage are almost always perpendicular, or very close to perpendicular to bedding indicate that it developed mostly before folding; if cleavage had developed after folding it would cross bedding at different angles, and removal of the dip of bedding would result in spreading out the poles. If, on the other hand, cleavage had developed entirely before folding, the resulting distribution of poles after removing the tilt of bedding would be very tight and closer to vertical, or vertical. An explanation for the failure of these poles to reach a tight, vertical attitude after removal of the tilt of bedding may be that some gentle folding had already taken place by the time cleavage started to develop by LPS (Fig. 8a). Stratigraphic observations discussed earlier indicate that gentle folds were nucleated as early as late



Fig. 11. Map of mesoscopic structures in the Piedras–Girardot area. Traces of veins were interpolated from field measurements. (a) Insets contain length-azimuth diagrams (dip of bedding removed) of intraclastic veins on 23 field photographs from 10 field stations (Table 1). (b) Rose diagram of field measurements of intraclastic veins in conglomerate of the La Tabla Formation (dip of bedding removed); (c) Lower hemisphere, equal-area Kamb contour diagram of poles to veins, and (d) poles to joints.

Campanian times. Hence, cleavage development must have occurred after initial fold growth (late Campanian), and may have continued to develop during latest Cretaceous and earliest Paleocene times. Cleavage formation, however, must have ceased before the propagation of large thrust sheets, and development of large folds, since it is passively translated by these structures.

3.3.3. Shattered pebbles and veins

The conglomeratic unit atop of the Cretaceous sequence of the Piedras-Girardot area (La Tabla Formation) contains a conspicuous intragranular fracture deformation fabric (Fig. 11). Clast-supported, quartzite-rich conglomerate mostly near the top of this unit consistently develops a systematic fabric of intragranular microveins perpendicular to bedding (Fig. 12), while matrix-supported conglomerate lacks this fabric. Microscopic analysis reveals that this fabric consists of calcite-filled microscopic veins (Fig. 13) that become visible in outcrop exposures due to differential weathering of calcite. The trend of this microscopic intragranular fabric remains roughly constant from pebble to pebble defining a systematic mesoscopic deformation fabric. Microscopic pock marks where dissolution occurred (Fig. 13) are also common at grain-to-grain contacts, although the poor framework packing in this conglomerate precludes a high density of these contacts. Microscopic examination also revealed that the calcareous matrix is twinned.

Although less systematic than cleavage, the overall trend of intraclastic and mesoscopic veins is north-



Fig. 12. Shattered pebbles of La Tabla conglomerate in a clastsupported conglomerate.



Fig. 13. Photomicrograph of La Tabla conglomerate looking onto bedding. Sutured contact between two quartzite pebbles, note the coordination between calcite-filled veins and the sutured contact. Width of photograph is 4.2 mm.

west, turning to more east–west trends towards the Ibagué fault. Locally, significant and abrupt changes in trend take place: northeast-trending mesoscopic veins were measured along both flanks of La Tabla Ridge parallel to the Camaito fault and west of the Santuario fault, which is a 90° change from veins trends to the south, north, and east (Fig. 11a, insets 4 and 5). Southeast of the Cotomal fault mesoscopic and intraclastic veins have nearly identical northwest trends at almost right angles to the axial traces of major folds and the northern, northeast-trending segment of the Cotomal fault (Fig. 11, insets 6 to 10). Intraclastic and mesoscopic veins also trend northwest although the variability is more pronounced, and change to WNW near Piedras.

Because this microscopic intragranular fabric is conspicuous in outcrop exposures, field photographs looking onto bedding were used to measure the orientation and length of intragranular veins, recording every visible vein tracelength in the photographs. Field photographs at 10 different stations were analyzed to count the number, length, and orientation of all visible veins. In total, 3463 intragranular veins were measured with a total length of 2348 cm in a total surface area of 11,201 cm² (Table 1). The percentages of matrix versus grains in three thin sections of a conglomerate from La Tabla Formation (station 5 in Table 1, and Fig. 11) were also measured, and the total area of intraclastic veins was calculated for the purpose of obtain total area change.

Overall, the average vein length is 0.73 cm, which is indicative of the average grain dimension parallel to

Station number (Fig. 11)	Number of veins measured	Total cumulative length (cm)	Area measured in photo (cm ²)	Number of veins per cm ²	Length of vein per cm ²	Average vein length
1	152	145.07	1847	0.08	0.08	0.95
2	60	55.41	326	0.18	0.17	0.92
3	322	263.87	3761	0.09	0.07	0.82
4	59	47.55	267	0.22	0.18	0.81
5	928	583.52	2643	0.35	0.22	0.63
6	278	127.02	61	4.56	2.08	0.46
7	340	290.5	532	0.64	0.55	0.85
8	656	489.49	1062	0.62	0.46	0.75
9	464	211.15	505	0.92	0.42	0.46
10	204	135.38	199	1.03	0.68	0.66
Total	3463	2348.96	11203			
Average				0.87	0.49	0.73

Table 1 Measurements of intragranular microveins in the La Tabla formation conglomerate

veins because most veins completely cross pebbles. A more meaningful number is the intensity or average length of intragranular veins per square centimeter (Wu and Pollard, 1995), and its variation across the area mapped (Fig. 11). Exposures of the La Tabla Formation conglomerate on the northwestern side of the Piedras-Girardot area have smaller length values per square centimeter (between 0.07 and 0.22), than exposures on the southeastern side (between 0.42 and 2.08). This increase in intensity is indicative of a greater amount of extension by vein development to the southeast, away from the Ibagué fault. Despite the conspicuous appearance of this fabric, the measured total area change accommodated by intraclastic veins with an average intensity of 0.22 on the northwestern side of the Piedras–Girardot area (inset 5 in Fig. 11a) is only between 1% and 2%. Extreme values (inset 6 in Fig. 11a) are not representative, and may be due to local effects along discrete fracture zones. More average values of about 0.4 (Table 1), therefore record less than 5% extension.

Previous studies in similarly deformed conglomerates indicate that the direction perpendicular to the intragranular veins or fractures is parallel to the maximum finite stretch axis of the finite strain ellipse (Tyler, 1975; Tanner, 1976; Wiltschko et al., 1982; Jerzykiewicz, 1985; Calamita and Invernizzi, 1991; Lin and Huang, 1997). Similar fractured clasts have been reported in recent deposits adjacent to active faults (Tanner, 1976; Eidelman and Reches, 1992; Lin and Huang, 1997), and may develop under little overburden (Wiltschko et al., 1982). Photomechanical and experimental studies indicate that extensional fractures in grains of stressed cemented aggregates where grains and matrix having similar elastic modulii exhibit a higher degree of preferred orientation than uncemented aggregates (Gallagher et al., 1974). These experiments indicate that microfractures tend to develop parallel to the greatest principal stress trajectory, with little influence of stress concentration at grain-to-grain contacts in cemented aggregates. Therefore, the deformation fabric in La Tabla Formation conglomerate may have developed shortly after its accumulation, with little overburden, and it may be used to deduce the orientation of the maximum finite stretch axis of the finite strain ellipse, as the direction perpendicular to the northwesttrending veins. This stretch is approximately perpendicular to the direction of minimum finite stretch independently deduced from cleavage and deformed fossils. The intraclastic microscopic veins in La Tabla Formation conglomerate record small amounts of extension (less than 5%) in this direction.

4. Kinematic development of the Piedras–Girardot area

In summary, microscopic and mesoscopic fabric elements in the Piedras–Girardot area, albeit prominent and pervasive, record less than 5% shortening in a general northwest direction, and less than 5% extension in a northeast direction. Within the limits of error, these two types of structures may represent mutually compensating mechanisms of dissolution and precipitation of soluble materials within a nearly closed system (as suggested by Hobbs et al., 1976). The orientation of all fabric elements and the magnitude of internal strain are a function of horizontal distance to the Ibagué fault: the trend of cleavage and of long axes of deformed ammonite molds become closer to the trend of this fault as the distance to it decreases. Similarly, vein intensity increases and internal strain magnitude decreases as distance to the fault increases. Mesoscopic fabric elements, like map-scale faults and folds, are oblique to the independently constrained relative direction of tectonic transport (ENE, Montes, 2001).

Late Campanian deformation fabrics like cleavage and veins started to develop very early, after the initial propagation of the northernmost, northeast-trending segments of the Camaito and Cotomal faults, and the El Guaco anticline. Although these early faults did not breach the surface, they were probably rooted along a basal detachment along the lower part of the Villeta Group (Montes, 2001). These gentle structures were later overlapped by La Tabla Formation conglomerate, which records Maastrichtian unroofing to the west. which were most likely supplied by erosion in the Cordillera Central. The early Paleogene marks a time of segmentation of a once continuous accumulation environment due to the relative west- or southwestward propagation of the Cambao fault (and probably other faults in this foldbelt), and generation of accommodation space in the Guaduas and Gualanday synclines, to the northeast and southwest of the Piedras-Girardot area respectively. Paleogene and younger deformation, although spectacularly recorded by thick, folded molasse deposits and map-scale faults and folds, is missing a mesoscopic deformation fabric. Earlier fabric elements were passively translated within propagating thrust sheets. This foldbelt has been a positive area since then, shedding clastic material into these actively deforming Paleogene depocenters. Only the northern part of the study area contains evidence for post-Miocene deformation, which is at least partially related to the latest motion along the Ibagué fault.

The observations above outlined may be framed within a kinematic concept where the ENE-trending Ibagué fault represents one of the synthetic shears in a regional northeast-trending dextral shear zone parallel to the overall trend of the Cordillera Oriental (Fig. 2). In such a zone, the orientation of fabric elements such as cleavage would initially be oriented north-south; as deformation progressed, the infinitesimal strain axes would rotate toward orientations closer to the boundary of the shear zone (Sanderson and Marchini, 1984; Tikoff and Teyssier, 1994; Krantz, 1995). Hence, at advanced stages of deformation, and more intensely near the fault, the orientation of the maximum finite strain axis would become more easterly. A northeast trend (~N40E) for the maximum horizontal finite strain axis in the Piedras-Girardot area, near the fault, is consistent with the theoretical prediction (Sanderson and Marchini, 1984; Krantz, 1995) based on the kinematics of a shear zone with a convergence factor of 2.0, and a shear strain of 0.8 (Fig. 5).

Paleostress analyses in other parts of the Cordillera Oriental (Mojica and Scheidegger, 1981, 1984; Kammer, 1999; Taboada et al., 2000) were not incorporated here because methods for determination of paleostress from naturally deformed rocks must assume a single, coaxial, strain-inducing event (Angelier, 1979; Ramsay and Lisle, 2000) without the interaction of neighboring faults (Maerten, 2000). This is not the case of the northern Andes, a region with a long history of deformation and fault reactivation extending back to Early Mesozoic rifting.

5. Speculations on northern Andean tectonics

The kinematic results outlined above markedly contrast with traditional two-dimensional kinematic models of the Cordillera Oriental. The key difference is that directions of tectonic transport have not been established outside the Piedras-Girardot area, and that additional modern structural analyses are absent in this part of the Andes. For simplicity's sake, regional models reconstructing the predeformational geometry of the Cordillera Oriental must assume plane strain and large (>300 km wide) composite thrust sheets riding along a crustal-scale, east-verging master detachment rooted beneath the Cordillera Central (Colletta et al., 1990; Dengo and Covey, 1993; Cooper et al., 1995; Roeder and Chamberlain, 1995), with subduction of the Caribbean Plate, or another oceanic plate, driving deformation (Kellogg

and Bonini, 1982; Freymueller et al., 1993; Taboada et al., 2000).

These simplified models, however, fail to produce acceptable solutions because: (1) the aforementioned difficulty to subduct buoyant Caribbean Plate crust, which more likely only bends under northwestern South America, and (2) restoration of foreland foldthrust belts also requires displacing the metamorphic assemblages and basement overlying the internal parts of the basal detachment (Oldow et al., 1990). Restoration of these large composite thrust sheets along a middle crustal detachment (Colletta et al., 1990; Dengo and Covey, 1993; Cooper et al., 1995; Roeder and Chamberlain, 1995), as suggested in some of these models, also requires displacing the metamorphic rocks and basement above the basal detachment. This operation, however, would displace the metamorphic core of the Cordillera Central beyond the edge of the continental crust (Romeral fault zone, Case and MacDonald, 1973; Etayo Serna et al., 1986) in the northern Andes, and so subdetachment mass would be missing (Fig. 14).

Such geometric contradiction could be easily explained by full deformation partition of the oblique convergence imposed by the Caribbean Plate between the Cordilleras Oriental (dip-slip) and Central (strikeslip). However, regional-scale thrust faults near the Piedras–Girardot area accommodate oblique displacements (Montes, 2001), ruling out the possibility of full deformation partitioning, and indicating that oblique convergence imposed by the relative eastward advance of the Caribbean Plate is distributed in a wide zone of deformation that at least includes the western flank of the Cordillera Oriental. The width of this zone and its gradient are unknown, but most likely comprises the domain of the Cordillera Oriental, with a larger contribution of strike slip towards



Fig. 14. Schematic section highlighting the contradiction existing in current models attempting to restore deformation of the Cordillera Oriental along a middle crustal detachment (MCD). RF: Romeral fault zone, marking the edge of continental crust.

the west, and a larger component of dip slip towards the east. The scarcity of detailed modern structural studies elsewhere prevents a more complete characterization of the structural style from being made; nonetheless, a new kinematic model based in the field observations and kinematic reconstructions of the Piedras–Girardot area (Montes, 2001), as well as other published outcrop, paleomagnetic, and tectonic data are presented below.

5.1. Dextrally transpressional kinematic model

The new kinematic model presented here is based on observations made in the Piedras–Girardot area. These observations have regional significance because structural features such as the Guaduas syncline, and the Ibagué, Alto del Trigo, and Cambao faults accommodate significant amounts of deformation with respect to the whole set of faults and folds in the western margin of the Cordillera Oriental and Magdalena Valley. The Piedras–Girardot area is therefore an anchor point that allows independent determination of tectonic transport direction, finite strain and timing of deformation (Montes, 2001). Hence, the structural style described for the Piedras– Girardot area must represent the dominant deformation style, not an isolated oddity.

Because the observations made here have regional significance, we postulate that dextral transpressional deformation played a fundamental role in the structural development of the Cordillera Oriental and Magdalena Valley. This does not mean that dip-slip along thrust faults is absent; it simply means that, for the sake of simplicity, previous studies have chosen to ignore a very significant component of deformationdextral strike-slip. If this component of deformation is accounted for, the structure of the northern Andes must be modeled as a dextrally transpressional system.

5.1.1. Assumptions and boundary conditions

Modeling the structure of the northern Andes as a dextrally transpressional margin requires a number of assumptions and simplifications. These assumptions delimit the number of variables to be considered, facilitate construction of the model, and allow broad predictions and regional comparisons to be made. A simple, broadly generalized model of the structure of the northern Andes is preferred at this time because the paucity of constraining structural data preclude the development of a more elaborated reconstruction. Simple reconstructions such as the one presented in this paper should highlight key areas for additional study, and help to establish conceptual frameworks to study the structure of the northern Andes.

First, the remarkably linear northeast-trending eastern flank of the Cordillera Oriental (Fig. 2) can be assumed to represent the eastern limit of deformation because the craton and overlying strata to the east are essentially undeformed (Cooper et al., 1995). Second, cross-sections containing subsurface information (Schamel, 1991; Namson et al., 1994), field data (Hubach, 1945; Restrepo Pace, 1989; Restrepo Pace, 1999), and geologic maps (Cediel and Cáceres, 1988) show that in a very general sense, the structural style of the Cordillera Oriental is relatively uniform. Briefly, this style is characterized by north- and northeast-trending faults and folds arranged in deformed belts on both flanks of the Cordillera verging outwardly in opposite directions from a relatively undeformed and topographically high axial zone (Scheibe, 1938). The assumption here is that this relatively uniform style reflects a common genetic process throughout the length of the Cordillera Oriental. Finally, in order to model the structural development of the northern Andes, deformation must be synthetically factored in two components: one of rigid-body translation to the ENE, oblique to the structural trends; and second, a component of homogeneous, simple shear deformation. The first component represents the rigid-body translation of large, regional scale fault systems (Fig. 2). The second component attempts to incorporate rigid-body translation and rotation below the resolution of this reconstruction; it does not attempt to take into account internal strain. Internal deformation, at least in the Piedras-Girardot area, was shown here to be minor (less than 5%) when compared with rigid-body translation.

5.1.2. Crustal blocks

A kinematic reconstruction of the northern Andean puzzle also requires defining the fragments of continental crust as well as their mechanical behavior. Because structural style reflects the mechanical response of the crust to deformation, it is used here as the primary criterion for this division. Major faults or fault systems outlined in Fig. 2 are used to define the boundaries of three major blocks in the northern Andes: (1) Cordillera Central–Middle Magdalena block, (2) Cordillera Oriental–Upper Magdalena block, and (3) Maracaibo block (Fig. 15). In this simple scheme, the craton to the east is considered stationary and rigid, while the oceanic terranes west of the Cordillera Central, and north of the Maracaibo block are added as the Caribbean deformation front advances along the northwestern margin of South America.

The Cordillera Central-Middle Magdalena and the Cordillera Oriental-Upper Magdalena were separated at the latitude of the Ibagué fault on the basis of their contrasting structural styles, the former dominated by strike-slip faults, and the latter by thrust faults. These changes in structural style likely reflect contrasting mechanical properties resulting from different tectonic histories. The Cordillera Central did not accommodate large volumes of Cretaceous strata, and it may have been a positive area since early Mesozoic times (Barrero et al., 1969; Villamil, 1999), making it a relatively rigid crustal block (Fig. 15). The relative rigidity of the Cordillera Central-Middle Magdalena block is expressed by undeformed, westward-onlapping Mesozoic and Cenozoic strata along the eastern flank of the Cordillera Central north of the Ibagué fault (Raasveldt, 1956; Raasveldt and Carvajal, 1957a; Feininger et al., 1970; Barrero and Vesga, 1976; Schamel, 1991). The relative weakness of the Cordillera Central south of the Ibagué fault is, in turn, indicated by pervasive deformation of Mesozoic and Cenozoic strata along its eastern flank (Raasveldt and Carvajal, 1957b; Schamel, 1991; Amézquita and Montes, 1994).

The rigidity of the Cordillera Central north of the Ibagué fault may also be the cause of the radically different outcrop patterns between the Ibagué and Antioquia batholiths; while the former shows an elongated outcrop pattern (Fig. 2), and is usually fault-bounded (Cediel and Cáceres, 1988; Restrepo Pace, 1992), the latter shows a nearly circular outcrop pattern (Fig. 2), and its contacts are commonly intrusive (Feininger et al., 1970; Cediel and Cáceres, 1988). Tentatively, these relationships may show that the Cordillera Central domain north of the Ibagué fault has undergone little internal distortion since



Fig. 15. Crustal blocks of the northern Andean region used for this reconstruction. Cordillera Oriental–Upper Magdalena in shades of grey, Cordillera Central–Middle Magdalena in line patterns, and Maracaibo block in dotted patterns. Note that each block is subdivided along major fault systems.

intrusion of the Mesozoic Antioquia batholith. In contrast, the Cordillera Oriental and Upper Magdalena Valley are thoroughly deformed, and have accommodated a great thickness of sediments (Sarmiento, 2002), on a severely thinned crust (Roeder and Chamberlain, 1995). The relative weakness of the Magdalena Valley south of the Ibagué fault, and of the Cordillera Oriental may have resulted from crustal thinning following Mesozoic rifting, elevation of geothermal gradients, and the thermal blanketing effect of a thick sedimentary cover. The weak Cordillera Oriental block can be further subdivided using the traces of the largest fault systems (Fig. 15) to attempt to model the rigid-body translations that evidently took place along these systems (Restrepo Pace, 1989; Amézquita and Montes, 1994; Namson et al., 1994).

The third crustal element was defined between the Boconó, Oca, and Bucaramanga-Santa Marta faults. These faults define a roughly triangular block with an intervening northeast-trending foldbelt (Perijá mountains, Kellogg, 1984), a northwest corner out of isostatic equilibrium (Sierra Nevada de Santa Marta, Tschanz et al., 1974), a northeast region limited to the east by allochthonous oceanic sequences (Villa del Cura, Bell, 1971), and a central depression where a great thickness of sediment has accumulated (Maracaibo basin, James, 2000). The relatively undeformed stratal geometry reported in the central part of this block (Maracaibo basin) is evidence of its relative rigidity. The kinematics of two of the bounding faults (dextral Oca, and sinistral Bucaramanga-Santa Marta faults) have been used to propose a general northwestward escape of this block with respect to stable South America (Kellogg and Bonini, 1982). This hypothesis is supported by a GPS study that indicates relative migration consistent with the proposed kinematics (Kellogg et al., 1995), and by a kinematic analysis within the Perijá Range (Kellogg and Bonini, 1982). This hypothesis, however, ignores the third bounding fault on this block (dextral Boconó, Schubert, 1981), as well as paleomagnetic data (Table 2) indicating that this block has undergone large clockwise rotations (Hargraves and Shagam, 1969; MacDonald and Opdyke, 1972; Skerlec and Hargraves, 1980; Castillo et al., 1991; Gose et al., 2003). Some of these paleomagnetic studies have obtained ambiguous results, such as counterclockwise rotation for the Perijá Range (Maze and Hargraves, 1984), or no rotation at all in the Sierra Nevada de Santa Marta (MacDonald and Opdyke, 1984), studies that were rejected here on the basis of the large limits of error reported (Table 2). The alternative hypothesis presented in this paper incorporates all kinematic and paleomagnetic data to model the Maracaibo block as a rigid block that underwent large clockwise rotations that are expressed in the paleomagnetic data and the seemingly contradicting kinematics of the faults bounding this block.

5.1.3. Reconstruction

Table 2

Reconstruction of a hypothetical pre-deformational state of the northern Andes involves two modes of retrodeformation of blocks: first, weak blocks are retrodeformed applying homogeneous simple shear to

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simulate map-scale deformation that otherwise cannot be accounted for regionally. The amounts and directions of angular shear used here agree with quantitative measurements made in the Piedras– Girardot area. Second, rigid-body rotation and translation of blocks (weak or rigid) accounts for displacements measured along and across strike on regional faults and fault systems. The combination of these modes generates a geometric puzzle where gaps between blocks represent crustal shortening taking place along regional fault systems, and distorted grids represent smaller-scale deformation within weak blocks.

Applying the quantitative kinematic data derived from the analysis of the Piedras-Girardot area (angular shear of -40° along N45°E, convergence factor of ~2) to the Cordillera Oriental-Upper Magdalena block, and rotating the Maracaibo block 50° results in large gaps along fault systems that apparently do not accommodate significant amounts of shortening. In addition, the outline of the Ibagué batholith fails to reach a circular outcrop pattern. Using larger values of angular shear $(-55^{\circ} \text{ along})$ N45°E), a closer fit is obtained between blocks, and the Ibagué batholith attains a nearly circular outcrop pattern. Rotation of 75° for the Maracaibo block is necessary to close the gaps along fault systems that do not accommodate shortening, well within the range allowed by paleomagnetic data (Table 2). The second alternative is preferred here because the quantitative kinematic data gathered in the Piedras-Girardot area,

Author	Location	Sites/ localities of interest	Lithology	Age	Results	Age of rotation	α95, degrees
Hargraves and Shagam, 1969	Mérida Andes	84/4	Dacitic tuff	Triassic–Jurassic	Large clockwise	Post-Eocene	25
MacDonald and Opdyke, 1972	Guajira peninsula	9/2	Lava	Late Jurassic	90° clockwise	Post-Cretaceous	19.1
Skerlec and Hargraves, 1980	Caribbean Mountains	15/2	Mafic volcanic/ Schist/Gneiss	Cretaceous	90° clockwise	Cretaceous– Paleocene	21.2 (av.)
MacDonald and Opdyke, 1984	Sierra Nevada Santa Marta	10/3	Red beds/tuff	Triassic–Jurassic	None	n/a	26.7
Maze and Hargraves, 1984	Perijá	32/4	Red beds/dike	Triassic–Jurassic	20–60° counter- clockwise	Early Cretaceous	64 (av.)
Gose et al., 2003	Perijá	11/4	Various	Jurassic to Eocene	$50^{\circ} \pm 12$ clockwise	Neogene	9.6



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while likely reflecting the structural style dominant in the Cordillera Oriental-Upper Magdalena, do not necessarily record the average amounts of deformation throughout the entire system. Thus, in this reconstruction preference was given to the undeformed state that contains smaller unexplained gaps or overlaps (Fig. 16a). An undeformed state was thus constructed applying an angular shear of -55° along N45°E to the weak blocks of the Cordillera Oriental-Upper Magdalena, and translating the thrust sheets along a N71°E vector, oblique to structural trends. Reconstruction of the semi-rigid Cordillera Central-Middle Magdalena involves lesser amounts of angular shear $(-20^\circ, \text{ along N45}^\circ\text{E})$. The shortening component perpendicular to the structural trends was derived from standard local and regional cross-sections, most notably those with direct field measurements or seismic reflection data (Sarmiento, 2002). The Maracaibo block was rotated 75° until it closed the gaps opened by shearing and translation in the other two blocks.

Once the preferred hypothetical undeformed state is chosen (Fig. 16a), forward deformation can be applied step by step using the east-to-northeast propagation of the Caribbean deformation front with respect to stable South America (Pindell et al., 1998) to progressively drive deformation in the northern Andes. The contrasting mechanical behavior allowed for the three blocks causes simultaneous movement along dextral and sinistral strike-slip faults, dextral transpression, clockwise rotations, and extensional opening of basins. For instance, the sinistral slip along the Santa Marta-Bucaramanga fault is compatible with simultaneous dextral slip along the Oca and Mérida faults (Fig. 16c-f). No further constraints are used to control the timing of deformation, keeping the model simple and predicting a younger deformation age to the northeast and east as the deformation front advanced. The model predicts a dip-slip shortening component of approximately 120 km along a hypothetical NW-SE, two-dimensional cross-section (Fig.

16f), at about same latitude as other two-dimensional cross-sections of the Cordillera Oriental that suggest similar dip-slip shortening components of deformation (105 km, Colletta et al., 1990; 150 km, Dengo and Covey, 1993).

Such simple reconstruction highlights the possibility of combining, in a single kinematic framework, most of the observed puzzling kinematic features of the northern Andes with the regional kinematics of the Caribbean Plate. It also demonstrates that dextral transpressional deformation, driven by the advance of the Caribbean deformation front, can adequately explain the regional structure and evolution of this complex margin.

This model provides a plausible alternative conceptual framework for the interpretation of the northern Andes. This model is based on the kinematic understanding of the influence of the Caribbean Plate, and the application of kinematic compatibility criteria. From a critical review of the literature, it is clear that many solutions to this puzzle are possible, and that as long as kinematic data are systematically ignored, it will remain that way. It is hoped, though, that this model serves to plan intelligent data collection in the northern Andes by having highlighted key areas, and hypotheses to test.

6. Conclusions

The Piedras–Girardot area is a dextral transpressional system where approximately 32 km of ENE– WSW contraction is recorded as a result of the ENE insertion of a rigid block of the Cordillera Central within a N45E-trending transpressional shear zone with a shear strain of 0.8 and a convergence factor of 2.0. Microscopic and mesoscopic fabric elements in the Piedras–Girardot area record less than 5% shortening in a general northwest direction, and less than 5% extension in a northeast direction. The orientation of all fabric elements is oblique to the independently

Fig. 16. Speculative sequential restoration of the northern Andean blocks. (a) Predeformational state. (b) Northward translation of the rigid Cordillera Central block. (c) Fragmentation of the rigid Cordillera Central block as activity along the sinistral Santa Marta–Bucaramanga fault , and rotation of the Maracaibo block begin. (d) Initiation of significant dextral transpressional deformation in the Cordillera Oriental–Upper Magdalena block, further rotation of the Maracaibo block, causing the emplacement of the Villa del Cura rocks on the South American margin. (e) Rotation of the Maracaibo block is almost complete, and extensional opening of the Falcon–Bonaire basin starts as the Caribbean deformation front continues to migrate east, result of a right-lateral, releasing bend (Muessig, 1984).

constrained relative direction of tectonic transport (ENE). Late Campanian deformation fabrics like cleavage and veins started to develop after the initial propagation of the northernmost, northeast-trending segments of the Camaito and Cotomal faults, and the El Guaco anticline. These structures were later overlapped by the La Tabla Formation conglomerate, which records Maastrichtian unroofing in the Cordillera Central. The early Paleogene marks a time of segmentation of the accumulation environment due to the relative west- or southwestward propagation of the Cambao fault, and generation of accommodation space in the Guaduas and Gualanday synclines. Paleogene and younger deformation, although spectacularly recorded by thick, folded molasse deposits and map-scale faults and folds, lacks a mesoscopic deformation fabric. Earlier fabric elements were passively rotated along horizontal axes and translated within propagating thrust sheets. This foldbelt has been a positive area since then, shedding clastic material into actively deforming Paleogene depocenters. Only the northern part of the study area contains evidence for post-Miocene deformation, which is related to the latest activity along the Ibagué fault.

The orientation of fabric elements and the magnitude of internal strain are a function of horizontal distance to the Ibagué fault: the long axes of strain markers become closer to the trend of this fault as the distance to it decreases. Similarly, internal strain magnitude decreases as distance to the fault increases. Map-scale faults and folds are oblique to the independently constrained relative direction of tectonic transport. The ENE-trending Ibagué fault may represent one of the synthetic shears in a regional northeast-trending dextral shear zone parallel to the overall trend of the Cordillera Oriental where the orientation of fabric elements would initially be oriented north-south and rotated progressively toward orientations closer to the boundary of the shear zone as deformation progressed.

Three continental blocks: the rigid Maracaibo, the semi-rigid Cordillera Central, and the weak Cordillera Oriental blocks interacted complexly to generate simultaneous dextral and sinistral transpression, large clockwise rotation, and extension along the northwestern margin of South America. Each of these blocks was permitted here to accommodate deformation differently according to its relative rigidity. Rigid blocks accommodate deformation by rigidbody rotation and translation, whereas weak blocks accommodate deformation by internal distortion and dilation. This deformation was driven by the east- to northeast advance of the Caribbean deformation front with respect to stable South America. Values of strain, timing of deformation, tectonic transport direction, and structural style derived from the analyses made in the Piedras-Girardot area were used to perform this regional reconstruction. The resulting model explains seemingly incompatible kinematic situations recorded in the northern Andes such as the simultaneous movement on dextral and sinistral strike-slip faults, dextral transpression, and large clockwise rotation.

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