Unravelling the widening of the earliest Andean northern orogen: Maastrichtian to early Eocene intra-basinal deformation in the northern Eastern Cordillera of Colombia

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Abstract
The onset of deformation in the northern Andes is overprinted by subsequent stages of basin deformation, complicating the examination of competing models illustrating potential location of earliest synorogenic basins and uplifts. To establish the width of the earliest northern Andean orogen, we carried out field mapping, palynological dating, sedimentary, stratigraphic and provenance analyses in Campanian to lower Eocene units exposed in the northern Eastern Cordillera of Colombia (Cocuy region) and compare the results with coeval succession in adjacent basins. The onset of deformation is recorded in earliest Maastrichtian time, as terrigenous detritus arrived into the basin marking the end of chemical precipitation and the onset of clastic deposition produced by the uplift of a western source area dominated by shaly Cretaceous rocks. Disconformable contacts within the upper Maastrichtian to middle Palaeocene succession document increasing supply of quartzose sandy detritus from Cretaceous quartzose rocks exposed in eastern source areas. The continued unroofing of both source areas produced a rapid shift in depositional environments from shallow marine in Maastrichtian to fluvial-lacustrine systems during the Palaeocene-early Eocene. Supply of immature Jurassic sandstones from nearby western uplifts, together with localized plutonic and volcanic Cretaceous rocks, caused a shift in Palaeocene sandstones composition from quartzarenites to litharenites. Supply of detrital sandy fragments, unstable heavy minerals and Cretaceous to Ordovician detrital zircons, were derived from nearby uplifted blocks and from SW fluvial systems within the synorogenic basin, instead of distal basement rocks. The presence of volcanic rock fragments and 51–59 Ma volcanic zircons constrain magmatism within the basin. The Maastrichtian–Palaeocene sequence studied here documents crustal deformation that correlates with coeval deformation farther south in Ecuador.
and Peru. Slab flattening of the subducting Caribbean plate produced a wider orogen (>400 km) with a continental magmatic arc and intra-basinal deformation and magmatism.

KEYWORDS
northern Andes, onset of deformation, Palaeocene, provenance, synorogenic basins

1 INTRODUCTION

Active orogens that started as ancient passive margins (or post-rift system in interior continental rifts) still contain the record of the transition from passive to active tectonic settings represented by drastic shifts in depositional environments, provenance, magmatism, tectonic subsidence and sediment supply rates across unconformities (e.g. Bisbee basin in SW USA, Dickinson & Lawton, 2001, or the Atlas, Eastern Cordillera and Pyrenees, Babault et al., 2013). These changes can be used to document the tectono-magmatic evolution of plate margins like the northwestern corner of South America (Figure 1), where an ancient continental rift system (Sarmiento-Rojas, 2018) evolved into an active orogen during the latest Cretaceous to Palaeogene times (Bayona, 2018; Cooper et al., 1995; Mora et al., 2013; Villamil, 1999; Figure S1). This record, however, may be obscured if the passive margin (or post-rift) succession has been involved in successive phases of deformation, as is the case of the Cretaceous–Palaeogene sedimentary sequence in the northern Andes (Figure 1).

During the latest Cretaceous–Palaeogene, a major change in depositional conditions in the northern Andes is recorded by the disconformable contact between coarse-grained continental sandstones supplied from the orogenic wedge resting upon fine-grained marginal deposits supplied from cratonic sources (see review in Horton, 2018). Recent thermochronological and provenance studies carried out in the Eastern Cordillera of Colombia (Figure 1) have provided age constraints on deformation and cooling of the orogen since the Oligocene, with increased cooling rates during the late Miocene (e.g. Horton et al., 2010; Mora, Casallas, et al., 2015; Mora et al., 2019; Mora, Parra, et al., 2015; Ochoa et al., 2012; Parra, Mora, Sobel, Streecker, & González, 2009; Parra, Mora, Lopez, Rojas, & Horton, 2010; Siravo et al., 2018, 2019). However, only a few studies have shown support for exhumation during the latest Palaeocene to early Eocene times (Caballero, Mora, et al., 2013; Parra, Mora, Lopez, Rojas, & Horton, 2012; Velandia, 2018).

Presently, the Eastern Cordillera of Colombia exposes Campanian to Palaeocene sedimentary rocks that contain the record of both the transition from marine to continental deposition (Figures 1 and 2a) and the transition from passive to active margin sedimentation. These rocks allow us to examine competing models illustrating potential location and timing of the onset of basin inversion. Did synorogenic sedimentation take place in a single basin adjacent to an orogenic belt restricted to the west (Goméz et al., 2005; Horton et al., 2010, 2015; Nie et al., 2012; Parra et al., 2009; Figure 2b)? Or, in contrast, was this process compartmentalized by intra-basinal uplifts that bounded closed basins (Bayona et al., 2013; Reyes-Harker et al., 2015; Figure 2c,d)? The latter configuration would imply that deformation and magmatism were not restricted to the west, as in other segments of the Andes in Maastrichtian–lower Eocene synorogenic succession (Figure 1).

HIGHLIGHTS

• Slab flattening of the subducting Caribbean plate produced Palaeocene widening of the northern Andes
• Palaeocene deformation included magmatic arc
• Intra-basinal deformation and magmatism.
• Sedimentological and provenance data allow the identification of three separated basins
• Volcanic rock fragments and 51–59 Ma volcanic zircons suggest volcanism within the basin
• Two types of unconformities were identified in the Maastrichtian–lower Eocene synorogenic succession

We studied the initial phases of deformation in the northern Andes by integrating field mapping, with analyses of
palynology and provenance. We focused on the Cocuy region, an area located in the northernmost axial region of the Eastern Cordillera (Figure 1), where we document the earliest phases of deformation and erosion of shallowly buried sedimentary cover during the transition from passive margin to an active orogen. We document early phases of deformation by mapping the increasing supply of terrigenous sediments across unconformities of the Eastern Cordillera and adjacent basins of the northern Andes. This transition took place during the warmest period of the Cenozoic (Zachos, Pagani, Sloan, Thomas, & Billups, 2001), an interval of high temperatures and precipitation in the tropics (Jaramillo et al., 2007;}

FIGURE 1 (a) Location of the study area in the northern Andes. (b) Location of the study area (numbers 1–3) in the Central segment of the Eastern Cordillera of Colombia bounded by the Magdalena Valley to the west and the Llanos basin to the east (see Figure 2 for palinspastic distance in Palaeocene time). Note the unconnected distribution of the three Palaeocene clastic wedges: C1: sublitharenite–litharenite (Seca-Hoyón-Lisama); C2: quartzarenite (Cacho–Socha Sandstone) to sublitharenite–litharenite (Bogota–Socha Mudstone); C3: Quartzarenite (Barco) to sublitharenites–litharenites (Los Cuervos). Proposed source areas for those clastic wedges include: S1 = Central Cordillera basement rocks and Cretaceous sedimentary cover (Permo-Triassic metamorphic rocks; Lower to Middle Jurassic and Upper Cretaceous igneous rocks; Sedimentary cover from these sources); S2 = Structures of the western flank of Eastern Cordillera (shaly Cretaceous units, lithic- and feldspar-bearing Jurassic rocks); 80–90 Ma volcanic rocks; Igneous mafic rocks = 74–136 Ma; S3 = Santander Massif and sedimentary cover (Upper Triassic and Ordovician igneous rocks; Grenvillian metamorphic rocks; 98–108 Ma dykes); S4 = Structures of the eastern flank of the Eastern Cordillera (Lower Cretaceous rocks supplied from Grenvillian metamorphic and Palaeozoic rocks; Upper Cretaceous rocks supplied from cratonic sources; igneous mafic rocks = 74–136 Ma); S5 = Cretaceous sedimentary cover of the Llanos basin (Upper Cretaceous quartzarenites supplied from cratonic and Palaeozoic rocks); S6 = Palaeozoic and Guiana cratonic rocks (cratonic sources; >1.3 Ga; see text for references)
FIGURE 2 (a) Restored basin geometry for the Campanian (ca. 75 Ma) showing lateral extension of (1) basement rocks with location of Ordovician to Cretaceous magmatic events and (2) Jurassic to Cretaceous sedimentary thickness and lateral change in lithologies (modified after Reyes-Harker et al., 2015; Siravo et al., 2018; Tesón et al., 2013). Note the location of reported Cretaceous magmatic activity (plutonic activity was projected to the cross section; see text for references). (b–d) Schematic diagrams showing proposed hypothesis for the location of source areas and basin geometry in late Palaeocene at the latitude of the Cocuy area. Distances from source areas (S1–S6) to basin(s) where clastic sediments accumulated (C1–C3) taken from the late Palaeocene palaeogeographic map of Montes et al. (2019)
Jaramillo, Rueda, & Mora, 2006) that favoured the transport of terrigenous material and intensive palaeosol development in subaerial plains (Morón et al., 2013).

2 | REGIONAL GEOLOGICAL SETTING

The Eastern Cordillera of Colombia is a doubly vergent orogen that resulted from the Cenozoic inversion of a Cretaceous continental rift system (Colleta, Hébrard, Letouzey, Werner, & Rudkiewicz, 1990; Cooper et al., 1995; Mora, Casallas, et al., 2015; Mora, Parra, et al., 2015; Sarmiento-Rojas, Van Wess, & Cloetingh, 2006; Siravo et al., 2018; Figure 1), which, in turn, had overprinted other phases of Jurassic continental rift extension (Bayona, Bustamante, Nova, & Salazar-Franco, 2020; Sarmiento-Rojas et al., 2006). West of these continental rift systems, both the Cretaceous back-arc basin and volcanic arc are present and involved in the structures of the western Central Cordillera (Zapata et al., 2019).

The Eastern Cordillera contains 3- to 10-km-thick stratigraphic sequences of Jurassic to Cretaceous rocks that have been interpreted as the primary supply of sediments during the early stages of deformation (Bayona et al., 2013; Caballero, Mora, et al., 2013; Figure 2a). To the west of the Santander massif (Figure 1), Upper Jurassic immature feldspar- and lithic-bearing sandstones and conglomerates filled extensional basins (Figure 2a) overlying unconformably Upper Triassic to Middle Jurassic plutonic, volcanic and volcanoclastic strata (Bayona et al., 2020; Sarmiento-Rojas, 2018; Sarmiento-Rojas et al., 2006). Jurassic rocks are not recorded east of the Santander massif (Figure 1). During the Early Cretaceous, two extensional basins formed to the east (Cocuy graben) and west (Tablazo graben) of the Santander massif (Figure 2a). In the eastern graben, sandy quartzose units dominated and the Cretaceous succession became sandier eastward, whereas in the western graben the Lower Cretaceous succession is mostly muddy and calcareous; sandstone units accumulated adjacent to the Santander massif or to the proto-Central Cordillera (Cooper et al., 1995; see Sarmiento-Rojas, 2018 for regional reconstruction of these two basins). Our study area is located near the extensional faults that bound the Cocuy graben on its western edge, where the Lower Cretaceous succession is thin and proximal to the Santander massif (Figure 2a).

Early Cretaceous rifting was succeeded by thermo-tectonic subsidence beginning in Albian time (Sarmiento-Rojas et al., 2006). During the Cenomanian and Turonian, extensive marine flooding events with anoxic waters covered the Santander massif and the palaeo-Central Cordillera, while the coastline moved eastward to the central Llanos basin (Sarmiento-Rojas, 2018). Both thermal and compaction subsidence favoured the accumulation of sandstone units along the Llanos and eastern margin of the Cocuy graben, whereas the rest of the shallow and open marine basin to the west and north is dominated by black shales, chert and limestone units (Sarmiento-Rojas, 2018).

The Coniacian–Palaeocene collision and accretion of intra-oceanic arcs with the continental margin switched the continental margin to an active orogen with synorogenic basins that developed on both the oceanic and continental sides (Pardo-Trujillo et al., 2020). Two source areas filled the basin formed on the continental side of the orogen. A single and distal eastern crustal source (Guiana craton) for the Coniacian to Campanian siliciclastic wedge that accumulated to the east (Sarmiento-Rojas, 2018; Sarmiento-Rojas et al., 2006), and to the west, erosion of the Central Cordillera, which supplied igneous and metamorphic rock fragments and reworked the sedimentary filling of back-arc basins (Guerrero, Sarmiento, & Navarrete, 2000; Zapata et al., 2019).

Several types of Cretaceous volcanic and intrusive rocks have been documented in the Eastern Cordillera (see Figures S11–S14 for detailed explanation of these volcanic deposits and magmatic intrusions). Localized pyroclastic tuffs and breccias interbedded with Turonian-Coniacian shallow-marine deposits (Martínez, 2010) indicate the presence of nearby intermediate volcanism, similar to the local Cenomanian volcanic cones documented in the Ecuador foreland basin (Baby, Rivadeneira, Barragán, & Christophoul, 2013). Beds of volcanic ash-fall deposits (bentonites) have been reported as thin interbeds in Turonian to Santonian marine black shales (Ballesteros-Torres et al., 2013; Terraza, 2012; Figure 2a). Local mafic Cretaceous magmatic intrusions affected the Cretaceous graben systems (Vasquez, Altenberger, Romer, Sudo, & Moreno-Trujillo, 2010; Figure 2a). Additionally, microdiorite and sienogranite dikes intruded the Santander massif in the Cretaceous (Correa, Rodríguez, Arango, Zapata, & Bermúdez, 2016).

Regional accumulation of the upper Campanian–Palaeocene synorogenic units along the Magdalena Valley (Umir, Seca and Lisama formations), the Eastern Cordillera (Guaduas–Cacho–Socha Sandstone formations) and the Llanos–Catatumbo-Cesar basins (Guaduas–Barco–Los Cuervos formations) records the increased supply of sandy continental detritus to the synorogenic basins (clastic wedges C1–C3 in Figure 1; Bayona, 2018). These units reveal the transition from shallow marine-marginal Cretaceous accumulation to continental deposition during the early Palaeocene time (Silva, Mora, & Cosgrove, 2013; Reyes-Harker et al., 2015; Sarmiento-Rojas, 2018; Villamil, 1999). However, the mechanism that drove this regional transition is still debatable, owing to the different hypotheses about the timing of onset of deformation in the Eastern Cordillera and surrounding basins (see discussion below) and how the first-order fall of sea level affected the filling of Palaeocene basins (e.g. Cooper et al., 1995).
Detrital zircon geochronology and sandstone petrographic studies have been recently used as indicators of source area composition and location for upper Campanian–Palaeocene synorogenic units. Recent provenance studies have assumed that detrital zircons younger than 150–170 Ma are derived from the Central Cordillera basement (e.g. Nie et al., 2010 and Nie et al., 2012; Silva et al., 2013; Reyes-Harker et al., 2015; Odoh, Saylor, Higuera, Copeland, & Lapen, 2019; Figure 2b). Erosion of basement rocks from the Central Cordillera or Santander massif had been proposed to explain the increase in metamorphic lithic fragments in Palaeocene units in the Cocuy area, Catatumbo and Floresta basins (Ayala-Calvo et al., 2012; Fabre, 1981; Montenegro et al., 2012; Odoh et al., 2019; Saylor, Horton, Nie, Corredor, & Mora, 2011), and Nuevo Mundo syncline in the Magdalena Valley (Caballero, Mora, et al., 2013; Caballero, Parra, et al., 2013; Moreno et al., 2011; Naranjo, Gómez, Gélvez, Duque, & Moreno, 2013; Nie et al., 2012; Prince et al., 2016). However, these studies did not consider that (a) local Cretaceous plutonic and volcanic rocks within the Eastern Cordillera and Santander massif could be a possible source of igneous rock fragments and detrital zircons or (b) the erosion of Jurassic sedimentary rocks could provide igneous and metamorphic rock fragments and zircons of 200–480 Ma ages; this population has been reported in the Jurassic samples documented in Horton et al. (2010) and Saylor et al. (2011).

Several cross-section restorations have been used to argue that deformation during pre-middle Eocene time was either null (Mora et al., 2013; Tesón et al., 2013) or very incipient (Bayona et al., 2008; Teixell, Teson, Ruiz, & Mora, 2015; Figure S2). However, multiple lines of evidence suggest a late Maastrichtian to Palaeocene onset of deformation in the Eastern Cordillera. Thermochronology, provenance and sedimentological data that support late Maastrichtian to Palaeocene deformation are found in buried palaeo-highs in the northern Middle Magdalena basin (Parra et al., 2012), the western foothills of the Eastern Cordillera (Bayona et al., 2013; Caballero, Mora, et al., 2013; Montaño et al., 2016; Parra et al., 2012), and the Santander massif (Prince et al., 2016). Additionally, reworking of Cretaceous material in Palaeocene rocks has been reported in the Middle Magdalena basin (Navarrete-Parra, et al., 2015), western foothills (Caballero, Mora, et al., 2013; Montaño et al., 2016), Floresta Basin (Céspedes & Peña, 1995; Pardo & Jaramillo, 2013) and Llanos foothills and Llanos basin (de la Parra, Mora, Rueda, & Quintero, 2015). Geodynamic models also indicate that tectonic loads in the Eastern Cordillera are needed to explain the geometry of the Palaeocene Llanos foreland basin (Bayona et al., 2008, 2009, 2013). Additional lines of evidence supporting Palaeocene deformation, including local angular unconformities, paraconformities, growth strata and fault-related emerald mineralization are summarized in Figure S3.

Three tectonic configurations have been proposed to explain the basin geometry and location of source areas during the Palaeocene. The first configuration proposes a single 740-km-wide foreland basin bounded to the west by the Central Cordillera with east-verging deformation (Goméz et al., 2005; Horton et al., 2010, 2015; Nie et al., 2010, 2012; Odoh et al., 2019; Parra et al., 2009; Saylor et al., 2011; Saylor, Stockli, Horton, Nie, & Mora, 2012; Figure 2b). The second configuration proposes a foreland basin broken by western uplifts since the Palaeocene, based on thermochronological results from structures buried in the Magdalena Valley or exposed in the western foothills of the Eastern Cordillera (Parra et al., 2012; Caballero, Parra, et al., 2013; Silva et al., 2013; Reyes-Harker et al., 2015; Mora et al., 2019; Figure 2c). Teixell et al. (2015) concluded that the foreland basin was affected by low-amplitude uplifts located at the western, axial and eastern margin of the Eastern Cordillera, but this deformation did not affect the lateral continuity of the foreland basin (Figure S2).

An alternative to this second configuration is that of Restrepo-Pace, Colmenares, Higuera, and Mayorga (2004), who proposed an eastward progradation of deformation beginning with the uplift of the Central Cordillera followed by Palaeocene-early Eocene thrusting affecting the Magdalena Valley, the western margin of the Eastern Cordillera and the Santander massif to the west.

The third tectonic configuration calls for a different geodynamic model because it considers: (a) a wider zone of deformation from the Central Cordillera to the proximal Llanos basin; (b) multiple synorogenic basins associated with crustal tilting to the west, structural inversion within, and foreland deposition to the east and (c) Palaeocene magmatism both to the west and within the basins (Ayala-Calvo et al., 2012; Bayona et al., 2008, 2011, 2012, 2013, 2015; Bustamante et al., 2016; Cardona et al., 2011, 2014; Figure 2d).

## METHODS

### 3.1 Field work

The Cocuy region, in the northern segment of the axial zone of the Eastern Cordillera, includes NNW-striking beds of Palaeogene age (Fabre, Osorio, Vargas, & Etayo, 1984, 1985). Three areas were selected (Figure 1) and the Upper Cretaceous and Palaeocene units in each were described. In stratigraphic order, these are as follows: La Luna, Los Pinos, Guaduas, Socha Sandstone (Areniscas de Socha) and Socha Mudstone (Arcillas de Socha) formations. The description of each stratigraphic unit includes (a) lateral variation of thickness; (b) dominant lithological associations and morphology of the unit; (c) compositional results based on petrographic analysis; (d) heavy mineral and detrital geochronology data; (e) palaeocurrent indicators; (f) age control based on
palynological analysis and (g) interpretation of depositional environment.

Digital outcrop models, based on Structure from Motion techniques, were generated from photos collected systematically in terrain (see procedures in Fabuel-Perez, Hodgetts, & Redfern, 2010 or in Bilmes et al., 2019). These digital models were used to define the geometry of strata and palaeocurrents (Duarte, Bayona, Ramírez, Baquero, & Tabares, 2019; Gómez, 2019), to add mesoscale information to the stratigraphic analysis. Strike and dip data from major structures involving Upper Cretaceous and Palaeogene strata were used to produce structural cross sections, allowing a comparison of stratigraphic thickness within a given structure (Figure 3) and among the three study areas (Figures S3–S6).

Ecopetrol (2004) supplied one stratigraphic column from the northern area 1 (our Section 1, Arteza) measured by

FIGURE 3 (a) Geologic map of the Las Mercedes syncline (northern area 1 in Figure 1) showing the location of the three stratigraphic sections and dip angles of Palaeocene units between the western and eastern flanks. (b) Structural cross sections showing the westward increase in thickness of lower Guaduas and Socha Mudstone formations.
Jacob staff at scale 1:200 from the top of La Luna to lower beds of the Socha Mudstone formations, where geochronological data had already been reported by Silva et al. (2013) and petrographic data by Bayona et al. (2008). Two additional stratigraphic columns in the northern area 1 were also constructed using the information from the geologic map, cross sections and field stations (Section 2, Ventura and Section 3, San Roque). Stratigraphic and petrographic data from Gómez (2019) were included for Section 2. For the central area 2 and southern area 3, composite stratigraphic columns were constructed (Section 4, Panqueba; and Section 5, La Conquista, respectively). Palynological analysis provided time control to constrain the lateral correlation of lithological associations in these stratigraphic columns.

3.2 | Palynological analyses

We analysed 44 palynological samples of Upper Cretaceous–Palaeocene strata to provide age control and help to assess depositional environments. Palynological samples were prepared by digesting 30 g of sedimentary rock per sample in hydrochloric acid (HCl) and hydrofluoric acid (HF), to remove carbonates and silicates (Traverse, 1988). The residual organic matter was separated into two fractions; one of them was then oxidized with nitric acid (HNO3) to eliminate unstable organic particles and concentrate the palynomorphs (Traverse, 2007). The oxidized part was used for identifying the organic matter compounds and recognizing environmental and thermal alteration signals. Light microscopy was used for routine palynological analyses, and at least 200 palynomorphs per sample were counted where possible. Measurements of Thermal Alteration Index (TAI) followed the procedure of Traverse (2007). Fossils were identified by comparing them with a morphological electronic database that contains every pollen/spore taxon described for the tropics of South America (Jaramillo & Rueda, 2019). The dating of the samples follows the biostratigraphic zonation scheme proposed by Jaramillo, Rueda, and Torres (2011) for the Cretaceous. This zonation has been calibrated with magnetic stratigraphy, carbon isotopes, radiometric ages and microfossil biostratigraphy (Jaramillo et al., 2011). Samples yielded very high thermal alteration indices ranging from 3 to −4, suggesting mature rocks (~1 to 1.6% Ro), thus complicating the identification of palynomorphs.

3.3 | Provenance analysis

We integrated palaeocurrents, petrography, heavy minerals and detrital zircon geochronology to understand whether sediment supply was from nearby structures (e.g. Santander massif—intra-basinal uplifts) or from catchment areas located at a relatively large distance (e.g. Central Cordillera—Guiana craton). Palaeocurrent indicators (119 measurements in Duarte et al., 2019 and 50 new data for the Socha Mudstone) were measured in cross-stratified beds both in the field and using digital outcrop models. When possible, a pair of planes from trough cross stratification were used to calculate the axis of the trough, and a single plane was used for planar cross beds.

Sandstone petrography was conducted in 27 samples and the results were integrated with the results of previous petrography studies from the same area (6 samples from Bayona et al., 2008; and 10 samples from Gómez, 2019; Table S1). In all, 300 framework points were counted per thin section using the Gazzi-Dickinson technique (Ingersoll et al., 1984). Thin sections were stained for identification of potassium feldspars. Detrital modes exclusive of carbonate, glauconite, intraclasts and heavy mineral grains were calculated from the point-count results following the technique of Dickinson (1985) and were plotted in QtFL and QmFLt ternary diagrams. Point counting results and recalibration results of petrographic analysis are in the Table S1.

Five samples encompassing Maastrichtian to upper Palaeocene units were selected for analysis of heavy minerals. Samples were crushed, sieved and hydraulically concentrated on the Wilfley table. The 63–250 µm fraction was selected to minimize the hydraulic sorting effect and large apparent discrepancies on the mounts, following Mange and Maurer (1992) and Morton (1985). Minerals with density above 2.89 gr/cm³ were obtained using sodium polytungstate. Mounts were prepared using the Meltmount® resin with a refraction index of 1.539. A minimum of 300 translucent minerals were optically identified following the ribbon method (Mange & Maurer, 1992). Results of the heavy minerals conventional and varietal analysis are given in the Table S2.

We used the detrital zircon geochronology database published by Silva et al., 2013; see methods in the supplementary material of that paper) to select 10 samples that were collected all along Section 1, in the northern area 1. Here, we focused on the identification of detrital zircon ages younger than Ordovician (485 Ma) to constrain the location of source areas. If Palaeozoic to Mesozoic basement rocks were exposed in the source areas, we would expect the presence of ages younger than 485 Ma together with older ages (485–2800 Ma). If only the Cretaceous sedimentary cover was exposed in the source areas, the relative abundance of ages younger than 485 Ma would be rare (<5%) in comparison with older ages (485–2800 Ma; see Odoh et al., 2019 for the regional review of published geochronology data). Reported data from Silva et al. (2013) are presented in the Table S3. In our analysis, we report the 238U/206Pb age if the zircon
was less than 1.1 Ga, and the 207Pb/206Pb age if the grain was older than 1.1 Ga. According to Spencer, Kirkland, and Taylor (2016) criteria for young grains, the 206U/238Pb age is more precise than the 207U/206Pb age due to low 207Pb intensity. The cross-over in precision occurs at ~1.4 Ga, but due to the tendency of zircons to have more Pb loss with age, a threshold that compromises precision and accuracy for old grains is commonly chosen at 1.0 Ga (Gehrels, 2012). We rejected results that had a 2σ error of >10% and those that were >10% discordant for ages younger than 500 Ma or >25% discordant for ages older than 500 Ma. To differentiate samples that are likely similar or dissimilar, we used the quantitative method of multi-dimensional scaling (MDS) as outlined by Vermeesch (2013) and Vermeesch, Resentini, and Garzanti (2016). The distance between sample points allows us to propose supply from similar/different source areas. The MDS plots were developed using R Isotplot package described by Vermeesch (2018) and available from https://www.ucl.ac.uk/~ucfpve/isoplotr/.

4  RESULTS

4.1  Geologic mapping of Campanian–Palaeocene strata

In the northern area, the Mercedes syncline is an asymmetrical NNW-plunging fold with 45–65° dip values in the western flank and 10–30° dip values in the eastern flank (Figure 3). Good rock exposures at >4 km of elevation allowed strong control of stratigraphic thickness of Campanian to Oligocene strata in both flanks (Figure 4a and Figure S3). Local faults and folds affect the La Luna formation and older strata in both flanks. In the northern segment of the eastern flank, stratigraphic Section 1 (Arteza; Figure 5) surveyed rocks from the top of La Luna to lower beds of the Socha Mudstone formations. A second stratigraphic column (Section 2 Ventura, Figure S7) was constructed based on a local structural cross section in the southern segment of the eastern flank of the syncline (Figure 3 and Figure S6) from the top of La Luna to lower beds of the Socha Sandstone formations. A third stratigraphic column (Section 3 San Roque, Figure 6) was constructed based on a local structural cross section in the western flank of the syncline from the top of the Campanian La Luna to lower beds of the middle Eocene Picacho formation (Figure 3).

The central Panqueba area (area 2 in Figure 1 and Figure S4) corresponds to a structural block that includes strata of the upper Guaduas and Socha Sandstone formations with nearly horizontal attitude, facilitating the continuous trace of the lithological contacts and collection of palynological samples between these two units (Figure 4b). This structural block is cut by several high-angle normal faults, and the area has been affected by recent mass movement processes. We limited our control to the block with horizontal bed attitude for stratigraphic control (Figure S8).

In the southern La Conquista area (area 3 in Figure 1), Campanian to Palaeocene rocks are exposed in two structural blocks bounded by northwest-verging inverse faults (Figure 4c and Figure S5). The eastern block includes Campanian to lower Palaeocene strata with moderate to gentle dip angles to the west. In contrast, the western block exposes lower Palaeocene to Oligocene strata with moderate to high dip angle to the east; locally these strata are overturned allowing a good exposure of the complete Palaeocene succession at >4 km of elevation (Figure 4c). A stratigraphic column was constructed based on field descriptions of a local structural cross section in the western block from the middle interval of the Guaduas formation to lower beds of the Picacho formation (Figure S9).

4.2  Lithological associations and contacts in studied areas

4.2.1  La Luna formation

The best exposures of this unit are in the eastern flank of the Mercedes syncline (Figure 3). Uppermost strata of this unit include medium to thin beds with tabular geometry of black siliceous siltstone and chert beds interbedded with shales and minor packstone beds with fragments of bivalves and phosphatic fragments (fish remains; Fabre, 1981). These chert and limestone rocks show an orthogonal fracture system and generate hilly topography with moderate slopes (Figure S3). Detrital zircon age populations from three samples are >485 Ma (n = 321, 4 were rejected); only two zircons had younger ages: 364 Ma and 444 Ma from samples U16 and 444 Ma from U15, respectively (Figure 5; see Table S3 for age errors). Palynological analysis from five samples in Section 1 from the northern area recovered dinoflagellates, pseudo-amorphous organic matter and woody phytoclasts with high thermal alteration. A Santonian–Campanian age is reported, based on ammonites (Etayo-Serna, 1985). Fine-grained lithological association and biostratigraphic data indicate that these deposits accumulated in offshore marine environments under anoxic conditions.

4.2.2  Los Pinos formation

The best exposures of this unit are in the eastern flank of the Mercedes syncline (Figure 3), where the stratigraphic thickness ranges from 179 m (in Section 2, Figure S6) to 235 m (in Section 1, Figure 3). In Section 2, the lower segment consists of thin beds of shales with a ‘pencil’-like fracture
FIGURE 4  Disconformable contacts bounding the upper Guaduas interval (discontinuous lines: top = red; base = light blue): (a) In the northern area 1: west-dipping beds showing minor truncation in the upper contact; (b) In the central area 2: meter-scale truncation of the fluvial Socha Sandstone over wavy-bedded and heterolithic laminated sandstones of the upper Guaduas; inset shows the disconformable contact between the middle and upper Guaduas intervals; (c) In the southern area 3: the upper Guaduas consists of light-coloured mudstone (valley) and a 3-m-thick fluvial channel deposit, whereas medium tabular beds of siltstones are common in the middle Guaduas interval. Palynological data indicate marine influence in the northern and central areas, whereas continental organic matter was recovered in the southern area.
FIGURE 5  Stratigraphic Section 1 (Arteza, see location in Figure 3) at the eastern flank of the Mercedes syncline (northern area 1 in Figure 1) with petrographic, palynological and detrital geochronology results. MDS diagram illustrates how the Socha Sandstone sample (U37) plots slightly apart from older samples, suggesting the input of new detrital zircons, and the Socha Mudstone sample (U70) plots apart suggesting the input of detrital zircons from a totally different source area than the other samples.
**FIGURE 6**  Stratigraphic section 3 (San Roque, see location in Figure 3) at the western flank of the Mercedes syncline (northern area 1 in Figure 1) with petrographic, palaeocurrents and palynological results.
system, forming a sharp contact with the underlying La Luna formation and a valley morphology (Figure S3). Packstone and wackestone beds with fragments of bivalves (e.g. oysters) and gastropods interbedded with phosphatic sandstones are present at lower and middle stratigraphic positions in the northern area. Fabre (1981) also reported the presence of bivalves and siliceous sponges at the base of this unit. The top of the formation includes subtabular thin to medium beds of bioturbated quartzarenite with glauconite (Section 2), or medium beds of quartzarenite with through cross stratification and wavy contacts (Section 3), forming a gradational contact with the lower interval of the Guaduas formation (Figure S3).

Petrographic analysis of one sample collected in Section 3 indicates the presence of monocrystalline quartz with straight extinction (91%) and undulose extinction (3.6%), authigenic glauconite (2%) and muddy matrix (Figure 13). Detrital zircon age populations reported in two samples (n = 204) are >485 Ma (Figure 5).

Palynological analysis of four samples in Section 1 of the northern area and one sample in the southern La Conquista area (eastern block) yielded a palynological assemblage dominated by dinoflagellates Trithyrodinium fragile, Senegalinium laevigatum, Senegalinium bicavatum, Polykrikos spp., Phelodinium tricuspe, Andalusiella-Paleocystodinium complex, Odontochitina spp. and Volkinitidinium spp. (Figure S10: plate 16–17, 19–23), the spore Ariadnasporites sp. (Figure S10: plate 1) and the pollen Proxapertites humbertoides, which indicate a Campanian–Maastrichtian age (Guerrero & Sarmiento, 1996; Jaramillo & Rueda, 2004). Foraminifera linings, pseudo-amorphous and amorphous organic matter, marine algae and woody phytoclasts dominate the samples.

Lithological association, the presence of authigenic glauconite, and marine macro and microfossils indicate accumulation in middle to lower shoreface environments under oxic conditions and adjacent to a distal coastline that supplied fine-grained terrigenous material.

### 4.2.3 Guaduas formation

Three informal intervals were identified for this formation in the Mercedes syncline (area 1; Figure S3). Stratigraphic thickness of the lower interval measured in the eastern flank of the Mercedes syncline (area 1) ranges between 89 m in Section 1 and 147 m in Section 2, whereas the thickness at the western flank is 160 m (Figure 3). The middle and upper intervals were mapped as a single unit with stratigraphic thickness ranging from 257 to 300 m in the eastern flank (Sections 1 and 2, respectively) and 260 m in the western flank (Section 3). Stratigraphic thicknesses of the Guaduas formation were not possible to calculate in areas 2 and 3 because the middle interval behaves as a detachment level.

### Lower interval

This interval includes upward-coarsening and upward-finining successions. The former succession consists of black shales, thin to medium bedded siltstones with internal wavy and planar lamination, and very fine to fine-grained sandstones to the top that either have meter-scale wave dune geometries or are internally massive due to intense bioturbation (Figures 5 and 6). The gradational nature of the lower contact with the underlying Pinos formation and the upper contact with the middle member of the Guaduas formation generates a ridge morphology bounded by two valleys (Figure S3). Upward-finining successions were identified in the northern area 1 (Section 1, Figure 5) and in the eastern block of the southern area 3 (Figure S5). They consist of medium-grained quartzarenite with fragments of bivalves at the base, wavy and planar cross stratification and bioturbation levels at the top. These sandy beds are interbedded with banks of black shales and thin limestone beds. Ferruginous nodules are present at the top of this unit. Fabre (1981) reported a conglomeratic level with angular quartz fragments, reworked mudstone and fossil fragments in the northern area.

Palynological analysis of four samples indicate that sandstones are quartzarenite with straight monocrystalline quartz (up to 89%), undulose monocrystalline quartz (<6%) and a trace of chert (<3%; Figure 8). Other framework grains are unstable sedimentary lithic fragments (<6%, muddy fragments), micas (<1%) and reworked glauconite (<5%; Figure 9a). Fabre (1981) also reported glauconite as grains (3%) and as part of the matrix, whereas isolated foraminifera shells and oyster fragments were identified in Section 1. Interstitial material consists of clayey matrix and ferruginous and calcite cement.

Detrital zircon age populations from two samples (n = 197) are >1.1 Ga, missing zircons of 485–1100 Ma, which are present in the other analysed samples (Figure 5). Heavy minerals separated in one sample consist of 18.6% ultrastable minerals, 78.4% clinzoisite and zoisite, 3% apatite and titanite, and muscovite (Figure 8). Two palaeocurrent indicators measured in the southern area 3 yielded an eastward flow direction (Figure S9).

Palynological analysis of four samples in Section 1 of the northern area revealed continental palynomorphs (Psilatrites guadensis, Annutriporites iversenii), woody phytoclasts, fungal remnants, pollen and spores, foraminifera linings and dinoflagellates. A late Maastrichtian age is assigned to the lower interval by its stratigraphic position (Figure 5).

The vertical and lateral mixing of upward-coarsening and upward-finining successions, together with the presence of macro and microfossils of continental and marine origin indicate accumulation in mixed muddy and sandy upper shoreface conditions (e.g. Sections 2 and 3) adjacent to extensive coastal plains with swamps cut by fluvial-estuarine channels.
Middle interval

This unit consists of thick intervals (10–20 mts) of carbonaceous-rich siltstones and shales that grade up-section to wavy to ripple laminated fine-grained sandstones with poor bioturbation and laminae enriched with organic matter. The gradational contact with the lower interval generates a valley morphology (Figure S3). In Section 2, this interval includes upward-fining medium- to fine-grained sandstone successions with medium sets of cross stratification (Figure S7). Sandstone beds show a bimodal distribution of fine and medium sand size. These sandstones are interbedded with banks of light-coloured mudstones with siderite nodules and thin to medium beds of coal. In Section 1, heterolithic interbeds of mudstones and siltstones are present towards the top (Figure 5). Thin lenses of coal and plant remains were reported by Fabre (1981) towards the top of this unit.

Petrographic analysis of four samples indicates that these sandstones are quartzarenites with straight monocrystalline quartz (up to 93%) and undulose monocrystalline quartz (<4.6%; Figure 8). Polycrystalline quartz fragments include sedimentary (<1.3%), diffusive (<4%) and chert (<3%; Figure 9b). Other framework grains are unstable sedimentary lithic fragments (<3%), micas (<6%), reworked glauconite (<5%) and a trace of silicified foraminifera in the northern section. Interstitial material consists of clayey matrix with ferruginous and calcareous cement.

Detrital zircon age populations from one sample (n = 100) are 99% > 485 Ma, and one zircon yielded an age of 80.4 ± 3.7 Ma (Figure 5). Heavy minerals separated in one sample consist of 44.7% ultrastable minerals, 43.5% clinozoisite and zoisite, 3.5% titanite, 8.2% amphibole, and muscovite (Figure 8).

In northern area 1, flow direction is to the west for six sets of cross beds and to the east for only 1 (Figure S7), whereas a current ripple measurement indicates a northward flow direction (Figure 6). In the central area 2, a metric-scale cross stratification set grades downdip from very fine-grained sandstone to siltstone and has a NNE dip direction; this bedform corresponds to a mouth bar. In the southern area 3, palaeocurrents show an eastward orientation (Figure S9).

Palynological analysis of five samples in Section 1 of the northern area indicates an early Palaeocene age. We analysed three samples from Sections 2 and 3 of the northern area 1. Only one sample yielded palynomorphs including *Proxapertites humbertoides* and *Spinizonocolpites baculatus*, which suggest a Maastrichtian to early Palaeocene (Guerrero & Sarmiento, 1996; Jaramillo et al., 2011; Figure S10; plate 7). In six samples analysed in central area 2 and southern area 3, only three samples yielded palynomorphs; these included *Spinizonocolpites baculatus* and dinoflagellate *Cerodinium* spp. (Figure S10; plate 18) suggesting an early Palaeocene age (biozone T01). Therefore, analysed samples yielded continental palynomorphs, woody phytoclasts, fungal spores, and in minor proportion, foraminifera linings and dinoflagellates.

Dominance of fine-grained successions grading up-section to wavy laminated sandstones, the presence of carbonaceous material, the large variation of palaeocurrents (W, N, E), the identification of mouth bars and the dominance of continental micropalaeontological association indicate the progradation of muddy coastal plains with frequent variation in flow regime (heterolithic laminations and bimodal grain size distribution) and extensive development of swamps with short periods of marine incursions. Shallow fluvial channels cut the coastal plains and end in the extensive swamps/lake systems.

Upper interval

We made a careful field control of the lower and upper contacts of this interval. At its base, thick fine- to medium-grained sandstone beds with trough and planar cross stratification rest in a sharp contact upon black laminated mudstones of the middle interval in all three areas (Figure 4). In central area 2 and northern area 1, sandy units within the upper interval grade up-section to interbeds of fine-grained sandstones and mudstones with bedforms showing wavy contacts (Figure 4b) and heterolithic laminations (ripples, wavy and lenticular laminations). The upper contact shows evidence of truncation and thick beds of the Socha Sandstone formation rest upon the irregular contact (Figure 4). This interval is <30 m thick but generates a gradual change in slope geometry with respect to the middle interval of the Guaduas formation (Figure 4a). In Section 5 (southern area 3), a 10-m-thick upward-fining sandstone bed is overlain by a 40 m thick succession of massive, light-coloured mudstones and medium-bedded lenticular siltstone interbeds (Figure 4c, Figure S9). Fabre (1981) included this upper interval as part of the Socha Sandstone formation because of the change in the topography, but we consider that the fine-grained interval and sedimentary structures of the sandstones are more characteristic of the Guaduas formation than of the Socha Sandstone formation, as described below.

Petrographic analysis of five samples indicates that these sandstones are quartzarenites with straight monocrystalline quartz (up to 92%); undulose monocrystalline quartz is present in all the samples (<6.7%; Figure 8). Polycrystalline quartz fragments include foliated (<0.7%), diffusive (<2%) and chert (<0.7%). The altered feldspar fraction is <1.5%. Other framework grains are unstable sedimentary lithic fragments (<3%), metamorphic (<1.3%), volcanic (<0.7%), micas (<1.7%) and reworked glauconite (<4.6%; Figure 9c). Interstitial material consists of clayey matrix and ferruginous cement. Directions of 14 palaeocurrents in central area 2 are scattered, but with a regional northward orientation (Figure S8).

Palynological analysis of six samples indicates the association of *Monocolpopollenites ovatus*, *Echitripores*
4.2.4 | Socha Sandstone formation

This sandstone-dominated unit offers good exposures in the three studied areas, and the irregular and sharp contact with the upper sandy interval of the Guaduas formation shows evidence of truncation of tens of meters in all studied sections (Figure 4). Measured thickness is 245–260 m in the Mercedes syncline (Figure 3) and in the southern La Conquista area (Figure S6). This unit consists of cross-stratified, medium-grained sandstones grading up-section to cross-stratified and massive fine-grained sandstones. In the middle of the Socha Sandstone, fining-upward successions include pebble-size fragments in the cross-stratified beds, these being the coarsest fragments observed in the studied succession. Thickness of cross stratification sets is medium to thick (<1 m). Light-coloured silty mudstones and thin coal beds are present as thin to thick layers interbedded within the sandy successions.

Petrographic analysis of six samples indicates that these sandstones are quartzarenites with straight monocrystalline quartz (up to 91%) and undulose monocrystalline quartz (<4.6%; Figure 8). Polycrystalline quartz fragments include sedimentary (<1.3%), foliated (<1%; Figure 9d), diffusive (<6.3%) and chert (up to 4%). The altered feldspar fraction is <1%. Other framework grains are unstable sedimentary lithic fragments (<7.6%), metamorphic (up to 1.6%), volcanic (<0.7%) and micas (<1.3%). Fabre (1981) indicated that chert fragments composed up to 5% of the sandstone composition at the top. Interstitial material consists of clayey matrix and ferruginous cement.

Detrital zircon age populations from one sample (n = 105) indicate that 94% of the zircons are >485 Ma, while six grains have younger ages: 328.1, 236.6, 82.1, 77.5, 77.1 and 64.6 Ma (Figure 5; see Table S3 for age errors). Heavy minerals separated in one sample consist of 35.9% ultrastable minerals, 59.8% clinozoisite and zoisite, 4.3% apatite and titanite, and muscovite (Figure 8).

Palaeocurrents from the lower segment in the southern and central areas have dominant NNE flow directions (16 measurements; Figures S8 and S9), three palaeocurrents in the central area have an ESE direction, and one palaeooccurt a WSW direction (Figure S8). In the northern area, flow directions from the lower segment show two nearly opposite trends, ENE (n = 6) and WSW (n = 5; Figure 6 and Figure S7). In the southern and central areas, palaeocurrents in the upper segment (the interval with pebble and granule size quartzose fragments) have a westward flow (n = 10), one direction to the NW and another to the SW (Figures S8 and S9), whereas in the northern area 13 palaeocurrents have a direction to the NEE (Section 3, Figure 6) and three palaeocurrents have an almost orthogonal WNW direction (Section 2, Figure S7).

Palynological analysis of one sample in Section 4 yielded Echitriporites suescae, Foveotremites margaritae, Psilabrevitricolporites annulatus and Syncolporites lisa-mae, suggesting a Maastrichtian–Palaeocene age (biozones T01 to T02; Figure S10; plates 5, 9, 11); a similar association is reported for the Guaduas formation by Sarmiento (1992). One sample at the top of this unit in Section 5 contains Cricotriporites minutiporus, Syncolporites marginatus and Ulmoideipites krempii, which suggests an age no younger than middle Eocene (biozone T06; Figure S10: plates 4, 12, 15). For this unit, a mid-Palaeocene age is inferred from its stratigraphic position and the youngest zircon age. These samples show a dominance of continental palynomorphs, woody phytoclasts, cuticles and fungal spores.

The dominance of cross-stratified sandstones lithofacies with thin interbeds of coal and light-coloured mudstones, the large range in orientation of flow directions (SW to NNE) and the continental micropalaeontological associations indicate accumulation of fluvial channel sand bars. Lateral channel migration, as recorded by amalgamation of sand bars of different fluvial fills, left poor record of adjacent flood plains. Opposite palaeoflow directions in lower beds may indicate minor tidal influence affecting only the northern area.

4.2.5 | Socha Mudstone formation

In the Mercedes syncline in the northern area (Figure 3), stratigraphic thickness calculated in the cross section increases from 341 m in the eastern flank to 635 m in the western flank. In the southern area, stratigraphic thickness is 410 m in Section 5 (Figure S6). This unit consists at its base of meter-scale successions of very fine- to fine-grained lithic-bearing massive to planar laminated sandstones that grade up-section to thick successions (<20 m thick) of...
massive to wavy-ripple laminated light-coloured mudstones and siltstones, with local preservation of leaves and thin beds of coal. Massive mudstones, peds, reddish to brownish alteration and calcite cements in sandstones (Figure 9f) indicate pedogenic processes. Upward-coarsening successions of fine- to medium-grained sandstones with siltstone and thin coal beds at the base and thick sets of cross-stratified beds at the top are common in the lower and middle intervals. The uppermost beds consist of >20 m thick light-coloured massive mudstone intervals (Figure 7).

**FIGURE 7** (a) Stratigraphic correlation of Maastrichtian to lower Palaeocene strata. (b) Stratigraphic correlation of upper Palaeocene to lower Eocene strata. Accommodation rate calculation used biostratigraphy, geochronology and thickness data from the Arteza section 1 showing an up-section decrease from Maastrichtian to earliest Eocene, and an increment in early Eocene. Supply of sandy terrigenous detritus increased in middle to late Palaeocene, but important supply of muddy terrigenous detritus in Maastrichtian and early Eocene favoured the filling of the sedimentary basin to keep the accumulation in continental basins.
Petrographic analysis of 23 samples collected in the three areas indicate sublitharenite to litharenite composition, with a significant increase up-section of total unstable lithic fragments (up to 41%; Figure 8). Monocrystalline quartz has straight (up to 86%) and undulose extinction (<8.3%). Polycrystalline quartz fragments include sedimentary (<7%), foliated (<1.3%), diffusive (<5.3%) and chert (<13%). The altered feldspar fraction is <2%. Unstable lithic fragments (Figure 9e and f) are sedimentary (<26.6%), metamorphic (<18.6%) and volcanic (<20.3%). Muscovite is also a very common fragment in this unit (<11%). Interstitial material consists of clayey matrix and ferruginous and calcareous cement. Fabre (1981) also documented the highly immature composition in this unit with quartz in the range of 30%–40%.

Detrital zircon age populations in one sample (n = 108) have 73 grains >482 Ma, eight grains in the range of 223–477 Ma, nine grains of 135–154 Ma, six grains of 75–97 Ma and 12 grains of 51–59 Ma (Figure 5; see Table S3 for age errors). Heavy minerals separated in two samples at the base consist of 9.5%–30.5% ultrastable minerals, 55.1%–74.3% clinozoisite and zoisite, 2.9%–4.6% apatite and 1.7%–7% titanite. Garnet is present in the northern area (4.4%), whereas the sample of the southern area includes epidote (7.1%), amphibole (1.7%) and biotite (1.2%; Table S2). Muscovite is present in both samples.

A total of 68 flow directions collected in Section 3 (Figure 6) and 16 flow directions collected in Section 5 (Figure S9) show a regionally high dispersion for the lower sandstones, even with some directions showing a southward orientation in both areas. In contrast, palaeocurrents in lower Eocene strata show a dominant NNE and E flow orientation in the southern area (Figure S9) and NW to NE (total of 9) and EN to ES (19) flow orientation in the northern area.

Palynological analysis of four samples yielded Proxapertites operculatus, Retidiportes magdalenensis, Bombacacidites brevis, Striatopollis catatum-bus, Proxapertites psilatus, Spatiphyllum vanegensis, Gymmonocolpites barbatus and Psilamonocolpites spp., indicating an early Palaeocene to early Eocene age (biozones T01 to T04; Figure S10: plates 8, 10, 13). We infer a late Palaeocene age at the base and an early Eocene age at the top. These samples show a dominance of continental palynomorphs, woody phytoclasts, cuticles and fungal spores.

FIGURE 8 Compositional ternary diagrams of Folk (1974) showing the dominance of quartzarenite composition in the Guaduas and Socha Sandstone formations, and sublitharenite to litharenite in the Socha Mudstone formation. Results of heavy mineral analysis in selected samples show the up-section decrease in ultrastable minerals and the presence of unstable minerals at different stratigraphic levels.
The dominance of fine-grained lithologies, the upward-fin-
ing successions ending in ripple and wavy laminated siltstones and mudstone beds, the thick interbeds of palaeosol profiles and the continental micropalaeontological associations indicate accumulation on flood plains dominated by palms and gallery forest and cut by low-energy fluvial channels; the resulting sandy bars and silty channel fills, along with high channel stability, allowed the preservation of the adjacent floodplains and associated crevasse splays. Meter-scale upward-coarsening successions are interpreted as the advance of fluvial systems over lacustrine settings. Fluvial systems in the middle interval, where grain size reached medium-grained sandstones, show a slight increase in fluvial flow energy.

5 | DISCUSSION

5.1 | Identification of stratigraphic units and stratigraphic surfaces of correlation

We identified two types of unconformities. A xenocon-
formity, as defined by Carroll (2017), is identified at La
Luna-Los Pinos formation boundary because of the chemical variation of seawater conditions. This xenoconformity is indicated by an abrupt change from chemical precipitation of chert and calcareous mudstone of La Luna deposits with abundant planktonic forams (Fabre, 1981) to detrital accumulation in a fine-grained clastic-dominated system interbedded with fossiliferous limestone beds at the base of Los Pinos formation (Figure 5 and Figure S3). This xenoconformity is well documented to the north and west of the Cocuy basin. To the north, in the Maracaibo, Catatumbo and other basins to the west, this unconformity corresponds to the La Luna-Colon formations contact, where foraminifera assemblages from the uppermost La Luna formation are indicative of dysoxic marine environments with restricted circulation, whereas those in the lower Colon formation indicate oxygenated conditions (Martínez & Hernandez, 1992; Patarroyo, Torres, Rincón, Cárdenas, & Márquez, 2017). Glauconitic limestones at the base of the Colon formation have been interpreted as a condensed section of the lower and middle Campanian in the Cesar basin (Martínez & Hernandez, 1992). The same time interval is missing in the Catatumbo basin (Patarroyo et al., 2017). In the Middle Magdalena basin and western foothills of the Eastern Cordillera, the contact between the La Luna and Umir formations has been interpreted as an unconformity (Ward et al., 1973); the presence of a 1-m-thick glauconitic packstone above this boundary has been interpreted as a condensed section (Quiroz, Romero, & Delgado, 2018). This unconformity correlates with conformable accumulation of fine-grained siliciclastic strata of the upper Chipaque and the lower Guadalupe formations in the Llanos foothills (Guerrero & Sarmiento, 1996).

The second type of unconformity consists of disconformable contacts, and they were documented at the base of the Guaduas formation, at the base of the upper interval of the Guaduas formation, and at the base of the Socha Sandstone formation (Figures 4 and 7). These contacts are the result of enrich in sandy terrigenous fragments and northward migration of continental depositional systems (Figure 7). The truncation is produced by dilute streamflow incision that does not appear to have eroded stratigraphic thickness enough to have caused the absence of a complete palynological biozone. The abrupt input of siliciclastic sandy detritus in very shallow marine environments, likely close to the mouth of a riverine system (Figure 10), may explain the local disconformity at the base of the lower member of the Guaduas formation in Sections 1 and 5, whereas the progradational advance of the coastal plains of the Guaduas formation explains the transitional nature of the lower contact of the lower member in Sections 2 and 3 (Figure 7). The disconformity at the base of the upper interval of the Guaduas formation is defined by an abrupt increase in quartzose sandy grains and development of sandy dunes with wavy contacts (Figure 4b), which together with bimodal grain-size distribution and palynological-foraminiferal association, support the interpretation of coastal-plain conditions for the upper interval with tidal influence in the northern and central areas (Figure 10).

The disconformity at the base of the Socha Sandstone formation shows evidence of meter-scale truncation associated with riverine fluid incision. This unconformity is regional, and it has been documented at the base of the Palaeocene clastic wedges C2 and C3 shown in Figure 1 (see references in Bayona, 2018). In the study area, as well as in the southern and central part of the clastic wedge C2, strata of the Socha Sandstone formation are dominated by amalgamated channel-fill deposits that accumulated on continental fluvial plains with poor preservation of flood plain deposits, making age determination difficult (Figures 10 and 11). This disconformity does not correspond to a long period of incision and sediment bypass; instead, it is a diachronous surface evolving by erosion and deposition. Amalgamated channel-fill deposits resting upon the disconformity are coeval with the disconformity surface generated elsewhere by migration of the channel system (see examples of channel-belt migration above of basal valley-fill surfaces in Blum, Martin, Milliken, & Garvin, 2013).

### 5.2 Stratigraphic variation of flow directions

Variability in palaeoflow directions of the Guaduas, Socha Sandstone and Socha Mudstone formations indicates that fluvial systems did not have a regional flow trend in the Palaeocene (Figures 10 and 11). In the Guaduas formation, flow direction is mostly northward, but with some directions towards the west or the east. Lower beds of the Socha Sandstone in the northern area 1 show opposite directions with dominance to the ENE, more likely related to minor tidal influence affecting only the river-dominated channel system (Kurcinka, Dalrymple, & Gugliotta, 2018), whereas upper beds have NE (Figure 6) or NW directions (Figure S7). In the central and southern areas, NNE directions dominate in lower beds, whereas in upper beds of the Socha Sandstone palaeoflow directions are mainly westward (Figures S8 and S9). Grain size increase in the middle and upper intervals of the Socha Sandstone; however, the increase in the flow regime did not result in more uniform palaeoflow patterns. For the lower segment of the Socha Mudstone, palaeoflow directions are also variable, including directions to the WSW and EWE, whereas palaeoflow directions to the top of the Socha Mudstone are more uniformly to the ENE (Figure 6 and Figure S9). We interpret that Palaeocene river systems export sediments to local river mouths leading into closed
FIGURE 10  (a–c) Local palaeogeographic maps with ternary provenance diagrams and palaecurrent domains from late Maastrichtian to middle Palaeocene, showing the change from a marginal basin with a western source area to fluvial systems ending into local lakes (closed basin systems) bounded by two source areas with different composition of the sedimentary cover. (d) Regional palaeogeographic map of the northern Andes for Maastrichtian showing the location of the Cocuy region (base map from Montes et al., 2019 and Bayona, 2018)
FIGURE 11  (a–c) Local palaeogeographic maps with ternary provenance diagrams and palaecurrent domains from middle–late Palaeocene to early Eocene, showing the filling of closed basin systems bounded by two source areas with different composition of the sedimentary cover and normal faulting in the early Palaeocene. (d) Regional palaeogeographic map of the northern Andes for late Palaeocene showing the location of the Cocuy region (base map from Montes et al., 2019 and palaeogeography from Bayona, 2018). Intra-basinal uplifts separate the three clastic wedges identified in Figure 1: C1: Seca-Hoyón-Lisama; C2: Bogota–Socha Mudstone; C3: Barco-Los Cuervos
lacustrine systems, as discussed below. Farther to the north, there are not Palaeocene sandy deltas reported along the ma-
rine coastline; instead, mixed calcareous rocks indicate poor
supply of clastic detritus from the continent (Bayona, 2018).

5.3 | Regional comparison of sandstone composition: inferring location and composition of source areas

A change in sandstone composition, consisting of an up-sec-
tion increase in unstable fragments, is notable in Palaeocene
units. This occurs in the northern segments of clastic wedges
C1 (Middle Magdalena and western foothills) and C2
(Florestra and Cocuy), as well as in the central C3 (Medina).
Sandstone provenance for lower Palaeocene units (GuaduAS
and Socha Sandstone formations) indicates a reworked inte-
rior craton to quartzose recycled orogen (Figures 10 and 11),
suggesting no major provenance change across the discon-
formities. Source areas likely included a sedimentary cover
with quartzose units rich in glauconite and chert interbeds.
Oxidized mud fragments may result from soil development
in these source areas. Sandstones of the upper Palaeocene-
lower Eocene Socha Mudstone are sublitharenite to litharen-
ite, with up to 5% unstable heavy minerals and an up-section
increase in metamorphic rock fragments. Polycrystalline
quartzose and chert fragments increase up-section, as well
as unstable lithic fragments of volcanic, metamorphic (phyl-
lite and schist) and sedimentary (siltstone and claystone) ori-
gin (Figure 9e,f). Sandstone provenance diagrams indicate a
quartzose to transitional recycled orogen (Figures 10 and 11).
Similar up-section increase in unstable lithic fragments is re-
ported in Palaeocene sandstones of the Catatumbo (Ayala-
Calvo et al., 2012), Floresta (Saylor et al., 2011) and Medina
(Bayona et al., 2012; Odoh et al., 2019) basins, whereas
Palaeocene litharenites in the Magdalena basins always in-
clude immature sedimentary, metamorphic and volcanic rock
fragments (Caballero, Mora, et al., 2013; Moreno et al., 2011;
Naranjo et al., 2013). Caballero, Mora, et al. (2013) indicated
that clasts of sedimentary origin are mainly of claystone, silt-
stone, sandstone, chert and glauconitic sandstones suggesting
erosion of Cretaceous units from the western foothills of the
Eastern Cordillera.

In the northern termination of the Nuevo Mundo syncline,
Prince et al. (2016) argued that the presence of conglomerate
clasts and feldspars in lower sandstones of the Lisama forma-
tion indicates supply from the Santander massif. Growth strata
affecting the top of the Lisama formation in the Acordionero
field (see location in Figure 1) suggest eastward-dipping in-
verse structures that support this interpretation.

The sandy to pebble-size quartzose terrigenous sediments
in lower to middle Palaeocene fluvial sandstones and imma-
ture detritus in upper Palaeocene sandstones in the Cocuy,
Floresta and Catatumbo basins, were likely derived from
nearby uplifts that bounded the accumulation of clastic wedge
C2, rather than from distal sources in the Central Cordillera.
Amorocho, Bayona, and Reyes (2011) found that unstable
rock fragments and glauconite in fluvial sands of the tropi-
cal Llanos basin do not persist during fluvial transportation
for more than 100 km from their source area in the Eastern
Cordillera. The high levels of precipitation and tempera-
ture of the Palaeocene/Eocene in northern South America
(Jaramillo & Cardenas, 2013), which would have enhanced
weathering, would also suggest that unstable fragments were
derived from nearby sources.

Supply of unstable terrigenous fragments from nearby
sources explains the presence of (a) reworked glauconite grains,
(b) chemically and mechanically unstable fragments (feldspars,
micaceous and graphitic metamorphic grains, amphiboles, gar-
net, biotite) and (c) oxidized muddy matrix as result of rework-
ing soil development in source areas (Figure 9). Reactivation
of nearby faults along the boundary of the Cocuy graben and
Santander massif (Figure 2a) caused the deformation of muddy-
dominated Cretaceous and sandy-dominated Jurassic rocks.
These nearby western uplifts correspond to palinspastically
restored segments of the southern Santander massif and Los
Cobardes Anticline (Reyes-Harker et al., 2015), where ther-
mochronological data indicate onset of exhumation since the
Palaeocene (Caballero, Mora, et al., 2013; Velandia, 2018). The
increase in immature rocks fragments in the Socha Mudstone
formation indicates the exposure and erosion of clastic units of
the Upper Jurassic Giron formation, a unit found within and to
the west of the Santander massif (Figure 11).

Reactivation of other crustal faults of the Cocuy graben
(Figure 2a) caused subtle deformation of Upper Cretaceous
quartzose sandstones with glauconite resting upon inter-
beds of chert and black mudstones of La Luna formation
(Figures 10 and 11). The up-section increase in quartzose
sandy detritus, up to pebble size in the Socha Sandstone, may
have been supplied from those eastern source areas. These
nearby eastern uplifts correspond to the tectonic loads re-
quired to generate the foreland basin recorded by the C3 clas-
tic wedge (Bayona et al., 2008, 2009).

5.4 | Regional comparison of U-Pb detrital zircon populations, inferring lateral connectivity of clastic systems

Detrital zircon populations reported in Maastrichtian to
Lower Palaeocene sandstones show 100% of ages >485 Ma
for samples from La Luna, Los Pinos and lower-middle
GuaduAS units in the Cocuy Basin (Figures 5 and 12b), and
only one age of 80 Ma in the sample of the middle GuaduAS
formation. Ages >485 Ma are also almost all in coeval strata
in the Floresta and Medina basins (Table 1 and Figure 12c;
Bayona et al., 2012; Odoh et al., 2019).
Saylor et al., 2011; Bayona et al., 2013; Odoh et al., 2019). Unroofing of the Cretaceous sedimentary cover is regionally documented by the dominance of detrital zircons >485 Ma, but the up-section increase in detrital zircons <485 Ma in Palaeocene sandstones requires a detailed geochronological provenance analysis. The following paragraphs describe in detail the possible source for each zircon population within the Cocuy basin and its comparison with proposed source areas in adjacent basins.

The internal subdivision of detrital zircons <100 Ma of the clastic wedge C1 suggests there was no internal north–south connection and that western sources may have been tectonically transported northward. In the northern Middle Magdalena basin, the detrital zircon of 66–100 Ma constitutes <2% (Table 1), whereas this population is 40% in the southern segment of the basin (Bayona et al., 2013). Additionally, Palaeocene zircons are absent in the northern segment, whereas Palaeocene ages have been reported in the southern segment. Upper Cretaceous and Palaeocene batholiths, widely exposed in the northern Central Cordillera (Figure 1), restore palinspastically to the west of the southern segment (Figure 14). The separation of the northern and southern segments of the C1 clastic wedge is further supported by Palaeocene deformation proposed by Restrepo-Pace et al. (2004) in the boundary of these two segments. Horton et al. (2015) proposed that absence of Palaeocene zircons in the northern segment of C1 may be related to limited contribution of magmatic centres in the southern segment of C1.

The internal subdivision of 66–201 Ma detrital zircons of the northern clastic wedges C1 and C2 (Table 1; Figures 12 and 13) suggests that the Cocuy area (northern C2) was laterally connected with fluvial systems of the southern C2 and disconnected from fluvial systems of the northern C1 (Figure 14). In the northern clastic wedge C2, the 66–100 Ma population decreases southward from Catatumbo (8.3%) to Floresta (4.2%); this population increases to ~20% in the southern segment of clastic wedge C2 (Bayona et al., 2013), whereas it is <2% in the northern C1. In the northern clastic wedges C1 and C2, the population of 100–164 Ma is less

| Sample | Unit | 49–66 Ma | 66–100 Ma | 100–164 Ma | 164–201 Ma | 201–485 Ma | >485 Ma |
|--------|------|----------|-----------|------------|------------|------------|---------|
| U821_N = 88 | Upper Lisama | 0 | 0 | 9.1 | 1.1 | 44.3 | 45.5 |
| LM1505097 + RS0114091 N = 157 | Lower Lisama | 0.6 | 1.9 | 0.6 | 1.3 | 5.1 | 90.4 |
| Cocuyo 1–7 N = 16 | Umir | 0 | 0 | 18.75 | 18.75 | 12.5 | 90.4 |
| PE–05 + PE06+RC–01 N = 254 | Los Cuervos | 2.4 | 8.3 | 3.5 | 2.4 | 7.5 | 76 |
| BA–01 N = 90 | Barco | 0 | 7.8 | 1.1 | 0 | 4.4 | 86.7 |
| U–70 N = 108 | Socha Mudstone | 11.1 | 5.6 | 8.3 | 0 | 7.4 | 67.6 |
| U–37 N = 105 | Socha Sandstone | 1 | 2.9 | 0 | 0 | 1.9 | 94.2 |
| U–38 N = 100 | Middle Guaduas | 0 | 1 | 0 | 0 | 0 | 99 |
| U–11 + U–39 N = 197 | Lower Guaduas | 0 | 0 | 0 | 0 | 0 | 100 |
| 3BUG228 + 4BUG9+110808–16 + SOG–10 N = 384 | Socha Mudstone | 9.6 | 4.2 | 5.7 | 1.6 | 2.6 | 76.3 |
| 110808–4 + SOG–09 N = 178 | Socha Sandstone | 0 | 1.7 | 0 | 0 | 0.6 | 97.8 |
| SOG–08 + 2BUG190 N = 181 | Guaduas | 0 | 0 | 0 | 0 | 0 | 100 |
| 1,108,086 N = 82 | Tierna | 0 | 0 | 0 | 0 | 0 | 100 |
| MEDINA (central clastic wedge 3) |
| GJ611 + NA46+GUA–01 N = 267 | Los Cuervos | 12.7 | 2.6 | 2.6 | 1.1 | 1.9 | 79 |
| GJ604 + GUA–03 N = 308 | Barco | 0 | 0.6 | 0 | 0 | 1 | 98.4 |
| GUA–02 N = 272 | Guaduas | 0 | 0 | 0 | 0 | 0.4 | 99.6 |
| GJ607 N = 100 | Tierna | 0 | 0 | 0 | 0 | 0 | 100 |

Note: See Figure 12 for the illustration of these populations.

, Maastrichtian-lower Palaeocene; □, Lower-Middle Palaeocene; ■, Upper Palaeocene.
Comparison of detrital zircon population from Maastrichtian to upper Palaeocene–lower Eocene units in: (a) the Magdalena depocenter (clastic wedge C1 in Figure 1), (b) Catatumbo and Cocuy basins (northern clastic wedge C2 in Figure 1) and (c) Floresta basin (central clastic wedge C2 in Figure 1) and Medina basin (central clastic wedge C3 in Figure 1). The change in age scale illustrates zircon populations from 0 to 100 Ma, 100 to 500 Ma and older than 500 Ma. Note that (1) population in A are very different to (b and c), (2) in (b and c) increase up-section the 50–100 Ma population while decrease the >500 Ma population and (3) the major difference between (b) and (c) is the higher zircon populations of 100–500 Ma in (b) than in (c) because (b) is closer to the Santander massif.

MDS diagrams showing the distribution of detrital zircon populations reported in the clastic wedges C1 (western foothills), C2 (Catatumbo, Cocuy, Floresta) and C3 (Medina). (a) Lower to middle Palaeocene sandstones of the Guaduas formation plot separately, but the samples of the Barco-Socha Sandstone units tend to cluster (see arrows). The Lisama sample is part of the large dispersion from the other basins and the scattering may reflect supply from several uplifts with Cretaceous sedimentary cover. (b) Upper Palaeocene–Eocene components plotted totally apart as result of the supply of detrital zircons from the Central Cordillera (Caballero, Mora, et al., 2013; Nie et al., 2010, 2012), whereas detrital zircon population of Socha Mudstone and Los Cuervos samples indicate that other source areas supplied sediments to the Catatumbo, Cocuy, Floresta and Medina areas (see text for discussion).
than <9% without clear geographic distribution; this population, however, increases to 10% in the southern segment of clastic wedge C2 (Bayona et al., 2013). The population of 164–201 Ma is 4% in Palaeocene sandstones of the northern clastic wedges C1 and C2 (Table 1). In the southern segment of clastic wedge C2, sediments were supplied from the Cretaceous sedimentary cover of the Central Cordillera (Zapata et al., 2019; Figure 14), or from Upper Jurassic igneous rocks of the southeastern border of the Central Cordillera (Bustamante et al., 2017). However, reworking of the southern clastic wedges C1 and C2 by intra-basinal deformation may have supplied those 66–201 Ma detrital zircons into the Floresta to Cocuy basins (Figure 14). Alternatively, 66–201 Ma detrital zircons may be also supplied from the erosion of volcanic/plutonic rocks interbedded with the Cretaceous sedimentary cover of the Eastern Cordillera.

FIGURE 14  Late Palaeocene palaeogeography (palinspastic map from Montes et al., 2019) showing the location of the marine shoreline and the boundary where lacustrine systems were subaerial to the south and more flooded to the north. In the Cocuy area, sediments were supplied from nearby W and E uplifts, as well as by axial transport from the SW (66–485 Ma detrital zircons). Maximum thickness of Palaeocene strata is reported in C2. The wide of deformation between 4º and 6ºN is very similar, if we include the fault activity in the southern Llanos from Mora et al. (2013). The aerial extension of the clastic wedge C3 also includes the extension of Palaeocene volcanic tuffs in the Llanos basin (see text for discussion).
patterns related to introduction of new and proximal source areas, (b) stacking patterns that differ among sub-basins, generated by local tectonism and climate and (c) effects in local basins due to multiple non-uniform tectonic loads. To evaluate the hypothesis that northern clastic wedges C1 (Middle Magdalena and western foothills), C2 (Floresta and Cocuy) and C3 (northern Llanos foothills and basin; Figures 1, 2a and 14) were separated during Palaeocene-early Eocene time, we examine thickness variation, lithological associations, proposed depositional environments, palynological constraints, palaeocurrents, sandstone composition (Table 2; see references in the legend) and the geochronological provenance analysis discussed above.

Variations in thickness of Palaeocene strata among the three clastic wedges indicate that a different tectonic subsidence mechanism governed each clastic wedge. Table 2 shows the thickness values for Palaeocene-lower Eocene strata from stratigraphic sections adjacent to the regional cross section shown in Figure 1, and only the maximum values are shown in Figure 14. The maximum thickness of Palaeocene-lower Eocene strata reported in each clastic wedge varies from 800 m in C1 to 1,200 m in C2 (see Figure 6), to 600 m in C3 (Figure 14); this variation does not follow an eastward decrease in thickness, which would be expected in a continuous foreland basin. Additionally, thickness varies internally within each clastic wedge. In C1, thickness decreases towards active structures along the western and middle segments of C1 where Palaeocene strata had not accumulated (Parra et al., 2012; Caballero, Mora, et al., 2013; Reyes-Harker et al., 2015; see fold axes in Figure 14). The thickness also decreases northward in areas with active deformation (Prince et al., 2016). The eastward increase in thickness is explained by clockwise tilting of the Central Cordillera due to the subduction of the Caribbean plate (Bayona et al., 2011, 2013), whereas active deformation took place to the west and north in the C1 clastic wedge. In C2, the thickness of Palaeocene-lower Eocene succession in both the Cocuy and Floresta areas is similar (upper Guaduas, Socha Sandstone and Socha Mudstone formations; Table 2). As illustrated by Bayona et al. (2013) and Reyes-Harker et al. (2015), the Cocuy and Floresta basins are part of a single basin bounded to the west by the inversion of the Lower Cretaceous Tablazo graben (Figure 2a). In C3, the Barco-Los Cuervos succession thins eastward and forms an eastward-stepping stratigraphic package onlapping the basal Cenozoic unconformity in the proximal Llanos basin, as a result of flexural deformation associated with tectonic loads, the product of the reactivation of crustal faults located in the middle of the Cocuy graben (Bayona et al., 2008, 2009; see discussion above).

Stratigraphic stacking patterns of Palaeocene-lower Eocene strata differ among the three basins, also supporting the hypothesis that a different tectonic subsidence mechanism governed each basin. In C1, the contact between the Umir
### TABLE 2  Comparison of stratigraphic, sedimentological and provenance characteristics of the northern segments of clastic wedges C1, C2 and C3

| Stratigraphic unit (age); lower contact | Thickness and strata architecture | Major lithologies and sedimentary structures | Depositional environments | Palynological constrains on depositional systems | Palaeocurrents | Sandstone composition | Proposed source areas and provenance markers |
|---------------------------------------|----------------------------------|---------------------------------------------|---------------------------|-----------------------------------------------|---------------|---------------------|---------------------------------------------|
| Lower Lisama Fm. (lower to middle Palaeocene); gradational with Maastrichtian Umir formation | From north to south: Acordionero field = 450 m (4) to totally eroded (or non-deposition) in buried structural highs to the west (1); Northern Nuevo Mundo syncline (NMS)=125–200 m (3); western flank NMS = 800 m; eastern flank NMS = 885 m (but only 550 of Palaeocene age (1, 2)). | Dominance of dark-coloured mudstones, carbonaceous siltstone, coal ((1 to 8) with interbeds of upward-coarsening and fining successions, ripple, wavy, massive and cross-stratified very fine to medium sandstones. Local bioturbation and muddy intervals with pedogenic processes (3, 6). At the base, local conglomeratic beds in easternmost sections (3,6,7) | Wave-dominated delta with distributary mouth bars (5); Tide-dominated delta (1,2); intertidal sand flats and point bars (4) | Western flank NMS = westward; eastern flank NMS = northward (1,5) | Litharenites to sublitharenites (1,2,3,5) and feldspathic litharenites (3); variable content of metamorphic, sedimentary and volcanic rocks fragments | Upsection increase of metamorphic rocks fragments (2, 5). Proposed source areas: south Central cordillera (1), Cobardes anticline (1,7), Santander massif (3) |
| Upper Lisama Fm. (upper Palaeocene-lower Eocene); gradational with lower Lisama | Southern western foothills = 750 m (8). Tabular to subtabular geometries (6). Growth strata in upper Lisama beds (3) | Dominance of light-coloured mudstones with interbeds of massive, cross-bedded and heterolithic fine to medium-grained sandstones (3). Thick palaeosol successions at the top (6) | Subaerial delta plains (5); prograding delta (1); fluvial-estuarine (4) | Both flanks NMS = eastward (1,5), in Acordionero is interpreted westward flow directions | | |
| | | | | | | | |

(Continues)
### TABLE 2 (Continued)

| Stratigraphic unit (age); lower contact | Thickness and strata architecture | Major lithologies and sedimentary structures | Depositional environments | Palynological constraints on depositional systems | Palaeocurrents | Sandstone composition | Proposed source areas and provenance markers |
|----------------------------------------|----------------------------------|-----------------------------------------------|---------------------------|-----------------------------------------------|----------------|----------------------|---------------------------------------------|
| Northern clastic wedge C2. Cocuy basin (this study) and Floresta basin (9, 10) | | | | | | | |
| Socha Sandstone Fm. (middle Palaeocene); disconformable with lower Palaeocene Guaduas Fm. | Cocuy basin: 245–260 m, n. Floresta basin: 100–130 m (9,10). Amalgamated fluvial system | Cocuy = cross-stratified, massive fine to granule conglomeratic sandstones with thin interbeds of silty mudstones, coal. Floresta = medium-grained sandstone to granule conglomerate in tabular to lenticular bedsets, with common cross stratification | Cocuy = continental amalgamated fluvial channel belts forming sandy bars (moderate energy); Floresta = amalgamated, braided fluvial channels (9) | Cocuy = dominance of continental palynomorphs. Floresta = continental with occurrence of Proxaperites operculatus (10) | Cocuy = Scatter. Lower beds: dominance of NNE and ESE directions. Upper beds dominate W and NEE directions. Floresta: Northward and eastward directions (9) | Cocuy = Quartzarenite, fragments of chert and sedimentary lithic fragments. Floresta = quartzarenite to sublitharenite, with metamorphic lithic fragments being dominant over volcanic and sedimentary (9) | Cocuy = intrabasinal deformation, erosion of Kr sedimentary cover. Floresta = supply from the Central Cordillera (9) |
| Socha Mudstone Fm. (upper Palaeocene-lower Eocene); abrupt change from amalgamated to non-amalgamated fluvial system | Cocuy = Mercedes syncline: western flank = 635 m; eastern flank = 341 m, this change occurs in <3.5 kn. Floresta basin = 220–330 m, this change occurs in less than 2.5 kms (9). Mud-dominated units with non-amalgamated fluvial system | Cocuy = massive to wavy-ripple laminated light-coloured mudstones and siltstones; sandstone interbeds are very fine to medium-grained, massive with planar, ripple and cross bedded structures with local bioturbation. Floresta: lower interval includes dark-coloured mudstones, thin beds of fine sandstones and coal (9). Upper interval consists of thick palaeosol profiles interbedded with amalgamated sandstones (9,10), and presence of gypsum (10) | Cocuy = continental floodplains with isolated channel belts filled with low-energy energy and crevasse splay systems that end in lakes/swamps. Floresta: fresh water lakes in the lower segment (10) passing to subaerial floodplains (9, 10). Isolated sandstone beds are the record of crevasse splays, anastomosing fluvial systems (9) or meandering rivers (10) | Cocuy = dominance of continental palynomorphs. Floresta basin = presence of Pediastrum sp. and Ovoidites sp indicate fresh water lakes (10). Reworking of Cretaceous palynomorphs (10) | Cocuy = Scatter directions, including southward directions. Floresta: Northward and eastward directions (9) | Cocuy = Sublitharenite to litharenite with mixing of sedimentary, volcanic and metamorphic rock fragments. The latter become more abundant to the top (Figure 11). Floresta = sublitharenite, with dominance of metamorphic rocks fragments over sedimentary and volcanic fragments | Cocuy = intrabasinal deformation, erosion of Kr and Jr sedimentary cover. Floresta = change from craton to Central Cordillera sources (9) |

(Continues)
**TABLE 2**  (Continued)

| Stratigraphic unit (age); lower contact | Thickness and strata architecture | Major lithologies and sedimentary structures | Depositional environments | Palynological constrains on depositional systems | Palaeocurrents | Sandstone composition | Proposed source areas and provenance markers |
|----------------------------------------|----------------------------------|---------------------------------------------|--------------------------|---------------------------------------------|---------------|---------------------|---------------------------------------------|
| Northern clastic wedge C3. Northern Llanos foothills and proximal Llanos basin (11, 12) |                                  |                                             |                          |                                             |               |                     |                                             |
| Barco Fm. (lower to middle Palaeocene in the foothills, and upper Palaeocene in the Llanos basin); unconformable with Palaeocene and older strata eastward | Barco-Los Cuervos succession thins eastward and forms an eastward-stepping stratal package onlapping the basal Cenozoic unconformity in the western Llanos basin. In the Llanos foothills, Barco formation is 100 m thick and unconformably overlies fine-grained strata of the Colon-Mito Juan formation. The Cuervos is ca 500 m thick. In the proximal Llanos, the Palaeocene succession is mostly sandy and <200 m thick. Normal faults affect Cuervos Fm in the central Llanos foothills (13) | Foothills: bidirectional cross-bedded and bioturbated heterolithic laminated sandstones at the top with organic rich mudstones and coal beds. In the Llanos basin: aggraded fine-to-coarse-grained sandstones to fining-upward sandstones at the top | In the Llanos foothills, palynofloras in the Barco formation are dominated by brackish-water palms |                          |               |                     |                                             |
| Los Cuervos Fm. (upper Palaeocene to lower Eocene); abrupt change from amalgamated to non-amalgamated fluvial system |                                             |                                             |                          |                                             |               |                     |                                             |
| Foothills: laminated to locally bioturbated dark-gray organic-rich claystones and mudstones with thin coal seams interbedded with fining- and coarsening-upward successions with cross-bedded, ripple and wavy laminated sandstones. Thick palaeosols succession toward the top | Mud-rich coastal plains with tidal influence and isolated sandstone belt systems (channel, crevasse) passing upsection to subaerial flood plains | Dominance of continental palynomorph, with very humid conditions at the base |                          |                                             |               |                     |                                             |
| Note: References: 1: Caballero, Mora, et al. (2013); 2: Naranjo et al. (2013); 3: Prince et al. (2016); 4: Price et al. (2018); 5: Moreno et al. (2011); 6: Pardo, Jaramillo, and Oboh-Ikuenobe (2003) and Pardo (2004); 7: Montaño et al. (2016); 8: Sanchez et al. (2012); 9: Saylor et al. (2011); 10: Pardo and Jaramillo (2013); 11: Bayona et al. (2007); 12: Bayona et al. (2008); 13: Cooper et al. (1995). |                          |                                             |                          |                                             |               |                     |                                             |
and Lisama formations is gradational (Ward et al., 1973), as observed in the increase in sandstone beds accumulated in deltaic to fluvial systems (Table 2). This gradational change could be explained by the continuous crustal tilting of the basin. In C2, a regional unconformity separates coastal-plain deposits of the lower Palaeocene Guaduas formation from amalgamated channel-fill succession of the middle Palaeocene Socha Sandstone formation; the inversion of nearby structures explains the abrupt increase in quartzose sandy detritus supply and the increase in flow regime across the unconformity (see discussion above). In C3, Barco formation sandstones overlie unconformably Palaecocene strata to the west and Maastrichtian–Campanian strata to the east as result of the eastward migration of lithospheric flexure (Bayona, Jaramillo, Rueda, Reyes-Harker, & Torres, 2007).

Lithofacies associations, palaeocurrents and sandstone composition support the hypothesis of three separate basins. In C1, heterolithic bedding and wavy lamination in very fine- to medium-grained litharenites to sublitharenites, and the dominance of tabular geometries have been used as criteria to define deltaic systems affected by either tidal or wave actions. However, the absence of marine macrofossils along with the presence of continental and lacustrine palynomorphs, coal seams interbedded with massive mudstones that reveal palaeosol development, and the documentation of coeval westward- and eastward-oriented deltaic systems in upper Palaeocene strata (see references in Table 2), support the interpretation of deltaic systems that transition into fresh-water shallow lakes. Massive fine-grained sandstone beds, interbedded with heterolithic bedding and wavy lamination documented in the Lisama formation of C1, and in the Guaduas and Socha Mudstone formations in C2, have been interpreted by other authors as the product of tidal influence in coastal and deltaic environments (see Table 2). These physical sedimentary structures, however, are also documented in fresh-water fluvial and lacustrine systems (Kurcinka et al., 2018; Thomas et al., 1987; Zavala et al., 2006), where deltas and coastal plains are part of the lacustrine system. The low degree of bioturbation reported in these units may also correspond to bioturbation processes in shallow lacustrine systems (Scott, Buatois, & Mángano, 2012).

A continuous continental depositional system, as illustrated in Figure 2b, should exhibit an increment of compositional maturity, a decrease in grain size and a dominant directional trend of palaeocurrents. Northern clastic wedges C1 (proximal segment) and C2 (distal segment) did not show these characteristics. Lisama formation sandstones (C1) are dominantly fine-grained litharenites, whereas the sandstone composition of the Socha Sandstone formation (C2) is quartzose and includes pebbles and granules (Table 2). In C2, the scattered flow directions may represent development of fluvial systems within fine-grained continental plains around lacustrine systems in closed basins (Figures 10 and 11). This depositional setting developed mainly during the accumulation of the middle interval of the Guaduas formation and in the Socha Mudstone formation, when supply of sandy terrigenous material decreased and coastal/flood plains were poorly drained (Figure 7).

Continental lacustrine systems might have been connected, for short time intervals, with depositional marine systems documented farther to the north, where tidal and marine macrofossils have been reported in lower to middle Palaeocene mixed carbonate-siliciclastic rocks (Hatunuevo, Manantial and Guasare formations, see references in Bayona et al., 2011; Figure 14). Palynological analysis has supported brackish water vegetation and marine incursions in the lower Palaeocene Guaduas formation in the northern clastic wedge C2, and in the Barco formation in the northern clastic wedge C3 (Table 2). Opposite palaeocurrents indicate tidal influence in channel-fill successions at the base of the Socha Sandstone formation in the Cocuy basin.

5.7 Late Palaeocene–early Eocene tectonism and magmatism

Thick palaeosol deposits (20–100 m thick) are documented in the three basins in the latest Palaeocene–early Eocene time, coeval with the broad development of intra-basinal magmatism and the warmest period in the Cenozoic (Zachos et al., 2001). The record of thick mud deposition and later palaeosol development in upper Palaeocene-lower Eocene strata in each of the three isolated basins may be explained by the onset of continental-scale dynamic loading subsidence mechanism. Early Palaeocene subduction occurred along the northern margin (Bustamante et al., 2017) and generated a viscous corner flow that initially did not affect the tectonic subsidence mechanism of the basins (as occurs in retro-arc foreland basins; Catuneanu, 2019). In latest Palaeocene time, at the peak of magmatism (Bayona et al., 2012) and easternmost extension of magmatic activity in the Llanos basin (Bayona et al., 2015; Franco et al., 2018), widespread dynamic subsidence outpaced local tectonic subsidence mechanisms, leading to (a) submerging low-amplitude palaeo-highs and (b) regional increase in tectonic subsidence. Hot and humid climatic conditions (Jaramillo et al., 2006, 2007) favoured the continuous supply of muddy detritus and extensive development of palaeosol profiles, as also documented in the southern segments of C1 (Bayona et al., 2013), C2 (Morón et al., 2013) and C3 (Bayona et al., 2015).

In the Cocuy and Floresta areas, stratigraphic sections in the Socha Mudstone formation separated <3 km increased in thickness by >50%. In the central foothills of the Eastern Cordillera, Cooper et al. (1995) reported active normal faults that affected the thickness of the Los Cuervos formation (Table 2). Difference in stratigraphic thickness of upper
5.8 | Comparison with other segments of the Andes

Palaeocene deformation is a dominant feature in the Andes of Ecuador and Peru, and synorogenic deposition took place over extensional Mesozoic basins (see review in Horton, 2018). In the northern Andes, deformation was concentrated mainly along the western margin, where accretion of oceanic terranes occurred; however, deformation and magmatism also occurred eastward. In the Ecuador foreland basin, Baby et al. (2013) reported a phase of inversion of Jurassic extensional structures located at 300 km from the collisional margin during Maastrichtian–Palaeocene time. In this study, we document how deformation and magmatism in the Colombian margin extended >400 km from the Palaeocene magmatic arc eastward (Montes et al., 2019). If we consider the synorogenic basins and the accretionary orogen that formed on the oceanic side (Pardo-Trujillo et al., 2020), the width of the deformation zone extended >100 km to the west of the Palaeocene magmatic arc.

A reasonable explanation of the eastward extension of deformation and magmatism is the slab flattening of the Caribbean subduction zone (Bayona et al., 2012, 2013), whereas other segments of the Andes to the south were controlled by the dynamics of the subduction of the Farallon plate beneath the South America plate (e.g. Baby et al., 2013; Faccenna, Oncken, Holt, & Becker, 2017). Horizontal subduction explains deformation far from the trench (e.g. Martinod, Husson, Roperch, Guillaume, & Espurt, 2010; Ramos & Folguera, 2009), as demonstrated with numerical models (see review in Martinod, Gérault, Husson, & Regard, 2020). Horizontal subduction also explains the inboard migration and termination of volcanism (e.g. Bayona et al., 2012; English, Johnston, & Wang, 2003; Gutscher, Spakman, Bijwaard, & Engdahl, 2000). The northernmost margin of South America recorded a first phase of widening of the Andean deformation in Palaeocene time, with reactivation of inherited Lower Cretaceous structures, whereas in the central Andes Palaeocene deformation concentrated in the western margin with a development of a continuous foreland basin to the east (DeCelles & Horton, 2003; Horton, 2018; Horton et al., 2001).

6 | CONCLUSIONS

Supply of synorogenic detrital material into the former passive-margin Cretaceous basin produced two types of unconformities. The earliest, interpreted as a xenocform, corresponds to the end of chemical precipitation of the La Luna formation and onset of synorogenic deposition of the Los Pinos formation at the earliest Maastrichtian time. Disconformable contacts at the base of the Guaduas formation, at the base of the upper interval of the Guaduas formation, and at the base of the Socha Sandstone formation are interpreted as the response to the increasing supply of quartzose fragments from nearby source areas. These basins record the shift of depositional conditions from anoxic marine, to oxygenated shallow marine, to coastal plains and finally to fully continental swamps, fluvial and lacustrine environments developed in closed basins.

These changes of depositional environment were accompanied by up-section increase in sandy to pebbly terrigenous fragments supplied from two nearby source areas. Eastern quartzose-rich source areas supplied quartzose fragments and reworked glauconite to the Cocuy depocenter, whereas muddy Cretaceous and sandy Jurassic sedimentary cover to the west supplied muddy and unstable lithic and heavy minerals fragments. The abrupt change from quartzose sandstones to lithic-bearing sandstones in the latest Palaeocene time, along with the ratios of stable and unstable heavy minerals, documents the unroofing of the Cretaceous–Jurassic sedimentary cover nearby, not the supply from basement rocks exposed in the Central Cordillera to the west. Local fault activity is documented by the abrupt change in stratigraphic thickness of upper Palaeocene strata. The comparisons of detrital zircons populations reported in Maastrichtian to Palaeocene units from the Magdalena, Cocuy, Catatumbo, Floresta and Medina basins indicate that the northern Middle Magdalena and Cocuy basins were disconnected, whereas a connection with southwestern basins is required to explain the content of Cretaceous to Ordovician detrital zircons. However, erosion of Cretaceous volcanic rocks may also have supplied some of the Late Cretaceous zircons reported in the Cocuy basin. Evidence of intra-basinal volcanism is constrained by the presence of poorly altered volcanic rocks fragments and reported volcanic zircons of Palaeocene–early Eocene age. Slab flattening of the Caribbean plate and dynamic topography played a role in the filling of early Eocene basins, eastward reactivation of inherited lower Cretaceous faults and the eastward migration of magmatism.
Our results contribute to the understanding of how to detect early phases of deformation of other major mountain belts adjacent in convergent margins, by identifying reworking of sedimentary cover in adjacent uplifts versus supply of terrigenous detritus from basement rocks in very distal source areas.

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CONFLICT OF INTEREST

Authors have no conflict of interest to declare.

PEER REVIEW

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DATA AVAILABILITY STATEMENT

The datasets used and/or analysed during the current study are available from the corresponding author on reasonable request.

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**SUPPORTING INFORMATION**

Additional supporting information may be found online in the Supporting Information section.

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