# Onset of fault reactivation in the Eastern Cordillera of Colombia and proximal Llanos Basin; response to Caribbean–South American convergence in early Palaeogene time

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Abstract: The inversion of Mesozoic extensional structures in the Northern Andes has controlled the location of syn-orogenic successions and the dispersal of detritus since latest Maastrichtian time. Our results are supported by detailed geological mapping, integrated provenance (petrography, heavy minerals, geochronology) analysis and chronostratigraphical correlation (palynological and geochronology data) of 13 areas with Palaeogene strata across the central segment of the Eastern Cordillera. Spatial and temporal variation of sedimentation rates and provenance data indicate that mechanisms driving the location of marginal and intraplate uplifts and tectonic subsidence vary among syn-orogenic depocentres. In the late Maastrichtian-mid-Palaeocene time, crustal tilting of the Central Cordillera favoured reverse reactivation of the western border of the former extensional Cretaceous basin. The hanging wall of the reactivated fault separated two depocentres: a western depocentre (in the Magdalena Valley) and an eastern depocentre (presently along the axial zone of the Eastern Cordillera, Llanos foothills and Llanos Basin). In late Palaeocene-early Eocene time, as eastern subduction of the Caribbean Plate and intraplate magmatics advanced eastwards, reactivation of older structures migrated eastwards up to the Llanos Basin and disrupted the eastern depocentre. In early Eocene time, these three depocentres were separated by two low-amplitude uplifts that exposed dominantly Cretaceous sedimentary cover. Synorogenic detrital sediments supplied from the eastwards-tilted Central Cordillera reached areas of the axial domain of the Eastern Cordillera, whereas unstable metamorphic and sedimentary fragments recorded in the easternmost depocentre were supplied by basement-cored uplifts with Cretaceous and Palaeozoic sedimentary cover reported in the southern Llanos Basin. This tectonic configuration of low-amplitude uplifts separating intraplate syn-orogenic depocentres and intraplate magmatic activity in Palaeocene time was primary controlled by subduction of the Caribbean Plate.

**Supplementary material:** Appendix 1 presents detailed descriptions of analytical methods used in this manuscript. Appendixes 2 to 4 include raw data of sandstone petrography, heavy minerals and U–Pb detrital zircon geochronology, respectively. All this material is available at http://www.geolsoc.org.uk/18597.

Reactivation of pre-existing structures occurs as one response to changes in tectonic regime. Positive inversion structures (Hayward & Graham 1989; Coward 1994; Sibson 1995; Bayona & Lawton 2003) involve the reactivation of a former normal fault; in the Northern Andes, major uplifts have occurred along inversion structures (e.g. Cooper *et al.* 1995; Colleta *et al.* 1997; Cortés *et al.* 2006;

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Mora *et al.* 2008*a*). Several factors may control whether an older structure may be reactivated: (a) fault and basin geometries (Butler 1989; Buchanan & McClay 1991, 1992; Bayona & Thomas, 2003); (b) potential of fault reactivation (e.g. pore pressure, friction, cohesion and fault gouge material: Sibson 1985, 1995; Etheridge 1986); and (c) reorganization of stresses along the collided margin (Kluth & Coney 1981; Ziegler *et al.* 1998; Marshak *et al.* 2000).

Inversion structures fragment former basin geometry and drive the evolution of compartmented syn-orogenic basins. As fault reactivation develops, tectonic fragmentation of the drainage system, depositional patterns and climate conditions begin to differ between localized depocentres (Bayona & Thomas 2003; Sobel & Strecker 2003; Sobel et al. 2003; Hilley & Strecker 2005; Strecker et al. 2007, 2009, 2011). Therefore, depositional environments, stratigraphical thicknesses and provenance signatures begin to vary among depocentres (Bayona & Lawton 2003; Hain et al. 2011). As a result, the Magdalena Valley Basin and the Llanos foreland basin of Colombia are two basins with drainage, climate and depositional systems that were totally disrupted by the rapid vertical uplift of the Eastern Cordillera of Colombia that occurred since late Miocene-Pliocene time (Fig. 1a) (see Mora et al. 2008b for review of the youngest events of Eastern Cordillera uplift). However, upper Cretaceous rocks of the Magdalena Valley, Eastern Cordillera and Llanos Basin accumulated within the same basin and represent a single depositional profile of a shallow platform in a passivemargin setting (Cooper et al. 1995; Villamil 1999; Cediel et al. 2003).

Conversely, modelling of undisrupted basins, such as foreland basins (DeCelles & Giles 1996), has allowed predictions of patterns of deformation (i.e. the advance of orogenic front and flexural subsidence), sediment routing (provenance signature) and accumulation patterns (stratal geometry and three-dimensional lithofacies distribution) to be made (Beaumont 1981; Jordan 1981; Heller et al. 1988; Flemings & Jordan 1989; Crampton & Allen 1995; Burbank et al. 1996; DeCelles & Giles 1996; Sinclair 1997; DeCelles & Horton 2003). In orogens with multiphase deformation, such as the Northern Andes (Bayona et al. 2008), comprehensive basin stratigraphy and provenance analyses become the key to document the timing and sense of fault movement along reactivated structures.

The timing of when a basin becomes a positive feature is relevant for the determination of the cessation of burial processes of source rocks and the formation of early structures in a petroleum system. Definition of the onset of positive inversion in the Northern Andes has been controversial, and the timings range from Maastrichtian-Palaeocene (Sarmiento-Rojas 2001; Montes et al. 2005; Cortés et al. 2006; Bayona et al. 2008; Moretti et al. 2010; Parra et al. 2012), to middle-late Eocene (Toro et al. 2004; Gómez et al. 2005a; Parra et al. 2009a, b; Horton et al. 2010b; Saylor et al. 2011, 2012), late Eocene (Mora et al. 2010) and late Oligocene-Miocene (Cooper et al. 1995; Villamil 1999). In order to address this problem of the timing of fault reactivation in a multiphase orogeny, like the Eastern Cordillera, we integrated the analysis of depositional environments, thickness variations and provenance signatures of Maastrichtian-lower Eocene strata that are presently compartmentalized by several structures within the double-verging Eastern Cordillera of Colombia (Fig. 1). Our study specifically examines the Maastrichtian-early Palaeogene time interval to document possible intraplate deformation events coeval with the oblique convergence of the Caribbean Plate or the collision of oceanic arcs with the South American Plate (see the review in Bayona et al. 2012). The approach used in this study may be applied in multiphase orogens where low-temperature thermochronometry does not reveal early phases of shortening but abrupt changes in provenance signatures and sedimentation rates record those early phases of deformation.

# **Regional setting**

Three major mountain belts are the result of the complex interaction of the Nazca, Caribbean and South America plates since the Late Cretaceous: Western Cordillera (WC), Central Cordillera (CC) and the double-verging Eastern Cordillera (EC) (Fig. 1a, b). The Eastern Cordillera is interpreted as a wide Cretaceous extensional basin formed during at least two stretching events (Sarmiento-Rojas et al. 2006) and tectonically inverted during the Cenozoic (Colleta et al. 1990; Dengo & Covey 1993; Cooper et al. 1995; Mora et al. 2006). The geometry of the double-verging Eastern Cordillera changes northwards from a narrow belt in the southern domain that broadens in its central segment where the Cordillera is bounded by east- and west-verging thrust belts (Fig. 1b). Further to the north, the Eastern Cordillera bifurcates into the Santander Massif and the Mérida Andes (Fig. 1a).

The areas selected for this study are located in the three structural domains of the central segment of the Eastern Cordillera (Fig. 1c). Its western domain includes a west-verging thrust and fold belt that places Cretaceous and Palaeogene strata in contact with Cenozoic strata and buried structures of the southern Middle Magdalena Valley Basin. In

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Fig. 1. (a) Location of the three Cordilleras in the Northern Andes and Romeral Fault System (RFS). (b) Regional cross-section from the Northern Andes (modified from Bayona *et al.* 2012). (c) Geological map showing major structures of the Eastern Cordillera and the location of selected studied areas. For reference, the studied areas were grouped into four structural domains: western; western axial; eastern axial; and eastern domains.

the frontal splay of the west-verging system is the north-south-trending Guaduas Syncline (Cortés *et al.* 2006). The axial domain consists of tight anticlines and broad synclines related to double-verging faults (McLaughlin & Arce 1975; Fajardo-Peña 1998; Toro *et al.* 2004; Bayona *et al.* 2008). This domain broadens northwards of Bogotá City, and is divided into western axial and eastern axial domains by the east-verging Pesca–Machetá Fault System and the Boyacá–Soapaga Fault System (Fig. 1c). The eastern domain includes east-verging faults placing Cretaceous and Cenozoic strata over Cenozoic strata of the Llanos foreland basin. Basement rocks of the Quetame Massif are exposed in the southern segment of the eastern domain (Fig. 1c). The NE-trending Medina Syncline is part of the frontal splay of the east-verging fault system (Parra *et al.* 2009*a*). Reverse structures in these three domains of the Eastern Cordillera have been interpreted as an inversion of former normal faults (Cooper *et al.* 1995; Branquet *et al.* 2002; Restrepo-Pace *et al.* 2004; Cortés *et al.* 2006; Kammer & Sanchez 2006; Mora *et al.* 2006, 2008*a*).

Interpretations of basin geometry for the early Palaeogene vary from a single and continuous foreland basin (Cooper et al. 1995; Gómez et al. 2005a; Parra et al. 2009b; Horton et al. 2010b; Saylor et al. 2011, 2012), a continuous negative flexural basin with an absence of bounding thrusts (Pindell et al. 2005) to a foreland basin disrupted by uplifts (Bayona et al. 2008). The location of intraplate uplifts has been proposed in the western border of the basin (Bayona et al. 2003; Restrepo-Pace et al. 2004; Cortés et al. 2006; Moretti et al. 2010), along the axial zone of the basin (Fabre 1981, 1987; Sarmiento-Rojas 2001; Pardo 2004) or at both borders of the basin (Fajardo-Peña 1998; Villamil 1999). Geodynamic modelling of Palaeogene strata encompassing the axial zone of the Eastern Cordillera to the Llanos Basin indicates an eastwards migration of tectonic loads within the restored Eastern Cordillera to explain the Maastrichtian-Palaeogene geometry of the syn-orogenic basin (Bayona et al. 2008).

Early Palaeogene (45–65 Ma) magmatic activity has been reported along the northern Andean continental margin (Central Cordillera–Santa Marta Massif in Fig. 1a, b) (Aspden *et al.* 1987; Cardona *et al.* 2011), and volcanic detrital zircons have been recognized in Palaeocene–lower Eocene sandstones as far as 400 km from the magmatic arc (Saylor *et al.* 2011, 2012; Bayona *et al.* 2012). Marginal and intraplate magmatism during early Palaeogene time is related to the convergence and shallow subduction of the Caribbean crust beneath the South America Plate, and the abrupt termination of volcanism in middle Eocene time to a change to a transpressive non-magmatic margin (Bayona *et al.* 2012).

# Maastrichtian-lower Eocene stratigraphy and depositional environments

In this section we summarize lithological associations and depositional environment interpretations proposed for Maastrichtian–lower Eocene units in the three structural domains of the central segment of the Eastern Cordillera, as these units record the regional shift from marine to continental depositional environments as a response to the accretion of oceanic terranes west of the Romeral Fault System (Fig. 1) (e.g. Etayo-Serna *et al.* 1983; McCourt *et al.* 1984).

Maastrichtian-lower Eocene stratigraphical units of the western domain have been defined in the western flank of the Guaduas Syncline (Fig. 1c), and include, from base to top, the Umir, Cimarrona, Seca and Hoyon formations (Fig. 2). The lower two units change laterally and vertically from calcareous dark-grey mudstones with abundant fossil content (bivalves, foraminifera), to cross-bedded conglomerates and fine-grained sandstones (Gómez & Pedraza 1994); conglomerate beds of the Cimarrona Formation change eastwards to limestone beds in the eastern flank (Gómez et al. 2003). The overlying Seca Formation includes dark-coloured mudstones with thin coal interbeds at the base, changing upsection to varicoloured massive mudstones with interbeds of fine- to medium-grained sandstones towards the top (De Porta 1966). The Hoyon Formation comprises interbeds of mudstones, sandstones and conglomerates that range in size from metre- to decametre-scale, and is divided into three informal units. The lower Hoyon unit is a conglomerate interval with upwards-coarsening successions, and has transitional contacts with sandstones and mudstones of the underlying Seca Formation and with mudstones of the overlying middle Hoyon unit; the latter units comprise dominantly massive red mudstones. In contrast, the upper Hoyon unit is a conglomerate interval resting unconformably upon the middle Hoyon unit. Both in the lower and the upper Hoyon units, horizontal bedding dominates over cross-bedding as internal structures, and an east-NE direction of transport is documented by palaeocurrent measurements (Gómez et al. 2003). Gómez et al. (2003) interpreted the Maastrichtian-lower Eocene succession as a regression from a shallow-marine environment with fan deltas (the Umir and Cimarrona formations), changing to coastal and alluvial plains (the Seca Formation), and ending up with fluvial and alluvial fan deposits (the Hoyon Formation).

In the axial domain, Maastrichtian–lower Palaeocene strata include the Guaduas Formation, overlain by Palaeocene–lower Eocene Cacho and Bogotá formations to the west of the Pesca– Machetá Fault System, or by the Lower and Upper Socha formations to the east of the Pesca– Machetá Fault (Fig. 2). The Guaduas Formation rests conformably over fine- to medium-grained sandstones of the Guadalupe Formation, with crossbeds and abundant ichnofossils (Perez & Salazar 1978; Fabre 1981). The lower half of the Guaduas Formation includes fine-grained strata with abundant coal seams, whereas the upper half includes

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**Fig. 2.** (a) Lithostratigraphical correlation of the Maastrichtian–Lower Eocene strata in the four structural domains, showing lateral variation of the thickness and lithology of stratigraphical units (the names of the lithological units are shown to the left of each structural domain). Datum, lower–middle Eocene unconformity. (b) Chronostratigraphical correlation of upper Campanian–middle Eocene strata based on palynological and geochronological data. See Figure 1 for the location of sections, and Table 1 for a summary of thickness and chronostratigraphical data per area.

laminated mudstones, and massive light-coloured mudstone and sandstone interbeds (Sarmiento 1992). The Cacho–Lower Socha formations are cross-bedded, fine-to coarse-grained upwards-fining

quartzarenites, locally conglomeratic; these units rest disconformably upon light-coloured mudstones of the Guaduas Formation (Pardo 2004; Bayona *et al.* 2010). In the western axial domain, the

Bogotá Formation overlies conformably the Cacho Formation, and consists dominantly of massive multicoloured sandy mudstone with isolated upwardsfining and -coarsening massive to cross-bedded sandstones (Bayona et al. 2010). In the eastern axial domain, the Upper Socha Formation overlies conformably the Lower Socha Formation and it changes upsection from bioturbated dark-grev organic-rich claystone and mudstone with thin coal seams to massive light-coloured sandy mudstone with isolated upwards-fining and -coarsening sandstones with cross beds, wavy lamination and ripple cross-lamination (Pardo 2004). Coarsegrained to conglomeratic sandstones of the Regadera-Picacho formations disconformably overlie the Bogotá-Upper Socha formations. Sarmiento (1992) interpreted the regression from shallowmarine to alluvial plain conditions for the Guaduas Formation, whereas Bayona et al. (2010), Pardo (2004) and Saylor et al. (2011) interpreted overlying units as deposition in fluvial systems varying from braided, to meandering to anastomosing.

In the eastern domain, as well as in the proximal Llanos Basin, the Guaduas Formation is very thin or absent, and the Barco and Cuervos formations rest unconformably on upper Cretaceous units (Cooper et al. 1995; Bayona et al. 2008) (Fig. 2). The succession described below corresponds to the western flank of the Medina Syncline. Maastrichtian strata consist of a coarsening-upwards succession from medium-grained to conglomeratic sandstones with cross-beds and reworked bivalves, oysters and corals of the Guadalupe Formation (Guerrero & Sarmiento 1996). The Barco Formation comprises fine- to medium-grained quartzose sandstones and is overlain by interbeds of upwardsfining sandstone and mudstone with bidirectional cross-bedded and bioturbated heterolithic laminated sandstone of the lower Cuervos Formation. The middle and upper Cuervos Formation resembles the lithological association of the Upper Socha Formation. Guerrero & Sarmiento (1996) interpreted Maastrichtian strata as shallow-marine deposits, whereas Cazier et al. (1995), Reyes (1996) and Bayona et al. (2008) interpreted the Barco-Cuervos succession as coastal plains with a tidal influence. The Mirador Formation is a coarsegrained sandstone unit of lower-middle Eocene age (Bayona et al. 2008; Parra et al. 2009a) that disconformably overlies the Cuervos Formation.

## Selection of study areas and methods

In order to determine spatial and temporal changes in basin geometry for a given time interval, it is crucial to consider the chronostratigraphical correlation of lithological units described earlier at different structural domains. The survey of Maastrichtian-lower Eocene strata included three areas in the Guaduas Syncline (San Juan de Río Seco East and West, and Guaduero), six areas in the western axial zone (Usme, Guatavita, Machetá, Checua, Tunja and Paipa), five areas in the eastern axial zone (Umbita, San Antonio, Rondón, Pesca and Paz de Río) and two areas in the eastern domain (Cusiana and Medina) (Fig. 1b).

The thickness of the Maastrichtian-lower Eocene lithostratigraphical units described previously was measured in synclinal structures and within areas with minimum structural complexity. Lithostratigraphical units were mapped in both flanks of each syncline structure, and the lateral variation of thickness was calculated using local structural cross-sections. Samples for palynological and provenance analyses (petrography, heavy mineral and detrital U–Pb zircon geochronology) were collected for each measured stratigraphical section.

Palynology and detrital volcanic zircons supply the age control for Maastrichtian–lower Eocene units. Descriptions of analytical methods and raw data for palynological, sandstone petrography, heavy minerals and geochronological analyses are presented in the Supplementary Material. Palynology and detrital volcanic zircons supply the age control for Maastrichtian-lower Eocene units. The results of 381 palynological samples are summarized in Table 1. Calculation of maximum age of deposition based on U–Pb ages in volcanic zircons follows the results presented in Bayona *et al.* (2012).

Sandstone petrography (123 samples), heavy mineral (46 samples) and U-Pb detrital zircon geochronology (28 samples) analyses were carried out in late Campanian-lower Eocene sandstones in order to compare composition, heavy mineral assemblage and detrital zircon age populations between structural domains. We integrate the results of seven U-Pb detrital zircon geochronology samples for Maastrichtian-Lower Eocene strata in the Floresta Basin (or Paz de Río area of the eastern axial domain: Saylor et al. 2011, 2012), one sample of lower Eocene strata in the Macheta area (Horton et al. 2010b) and two samples of the northern Guaduas Syncline (Caballero et al. 2013) in order to compare detrital zircon age populations from the western, axial and eastern domains from Campanian to early Eocene times.

Heavy minerals were grouped into ultrastable (zircon, rutile, turmaline), stable (apatite, garnet), unstable (amphibole, epidote, pyroxene, chlorite, talc and others) and muscovite. Zircon crystallization ages are divided into three major provinces: 65–360 Ma from the Central Cordillera; 360–1300 Ma from basement rocks of the Eastern

Cordillera and Santander Massif; and zircons supplied from basement rocks in the Guyana Craton have Neoproterozoic ages (600-650 Ma) and populations older than 1300 Ma (see Horton *et al.* 2010*a*, *b* for a detailed discussion of zircon crystallization ages in the Northern Andes). The Cretaceous sedimentary cover of the Eastern Cordillera includes a mixture of the last two provinces (see details later).

# Thickness variation and age control of Maastrichtian-lower Eocene strata

This section presents the most significant trends of stratigraphical thickness variations for studied stratigraphical units within structural domains. Figure 2a illustrates the measured stratigraphical thickness for each unit in each area and the position of samples with volcanic detrital zircons. Palynological results (summarized in Table 1) and calculations of maximum depositional age were used to construct the chronostratigraphical correlation presented in Figure 2b.

In the western domain, the stratigraphical thickness of the Seca Formation increases eastwards and northwards (Table 1; Fig. 2a). In the area of Guaduero, the thickness ranges from 490 m to more than 700 m, whereas in the San Juan de Rio Seco area the thickness ranges from 400 to 470 m. The lower and middle intervals of the Hoyon Formation are completely recorded in the eastern flank of the San Juan de Rio Seco area, and are partly eroded by the unconformity in the western flank of the San Juan de Rio Seco area and completely eroded in the Guaduero area (Fig. 2a). For the Seca Formation, a Maastrichtian palynological age at the base and detrital volcanic zircon ages at the top  $(61.9 \pm 2 \text{ Ma})$  constrain a Maastrichtianlower Palaeocene age (Table 1). The lower Hoyon unit yielded only two zircons with a Palaeocene age, 62 Ma being the youngest zircon, whereas the maximum age of deposition for the middle Hoyon unit is  $56.3 \pm 1.6$  Ma. The upper Hoyon yielded middle Eocene-lower Oligocene palynological ages, and only one zircon yielded a 55 Ma age.

In the western axial domain, the Guaduas Formation shows its maximum thickness in the central area of Checua (1098 m), and decreases to less than 600 m in northwards, southwards and eastwards directions (Table 1; Fig. 2a). Lower and middle intervals of the Guaduas Formation range in age from uppermost Campanian to Maastrichtian, and only the uppermost strata yielded a lower Palaeocene age (Fig. 2b). The Cacho Formation yielded four detrital zircon ages in the range of 61–64 Ma in the Checua area, yielding an early Palaeocene age of deposition (Fig. 2b). Thickness variations are similar to the Guaduas Formation, with a maximum thickness of 245 m in the Checua area. The stratigraphical thickness of the Bogotá Formation is very variable, with a maximum of 1415 m in Usme and a minimum of 169 m in Tunja; regionally, its thickness decreases northwards. Detrital volcanic zircon ages and palynology indicate a middle Palaeocene–lower Eocene age for this unit. The middle Eocene Regadera Formation disconformably overlies strata of the Bogotá Formation (Bayona *et al.* 2010).

In the eastern axial domain, the thickness of the Guaduas Formation varies from 290 to 609 m (Table 1; Fig. 2a) and the upper interval of the Guaduas Formation is Maastrichtian in age (Fig. 2b). In Umbita only, an eastwards decrease in thickness is documented from 525 to 350 m. The stratigraphical thickness of the Lower Socha Formation decreases northwards from 257 m in Umbita to a range between 90 and 138 m in Paz de Río (Pardo 2004; Saylor et al. 2011). Similarly, the thickness of the Upper Socha decreases from 450-460 m in Umbita and San Antonio to 255-280 m in Pesca and Paz de Río. Saylor et al. (2011) documented an abrupt change from 227 to 344 m in the Upper Socha Formation over a distance of less than 3 km. The Lower Socha Formation unit is poorly dated as lower-middle Palaeocene in the Paz de Río area (Pardo 2004) (Fig. 1b), whereas the Upper Socha Formation has palynological and mineral-age constraints indicating a late Palaeocene-early Eocene age (Table 1; Fig. 2b). The overlying Picacho Formation has been assigned an early-middle Eocene age (Pardo 2004).

The Guaduas Formation in the eastern domain is less than 50 m thick and yielded only a Maastrichtian age (Parra *et al.* 2009*a*); coeval strata in the Cusiana area are not reported (Fig. 2a). This unit overlies upper Campanian–lower Maastrichtian strata of the Guadalupe Formation (Guerrero & Sarmiento 1996). The sandy Barco Formation is early–middle Palaeocene in age, whereas the Cuervos Formation has palynological and mineral-age data that indicate a late Palaeocene–early Eocene age (Bayona *et al.* 2008; Parra *et al.* 2009*a*) (Table 1; Fig. 2). The thickness of both the Barco and Cuervos units decreases eastwards, as well as the thickness of the lower–middle Eocene Mirador Formation (Parra *et al.* 2009*a*).

# **Chronostratigraphical correlation**

Because stratigraphical units are diachronous (Fig. 2b), we established the following boundaries at 65, 59, 55 and 49 Ma for chronostratigraphical correlation of strata. The following criteria were used to determine the position of those times for

	Lithological	Thickne	ess (m)	Palynology			Mineral age
	unit	Western flank	Eastern flank	No. of samples	Results	Reworked pollen	(number of zircons for mean calculation)
Western domain							
Guaduero (this study)	Upper Hoyon	230	220	3	Base: Middle-Late Eocene	Campanian– Maastrichtian	
	Middle Hovon	0	0				
	Lower Hovon	Õ	Õ				
	Seca	490	>703	22	Base: Campanian– Maastrichtian boundary		Top: $61.9 \pm 2 Ma(5)$ ; 60 Ma(youngest) age)
	Umir/Cimarrona			2	Top: Late Campanian		
San Juan Río Seco; western flank: Gómez <i>et al.</i> (2003):	Upper Hoyon	460	250	19	Top: Middle Eocene–Early Oligocene	Late Cretaceous	
eastern flank (this study)	Middle Hoyon	310	1185	13	No recovery		
	Lower Hoyon	667	595	11	No recovery		
	Seca	400	>470	3	No recovery		
	Umir–Cimarrona	350		4	Top: Campanian– Maastrichtian boundary		
Western axial domain							
Paipa (this study)	Bogotá		485	1	No recovery		
	Cacho		200	5	No recovery		
	Guaduas		650	12	Top: Early Palaeocene; metre 500 and 600 is 65 Ma		
Tunja (this study)	Bogotá		169	11	Top: Palaeocene–Eocene	Campanian– Maastrichtian	
	Cacho		155	3	No recovery		
	Guaduas		600	14	Top: Late Maastrichtian; Middle: Campanian– Maastrichtian boundary		

**Table 1.** Measured stratigraphical thickness, palynological results and maximum age of deposition for Maastrichtian–Eocene strata (data from Bayona et al. 2012; error for the youngest age of a single crystal is not shown). Italic font correspond to the oldest unit studied in each area. These areas correspond to the location of major syncline structures, as shown in Figure 1

Checua (western flank (Sarmiento (1992 and this study); eastern flank (this study)	Bogotá	>629	>484	4 33 Top: Early Eocene– Oligocene; Base: middle–late Palaeocene			Top: 58 Ma (youngest age)
	Cacho	219	245	6	No recovery		Bottom: 64.3 + 2.1 Ma (3)
	Guaduas	908–1098		60	Top: Early Palaeocene; 65 Ma at metre 715; Middle: Maastrichtian; Base: Campanian– Maastrichtian		_ ()
Machetá (this study; mineral age from Horton <i>et al.</i> 2010 <i>a</i> , <i>b</i> )	Bogotá		646	7	Top: Early Eocene	Early Palaeocene– Late Maastrichtian	Top: 58.4 ± 1.2 (3)
	Cacho		84				
	Guaduas		669	_			
Guatavita (this study)	Bogotá	883	1180	5	Top: Early Eocene		
	Cacho		170	4	No recovery		
U (D 1 0010)	Guaduas	1 4 1 5	>450	3	No recovery	<i>a</i> .	T
Usme (Bayona et al. 2010)	Bogotá	1415			Top: Early Eocene; Middle: Base: middle– late Palaeocene; Base: Palaeocene?	Campanian– Maastrichtian	In metre 530: 56.2 $\pm$ 1.6 (6)
	Cacho	100			Palaeocene?		
	Guaduas	>600			Top: Palaeocene?		
Eastern axial domain							
Paz de Río (thickness and age from Pardo 2004; mineral age from Saylor <i>et al.</i> 2011, 2012)	Upper Socha		280		Top: Early Eocene; Middle and base: Late Palaeocene		Top: $55.4 \pm 0.5$ (6); Middle: $58.2 \pm 2$ (8)
	Lower Socha		138		Early-middle Palaeocene		
	Guaduas		400		Maastrichtian		
Pesca (this study)	Upper Socha		255	8	Base: middle-late Palaeocene		
	Lower Socha		125				
	Guaduas		290	7	Middle and Base: Late Campanian–Late Maastrichtian		

(Continued)

Table 1. Continued

	Lithological	Thickn	Thickness (m) P				Mineral age	
	unit	Western flank	Eastern flank	No. of samples	Results Reworked pollen		(number of zircons for mean calculation)	
Rondón (this study)	Upper Socha	355		25	Top: Early Eocene; 55 Ma is between metres 100 and 200			
	Lower Socha	200		4	Middle Palaeocene?	Late Maastrichtian		
	Guaduas	295		61	Top: Late Maastrichtian; Base: Early Maastrichtian			
San Antonio (this study)	Upper Socha		324-460	6	No recovery			
	Lower Socha		225 - 262		No recovery			
	Guaduas		609	3	Latest Campanian– Maastrichtian			
Umbita (this study)	Upper Socha	450	400	5	Base: middle-late Palaeocene	Maastrichtian	Top: 53 Ma (youngest age)	
	Lower Socha	230	257	2	Early-middle Palaeocene		2 )	
	Guaduas	525	350	19	Base: Maastrichtian			
Eastern domain								
Cusiana (Bayona et al. (2008))	Mirador	141						
	Cuervos	158						
	Barco	56						
	Guaduas	0						
Medina (western flank from	Mirador	237	135		Early-middle Eocene			
Parra <i>et al.</i> 2009 <i>a</i> ); eastern flank (from this study and a	Cuervos	477	348-440		Middle-late Palaeocene		$55.5 \pm 0.6 (9); 57.6 \pm 0.7 (6)$	
deep well)	Barco	231	125 - 187		Early-middle Palaeocene		_ ()	
-	Guaduas	50	53		Maastrichtian			

each stratigraphical section (Fig. 3) and the calculation of sedimentation rates presented in Table 2.

- Palynological data and maximum ages of deposition (Table 1) were used as the most reliable information for age constraints.
- When the stratigraphical unit does not have an age, we used the age determined for the same unit in an area within the same structural domain.
- Sandy units (i.e. amalgamated channels in the Cacho Formation) are considered to have lower rates of accumulation than muddy units (i.e. the floodplains in the Bogotá Formation).

Campanian–lower Maastrichtian conglomeratebearing strata are at the western and eastern domains, while fine-grained strata of the Guaduas Formation accumulated over sandstones of the Guadalupe Formation in the axial zone (Fig. 2b). Onset of fine-grained deposition of the Guaduas and Seca formations is older in the western domain than in the axial domain; in this later domain, onset of deposition is nearly coeval (Fig. 2b). The lacuna (non-deposition + erosion) overlying Seca and Guaduas strata increases eastwards between domains, and northwards in the four domains. Onset of fine-grained deposition of the Bogotá–Cuervos



**Fig. 3.** Identification of chronostratigraphical markers in the stratigraphical column of the Umbita area, and the location of boundaries of chronostratigraphical intervals. Note the change in rate of accumulation between the sandy and fine-grained intervals. See Table 2 for the identification of chronostratigraphical intervals in other areas.

Domain	Chronostratigraphical interval (Ma)	Western flank		Lithological	Eastern flank		Lithological
		Thickness (m)	Rate (m Ma <sup>-1</sup> )		Thickness (m)	Rate (m Ma <sup>-1</sup> )	unto
Western domain							
Guaduero	49-55	20	3.3	Seca	74	12.3	Seca
	55-59	38	9.5	Seca	63	15.8	Seca
	59-65	352	58.7	Seca	456	76.0	Seca
	65-70	17	3.4	Seca	no data	_	
San Juan Rio Seco	49-55	317	52.8	Middle Hovon	833	138.8	Middle Hovon
	55-59	442	110.5	Lower Hoyon	610	152.5	Lower Hoyon (260); Middle Hoyon (350)
	59-65	472	78.7	Seca (272); Lower Hoyon (200)	733	122.2	Seca (397); Lower Hoyon (336)
	65-70	478	95.6	Cimarrona (347); Seca (131)	no data	-	
Western axial domain							
Paipa	49-55				160	26.7	Bogotá
1	55-59				325	81.3	Bogotá
	59-65				322	53.7	Guaduas (122); Cacho (200)
	65-70				520	104.0	Guaduas
Tunia	49-55				70	11.7	Bogotá
	55-59				130	32.5	Cacho (30): Bogotá (100)
	59-65				224	37.3	Guaduas (100); Cacho (124)
	65-70				500	100.0	Guaduas
Checua	49-55	278	46.3	Bogotá			
	55-59	575	143.8	Bogotá			
	59-65	498	83.0	Guaduas (253); Cacho (245)			
	65-70	655	131.0	Guaduas			

**Table 2.** Estimated stratigraphical thickness for chronostratigraphical intervals in each study area (see Fig. 3 for an explanation of the methods) and the calculation of sedimentation rates ( $m Ma^{-1}$ ) shown in Figure 8. Italics and roman fonts correspond to the interval of 65–70 Ma in each area. Numbers in parentheses in the lithological units column make reference to stratigraphic thickness in metres

Macheta	49–55 55–59 59–65				278 368 334	46.3 92.0 55.7	Bogotá Bogotá Guaduas (250); Cacho (84)
Guatavita	65–70 49–55 55–59 59–65				419 708 472 340	83.8 118.0 118.0 56.7	<i>Guaduas</i> Bogotá Bogotá Guaduas (170); Cacho (170)
Usme	65–70 49–55 55–59 59–65	693 526 325	115.5 131.5 54.2	Bogotá Bogotá Guaduas (225); Cacho (100)	280	56.0	Guaduas
	65-70	375	75.0	Guaduas			
<b>Eastern axial domain</b> Paz de Rio	49-55				90	15.0	Upper Socha (44); Picacho (46)
	55–59 59–65				188 185	47.0 30.8	Upper Socha Lower Socha (138); Upper Socha (47)
Pesca	65–70 49–55				<i>400</i> 130	80.0 21.7	<i>Guaduas</i> Upper Socha (73); Picacho (57)
	55–59 59–65				178 125	44.5 20.8	Upper Socha Lower Socha
Rondon	65–70 49–55	163	27.2	Upper Socha (130); Bicacho (33)	232	46.4	Guaduas
	55-59	265	66.3	Lower Socha (40); Upper Socha (225)			
	59-65	160	26.7	Lower Socha			
San Antonio	65–70 49–55	236	47.2	Guaduas	291	48.5	Upper Socha (225); Picacho (66)

(Continued)

Domain	Chronostratigraphical interval	Wester	n flank	Lithological units	Eastern	n flank	Lithological units
	(Ma)	Thickness (m)	Rate (m Ma <sup>-1</sup> )		Thickness (m)	Rate (m Ma <sup>-1</sup> )	
	55-59				280	70.0	Lower Socha (40); Upper Socha (240)
	59-65				219	36.5	Lower Socha
	65-70				487	97.4	Guaduas
Umbita	49-55				250	41.7	Upper Socha (196); Picacho (54)
	55-59				245	61.3	Lower Socha (41); Upper Socha (204)
	59-65				214	35.7	L Socha
	65-70				280	56.0	Guaduas
Eastern domain							
Cusiana	49-55	54	9.0	Mirador			
	55-59	128	32.0	Cuervos			
	59-65	86	14.3	Barco (56); Cuervos (30)			
	65-70	0	0.0				
Medina	49-55	96	16.0	Mirador	0	0.0	
	55-59	477	119.3	Cuervos	261	65.3	Cuervos
	59-65	231	38.5	Barco	274	45.7	Barco (187); Cuervos (87)
	65-70	50	10.0	Guaduas	53	10.6	Guaduas

Table 2. Continued

strata is coeval in the eastern and axial domains, whereas, in the western domain, onset of finegrained deposition of the middle Hoyón Formation is younger than in the axial domain.

The Palaeocene–Eocene boundary is well constrained by palynology and detrital zircon ages in the four domains (Fig. 2b), showing a correlation, from west to east, of the middle Hoyón–Bogotá– Upper Socha–Cuervos formations (Fig. 2b). Lower Eocene strata in the eastern domain and eastern axial domain include sandstones of the Mirador– Picacho formations, whereas, in the western axial domain, it corresponds to the fine-grained strata of the upper Bogotá Formation. In the western domain, the middle Hoyon Formation in the south and strata mapped as upper Seca Formation in the north have an inferred mid-Palaeocene–lower Eocene age (Fig. 2b).

### Provenance

The description of provenance signatures (sandstone petrography, heavy minerals and U–Pb detrital zircon geochronology) follows the chronostratigraphical intervals defined earlier to understand the dispersal of terrigenous detritus by interval of time (see Figs 4–6). Glauconite was identified either as authigenic (e.g. clean faces filling pores: Fig. 4e) or as reworked fragments (clay coating and rounded fragments: Fig. 4f, g).

### Upper Campanian sandstones (>70 Ma)

Quartzarenites show authigenic glauconite, as identified at the top of the Guadalupe Formation in the Umbita (Fig. 4e) and Checua areas (e.g. Sarmiento 1992). Heavy mineral assemblage of the Guadalupe Formation in the eastern domain is dominantly ultrastable (Fig. 5a) but the presence of titanite makes up 11.7% of the sample (Fig. 5b).

Detrital zircons for Campanian strata from the axial and eastern domains reveal a dominant Proterozoic population, with several age peaks between approximately 0.96 and 1.8 Ga (see also the Cretaceous rock data in Horton *et al.* 2010*b* and Saylor *et al.* 2011), whereas one sample from the western domain additionally reveals a younger population (< 0.3 Ga) (Fig. 6). In these three areas, Palaeozoic age populations are a minor population.

## Maastrichtian sandstones (65-70 Ma)

The provenance signature of Maastrichtian sandstones in the axial and eastern domains fall towards the field of craton interior, whereas sandstones of the western domain fall in the field of transitional recycled orogen (Fig. 4a). However, lithic fragments are dominantly sedimentary in all of the four domains. Reworked glauconite was identified in five areas of the axial domain (Checua, Paipa, Úmbita, Rondon and Pesca) (Fig. 4f, g). Heavy mineral assemblage is dominantly ultrastable (Fig. 5a), with stable minerals in the western domain (apatite and garnet, Guaduero area: Fig. 5c) and enrichment of micas in northern areas of the axial zone (Rondon and Paipa).

The northern segments of the western axial and eastern axial domains are characterized by the appearance of Phanerozoic detrital zircon populations, including Cretaceous (c. 70–80 Ma) and Silurian–Cambrian (c. 424–534 Ma), as well as Proterozoic zircons similar to the population of Cretaceous strata (Figs 6 & 7). Cretaceous ages (65–100 Ma) are the dominant age population in the western domain. In the Umbita area, the age population is older than 1.2 Ga, and a single Guaduas Formation sample that shows ages older than 0.9 Ga (Saylor *et al.* 2011).

# Lower-middle Palaeocene sandstones (59-65 Ma)

The provenance signature defined by the sandstone petrography of the eastern and western domains varies in the content of the amount of quartz and lithic fragments (craton interior and transitional recycled orogen, respectively), whereas provenance signature of sandstones of the two axial domains overlap in the field of recycled quartzose orogen (Fig. 4b). Total lithic fragment content increases westwards (Fig. 4b, h-j). Metamorphic and plutonic rock fragments increase with respect to Maastrichtian sandstones; these fragments, volcanic lithic fragments and feldspars are more abundant in the western domain and the eastern axial domain (Fig. 4b). Reworked glauconite fragments were identified in the Checua and Usme areas of the western axial domain. Unstable and stable heavy mineral assemblages are more abundant in the western domain and in the Usme area, whereas, in the other areas of the axial zone and the eastern domain, ultrastable minerals dominate (Fig. 5a). Heavy minerals concentrate in laminae in the Umbita area, and in Pesca and Paipa muscovite is very common. Garnets fragments were identified in the Umbita area (eastern axial domain: Fig. 5d), while pyroxene fragments were identified in San Juan de Río Seco (western domain: Fig. 5e) and Umbita (eastern axial domain: Fig. 5f).

Cretaceous detrital zircons, together with more limited Triassic (248–230 Ma), Early Palaeozoic and Proterozoic detrital zircons, dominate in the western domain and the Usme area (Figs 6 & 7), as well as in the Paz de Río area (eastern axial

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Fig. 4. (a)–(d) Ternary provenance and lithical population diagrams for each chronostratigraphical interval, showing the calculated mean value. The shaded polygons represent  $1\sigma$  uncertainty envelopes. For lower Eocene sandstones (Fig. 4d), samples of the quartzose Picacho and Mirador formations are plotted separately because those units are above an unconformity. Note the upsection increase in lithical content in all domains, and the enrichment of metamorphic lithical fragments in late Palaeocene time for the western and eastern domains. (e) Authigenic glauconite. (f) & (g) Reworked glauconite and ferruginous matrix reported in Maastrichtian strata of the axial domain. Areas with reworked glauconite are indicated for each interval of time. (h)–(j) Eastwards increase in sandstone maturity of lower–middle Palaeocene sandstones, indicating at least three provenance fields (see b) from transitional recycled orogen to the west, to quartzose recycled orogen in the axial zone, and to craton interior to the east.

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**Fig. 5.** (a) Lateral distribution of heavy minerals in Maastrichtian–lower Eocene sandstones. (b)–(h) Photomicrographs of titanite, garnet, pyroxene and hornblende as examples of unstable and stable minerals reported at different domains and at different age intervals; the presence and good preservation of these minerals indicate provenance from metamorphic and igneous rocks, and accumulation in a first-cycle sandstone. (i) Muscovite fragments are very common in northern areas of the axial zone; this mineral is an indicator of accumulation in low-energy settings.

domain: Fig. 6) (Saylor *et al.* 2011). Although Phanerozoic zircon populations are similar along the western axial zone, there is a northwards increase of Proterozoic (0.8-2 Ga) age populations (Fig. 7). In contrast, one sample from the eastern domain shows older zircon age peaks of 600, 900 and 1500 Ma (Fig. 6).

#### Upper Palaeocene sandstones (55–59 Ma)

The provenance signature of the four studied domains plot near the boundary of quartzose and transitional recycled orogens, being more lithic and feldspar-rich in the western domain (Fig. 4c). The content of total lithic fragments increases slightly with respect to lower-middle Palaeocene sandstones. Metamorphic-plutonic rock fragments, as well as feldspar fragments, are more abundant in western, eastern and western axial domains, whereas sedimentary rock fragments dominate in the eastern axial domain (Fig. 4c). Reworked glauconite was identified in the Usme (western axial domain) and Umbita (eastern axial domain) areas. Unstable heavy minerals were found in the Usme (epidote), Rondón (chlorite) and Medina (hornblende: Fig. 5g) areas, and in the former two areas the unstable population is an important constituent



**Fig. 6.** Detrital zircon age population distributed by age interval and structural domain. Note the dominance of the young age population (65-360 Ma) in the western domain since the Campanian, while this population is absent in the eastern domain. Only Palaeocene ages of volcanic zircons are reported in the eastern domain. In the western axial zone, the dominance of 65-360 Ma increases in lower-middle Palaeocene sandstones, whereas, in the eastern axial zone, this population increases in upper Palaeocene sandstones. Palaeozoic and Proterozoic age populations dominate in the eastern axial and eastern domains in all of the intervals, and in the western axial domain in upper Cretaceous rocks, and lower-middle Palaeocene rocks of the northern areas (see Fig. 7). Archean ages (>2500 Ma) are absent in all of the study areas.

(Fig. 5a). In the other areas of the western, axial and eastern domains, the ultrastable assemblage dominates (Fig. 5a). Micas continue as important constituents in northern areas (Paipa–Pesca) of the axial zone. Garnet was identified in the Guaduero (western domain) and Medina (eastern domain) areas. The detrital zircon age populations from the western domain are dominantly Mesozoic (90–250 Ma: Fig. 6), whereas samples from the western axial and eastern axial domains show an increase in Phanerozoic detrital zircon ages (65-360 Ma) and decrease in Proterozoic (>1000 Ma) zircon ages when compared with the lower-middle Palaeocene

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65-360 Ma (Central Cordillera source ages) 360-1300 Ma (Eastern Cordillera Basement source ages) >1300 Ma (Amazon Craton source ages)

**Fig. 7.** Detrital zircon age populations of lower-middle Palaeocene sandstones from the western axial zone, showing the northwards increase in 900–1800 Ma ages. The increment of this age population is related to the reworking of Cretaceous rocks, as the exposure of these units increases northwards (see Fig. 1).

samples (Fig. 6). In contrast, the eastern domain shows a bimodal distribution with a minor population of 50-65 Ma zircons and a prominent Proterozoic population (>1.5 Ga) (Fig. 6).

#### Lower Eocene sandstones (49-55 Ma)

The provenance signature of the Mirador (eastern domain) and Picacho (eastern axial domain)

sandstones plot in the upper boundary of quartzose recycled orogen, whereas the other sandstones of the western axial and eastern axial domains (Bogotá–Upper Socha formations) plot near the boundary of quartzose and transitional recycled orogens (Fig. 4d). The former two units are above the unconformity, while the latter two units are below the unconformity. The content of metamorphic–plutonic and sedimentary rock fragments is

proportional in these sandstones, with an increase in the content of the latter fragments in the Picacho sandstones (Fig. 4d). The highest content of feldspar fragments is in the western axial domain. Reworked glauconite was identified in the Usme (western axial domain) and Umbita (eastern axial domain) areas. Stable and unstable heavy mineral assemblages are common both in the Usme (garnet in Fig. 5h: epidote and hornblende) and Umbita areas (garnet, titanite, epidote and hornblende), with a dominance of micas in the Rondón area (Fig. 5a, i). The ultrastable assemblage becomes dominant in quartzose sandstones of the Picacho and Mirador formations, above the unconformity (Fig. 5a); however, in the Medina area (eastern domain), apatite, hornblende, pyroxene and titanite are present at levels of between 2 and 5%.

Detrital zircon age populations from the western axial domain mostly include those of less than 290 Ma, with significant 58–70 and 220–290 Ma age peaks. Whereas, in the eastern axial domain, similar Phanerozoic detrital zircon age populations alternate with more abundant Proterozoic (>1000 Ma) zircon ages, with peak ages as old as about 1.8 Ga (Fig. 6). Samples from the Upper Socha Formation, reported in Saylor *et al.* (2011), include zircons in the range of 53–57 Ma. In contrast, detrital zircon populations reported for one sample of the Mirador Formation in the eastern domain (Horton *et al.* 2010*a, b*) is characterized by zircons older than 1.4 Ga.

# Spatial and temporal changes in sedimentation rates

Division of stratigraphical units (thickness and lithologies) into chronostratigraphical intervals (age) enable sedimentation rates to be calculated (see Table 2 for a summary). In this section, we document the spatial trend of sedimentation rates for each interval of time in order to locate depocentres, and infer its three-dimensional geometry in map and profile view (Figs 8 & 9). We use a sedimentation rate of 80 m Ma<sup>-1</sup> as an arbitrary reference to enclose areas with relatively high sedimentation rates as depocentres.

## Maastrichtian depocentres (65-70 Ma)

Two depocentres with localized areas of maximum sedimentation rates were formed at this time. In the western domain, a localized depocentre formed around the San de Juan de Río Seco area (95 m  $Ma^{-1}$ ). Sedimentation rates decrease abruptly further north in the Guaduero area (3.4 m  $Ma^{-1}$ ), and we infer that sedimentation rates of the depocentre increased eastwards (i.e. become thicker eastwards: Fig. 9b). Maastrichtian strata thin or are eroded by an unconformity further to the north (Río Minero area: Restrepo-Pace *et al.* 2004) and are not recorded by an unconformity further south (Fusa Syncline: Bayona *et al.* 2003). In contrast, the depocentre in the axial domain has an elongated geometry with a NNW strike (Fig. 8a). A maximum sedimentation rate of 131 m Ma<sup>-1</sup> is documented in the Checua area; other values decrease in all directions as far as the eastern domain, where Maastrichtian strata are not preserved (Fig. 8a). The profile geometry of the axial depocentre is like asymmetrical sag that is thicker to the west.

# *Lower–middle Palaeocene depocentres* (59–65 *Ma*)

Maximum sedimentation rates of  $122 \text{ m Ma}^{-1}$  in the western domain are located in the East San Juan de Río Seco region and decrease northwards to the Guaduero area  $(58-76 \text{ m Ma}^{-1})$ , suggesting a north-striking elongated depocentre geometry (Fig. 8b) that is thicker eastwards (Fig. 9c). In contrast, the former Maastrichtian depocentre in the axial domain broke into two depocentres. In the western axial domain, the Checua area continues as the area with a maximum sedimentation rate of  $83 \text{ m Ma}^{-1}$ ; sedimentation rates decreases slightly southwards to  $55-57 \text{ m Ma}^{-1}$  and more abruptly towards the eastern axial domain, where sedimentation rates are between 30 and 36 m  $Ma^{-1}$ . In the eastern domain, a new depocentre began to form as sedimentation rates increased to  $38-45 \text{ m Ma}^{-1}$ . The profile geometry of the axial depocentre continues as an asymmetrical sag that is thicker westwards, whereas the geometry of the eastern domain depocentre has a wedge-like geometry that is also thicker westwards (Fig. 9c).

## Upper Palaeocene depocentres (55–59 Ma)

During late Palaeocene time, the depocentre in the western domain had a similar pattern to that described before, with maximum sedimentation rates in the East San Juan de Rio Seco area  $(152 \text{ m Ma}^{-1})$  and a profile geometry that is thicker eastwards (Fig. 9c). The other two depocentres described for the axial and eastern domains, respectively, have a significant increase in sedimentation rates in comparison with rates reported in lower–middle Palaeocene time (Fig. 8c), but the profile geometry remains similar. In the axial domain, sedimentation rates increase to  $92-143 \text{ m Ma}^{-1}$  in the Checua–Guatavita areas, whereas, further south, they reached 131 m Ma<sup>-1</sup> in the Usme area. In other areas further north of the axial domain, sedimentation rates

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**Fig. 8.** (a)–(d) Palaeogeographical maps for the four chronostratigraphical intervals discussed in the text, showing the location of source areas, active structures, geometry of depocentres, sedimentation rates, depositional environments and dispersion of terrigenous detritus. (e) Palinspastic map for the Palaeogene time of Sarmiento-Rojas (2001), used as base map for the location of studied areas shown in (a)–(d). These diagrams show the northwards withdrawal of marginal deposits and how depositional systems of the western domain have been disconnected since Maastrichtian time. (b) & (c) During the Palaeocene, a single depositional system broke into two depositional systems due to eastwards migration of intraplate reactivation. The depositional system in the axial zone shows the presence of low-energy settings (i.e. continental lakes); further north, this depositional system is closed by the uplift of the Santander Massif at this time (Ayala *et al.* 2012). In contrast, the depocentre formed in the eastern domain shows northwards dispersion of terrigenous material. Depocentres with sedimentation rates in excess of 80 m Ma<sup>-1</sup> in the axial domain formed eastwards of reactivated faults. (d) In early Eocene time, magmatic activity continues along the Central Cordillera and advances onwards to intraplate settings (Fig. 9c) (Bayona *et al.* 2012); localized depocentres formed in the eastern axial domain and in the western domain, and a low tectonic subsidence regime dominated in the eastern axial and eastern domains.



**Fig. 9.** (a)–(c) Crustal-scale cross-sections for 75, 65 and 55 Ma (see Fig. 8a, c for the location) showing the relationship between Caribbean subduction, marginal and intraplate magmatism (from Bayona *et al.* 2012), crustal tilting, fault reactivation, and geometry of intraplate syn-orogenic basins. Tectonic subsidence mechanisms vary eastwards from crustal tilting of the Central Cordillera to reactivation of intraplate structures. Both mechanisms are related to subduction of the Caribbean Plate. In early Eocene time, reactivation in intraplate structures in both margins of the Eastern Cordillera formed a sag-like basin bordered by exposed Cretaceous rocks; deposition in the Llanos Basin corresponds to filling of a foreland basin (Bayona *et al.* 2008). (d) & (e) Regional tectonic setting, showing the shift in the style of collision of intraplate structures (modified from Bayona *et al.* 2012).

range between 47 and 81 m  $Ma^{-1}$ . In the depocentre of the eastern domain, sedimentation rates reached 119 m  $Ma^{-1}$  for the Medina area.

### Lower Eocene depocentres (49–55 Ma)

Only two depocentres formed at this time. The depocentre in the western domain maintains its location in the San Juan de Rio Seco area (up to  $138 \text{ m Ma}^{-1}$ ), with a wedge-like geometry thickening eastwards. As in previous intervals of time, sedimentation rates decrease abruptly northwards to less than  $12 \text{ m Ma}^{-1}$ . The other depocentre formed in the southern western axial domain, in the Usme (115 m Ma<sup>-1</sup>) and Guatavita (118 m Ma<sup>-1</sup>) areas. Further north and east, sedimentation rates decrease to values ranging from 15 to 45 m Ma<sup>-1</sup> in the eastern domain and from 0 to 16 m Ma<sup>-1</sup> in the eastern domain.

### Discussion

Integration of provenance data, spatial and temporal variations in sedimentation rates, and location of depocentres, using the systematic analysis carried out along the 16 studied sections (Fig. 1c), allows us to propose a new model for the tectonostratigraphical evolution of Maastrichtian–lower Eocene syn-orogenic basins, and to define how these syn-orogenic basins formed by the interaction of the Caribbean and South American plates. In this paper, we compare our model with pre-existing tectonostratigraphical models, suggesting either foreland basin setting or negative flexural basin for the early Palaeogene period.

# Maastrichtian-early Eocene tectonostratigraphical model and its relationship to Caribbean tectonics

Provenance signatures of Maastrichtian-middle Palaeocene sandstones (Fig. 4a, b) show at least two different source areas supplying sediments to the western and the axial domains. Supply from basement rocks of the Central Cordillera is well documented in both domains by the presence of a 65–360 Ma detrital zircon population, metamorphic lithic fragments, as well as stable and unstable heavy mineral associations. The emergence of the Central Cordillera in latest Cretaceous-early Eocene has been also documented using thermochronological and provenance data (Gómez *et al.* 2003; Villagómez *et al.* 2011; Caballero *et al.* 2013).

The other source area should be represented by the Cretaceous sedimentary cover separating the western domain from the axial domain, as indicated by the following criteria.

- The presence of reworked glauconite (Fig. 4f, g) and Cretaceous pollen in different areas of the axial domain (Fig. 8a, b).
- An increase in muddy matrix in the fabric of Palaeocene sandstones (Fig. 4f, g).
- A contrasting upsection change in heavy mineral assemblages. In the western domain, stable and unstable minerals are more common in lower-middle Palaeocene sandstones, and ultrastable minerals become dominant in upper Palaeocene sandstones (Fig. 5a). In the axial domain, stable and ultrastable heavy mineral associations dominate with a localized increase in unstable minerals in the Usme area in upper Palaeocene sandstones.
- A northward enrichment in 900–1800 Ma detrital zircon age population in sandstones of the axial zone (Fig. 7).

The criteria presented above also indicate that the uplifted area must have been located close to the place of deposition to the axial sections, as pollen and glauconite are very unstable fragments to survive tropical weathering and more than 40 km of fluvial transport, as documented in the fluvial sands of the Llanos Basin of Colombia (Amorocho *et al.* 2011). The reworking of compositionally mature Cretaceous sandstones concentrates ultrastable heavy minerals. The muddy fraction may be a product of the erosion of palaeosols (Potter *et al.* 2005), and 900–1800 Ma detrital zircon ages are a very common population in Cretaceous rocks (Horton *et al.* 2010*a, b*).

Separation of the western domain from the axial domain by the uplift of the Cretaceous sedimentary cover is also suggested by the eastwards increase in thickness in the western domain, and high sedimentation rates reported in the San Juan de Rio Seco area. Tectonic loads eastwards of the Guaduero and San Juan de Río Seco sections favoured higher tectonic subsidence of the western domain (Fig. 9a, b). Those tectonic loads correspond to the inversion of former normal faults of the western margin of the former extensional basin (Figs 8a & 9b), as proposed by Sarmiento-Rojas (2001) and Bayona et al. (2008), who used geodynamic models, and Cortés et al. (2006), who used the surface and subsurface mapping relationships of Palaeogene units (Fig. 1).

Uplift of the western flank of the Eastern Cordillera was not continuous southwards because continental deposits in the western domain further south of the San de Rio Seco area were connected with a fluvial system in the Usme area, as suggested by similar detrital zircon populations between these two areas (Figs 6 & 8a, b). Northwards and eastwards dispersion of detritus supplied from the Central Cordillera in the axial domain is further supported by palaeocurrent indicators reported for Palaeocene sandstones in the axial domain (see Bayona *et al.* 2008 for a review and Saylor *et al.* 2011 for palaeocurrent data in the Paz de Río area).

Reactivation of the western margin of the former extensional basin may be also explained by crustal tilting of the Central Cordillera–Santa Marta Massif (Fig. 9a, b), an uplift mechanism of the continental margin related to the subduction of the Caribbean Plate (Bayona *et al.* 2011, 2012; Cardona *et al.* 2011; Ayala *et al.* 2012). The presence of: (1) metamorphic rock fragments; (2) garnet, apatite and epidote fragments; and (3) the dominance of Cretaceous and Permo-Triassic zircon age populations over Jurassic age populations support the erosion of Cretaceous intrusive rocks and Permo-Triassic metamorphic rocks presently exposed adjacent to the Romeral palaeosuture.

Reactivation of the western fault system of the former extensional basin acted as the eastern boundary of the tilted rigid crustal block (Fig. 9). Eastwards thickening of the depocentre in the western domain (Figs 8 & 9) is also explained by eastwards tilting of the Central Cordillera. Similarly, eastwards tilting of the inverted hanging-wall block (Figs 9b, c) explains the elongated depocentre documented in Maastrichtian-late Palaeocene time (Fig. 8a-c) in the western axial domain. Sedimentation rates in the western axial depocentre are equivalent to, or higher than, the rate determined in the eastern San Juan de Rio Seco area in the western domain. The simultaneous eastwardstilting activity of uplifted blocks may explain the similar sedimentation rates in two areas that are 130 km apart (Fig. 8). However, local structurecontrolled patterns of sedimentation in the western domain (e.g. normal faulting in Fig. 8c) controlled localized deposition of fine-grained strata of the middle Hoyón Formation.

The distance between the Central Cordillera and the eastern axial and eastern domains is greater than 250 km (Fig. 8). Therefore, chemically unstable fragments are not expected to survive hundreds of kilometres of fluvial transport in tropical settings (Johnsson *et al.* 1991; Amorocho *et al.* 2011). Reworked Cretaceous pollen, unstable heavy mineral fragments (Fig. 5: garnet, titanite, hornblende, epidote and chlorite) and a high content of lithic fragments are common features in upper Palaeocene–lower Eocene sandstones of the eastern axial and eastern domains (Fig. 4c, d). Therefore, unstable terrigenous detritus in the eastern axial and eastern domains were not entirely derived from the Central Cordillera. In addition, a characteristic detrital zircon population derived from the Central Cordillera (65–360 Ma) is not recorded in the eastern domain (Fig. 6).

In the late Palaeocene, mud-dominated deposition and an increase in the content of metamorphic lithic fragments in sandstones of the eastern domain began earlier than in the axial domain (Figs 2b & 4c). Unstable titanite and hornblende fragments in Campanian and upper Palaeocene strata (Fig. 5a) also indicate the presence of nearby located basement-cored uplifts. The detrital zircon age population of the upper Palaeocene strata in the eastern domain (Fig. 6) is similar to the zircon population older than 1.2 Ga reported from a buried basement-cored high in the proximal Llanos Basin (Ibañez-Mejia et al. 2009). These buried basement-cored highs, presently under the Cenozoic foreland strata, include Cretaceous and/or Palaeozoic sedimentary units overlying basement rocks (Bayona et al. 2007). Therefore, erosion of the sedimentary cover and basement-cored uplifts supplied quartzose detritus, metamorphic lithic fragments, unstable heavy minerals and detrital zircons older than 1.2 Ga.

Reactivation of the eastern normal fault system of the former extensional basin and the development of structures in the southern Llanos Basin occurred during the Palaeocene (see Mora et al. 2013). This deformation is evident in late Palaeocene time by the immature composition of sandstones (Fig. 4c), and sedimentation rates in the eastern domain are higher than those in the eastern axial domain (Figs 8c & 9c). The location of the depositional axis in the eastern domain during the Maastrichtian (Figs 8a & 9b) and the reactivation of the eastern normal fault system preclude the eastwards transport of detritus supplied from the Central Cordillera to the eastern domain (Fig. 6). Tectonic loading between the western and eastern domains contributed to the flexure-driven subsidence and foreland pattern of the deposition between the eastern axial domain and the proximal Llanos Basin in Maastrichtian-late Palaeocene time (Fig. 9b, c) (Bayona et al. 2008).

Fault reactivation along the eastern margin is coeval with the record of volcanic zircons in upper Palaeocene sandstones (Bayona *et al.* 2012). These two processes may be explained by shallow eastwards-dipping subduction of the Caribbean Plate (Figs 9c, d) that produced intraplate magmatism and fault reactivation located as far east as the proximal Llanos Basin. Such intraplate magmatism affecting the Eastern Cordillera and adjacent basins in the east has been reported for Cretaceous (Vasquez *et al.* 2010) and Mio-Pliocene periods (Taboada *et al.* 2000; Vasquez *et al.* 2009).

## Comparison to pre-existing tectonic models

A major difference between our study and preexisting tectonostratigraphical model studies of the Northern Andes is the systematic analysis along 13 studied sections distributed across the four structural domains, and the integration of data from other studies along three sections. Some previous studies focused on data collected from only a single area (e.g. Paz de Río area: Saylor *et al.* 2011, 2012; Medina area: Parra *et al.* 2009*a*) or from a single structural domain (Middle Magdalena Valley: Gómez *et al.* 2003, 2005*a*, *b*). Other ones included a regional study with no details of specific sections (Cooper *et al.* 1995; Pindell *et al.* 2005) or used results from only one technique, such as thermochronology (Mora *et al.* 2010).

The model of a continuous foreland basin model, which lasted either from Maastrichtian to middle Miocene (Cooper *et al.* 1995) or from Maastrichtian to late Eocene (Gómez *et al.* 2005*a*; Horton *et al.* 2010*b*), predicts the following phenomena, which are not supported by our data (see discussion above) or by recently published data, including the following.

• Burial of Upper Cretaceous rocks was homogeneous across the western margin of the former extensional basin. The continuous foreland basin model implies that the thickest Palaeogene deposition occurred in the area where we propose the onset of inversion of Mesozoic structures (western margin of the former extensional basin). Moretti et al. (2010) used vitrinite reflectance as a palaeothermometer to compare burial processes in the footwall and hanging wall of the Bituima Fault. Their results show an eastwards decrease in maturity of the Umir Formation, and they interpreted an early Cenozoic reactivation of the Bituima-El Trigo Fault zone, similar to the suggestion by Cortés et al. (2006). Published thermochronometry studies indicate that exhumation of the western flank of the Eastern Cordillera is younger southwards. Parra et al. (2012) and Caballero et al. (2013) document the onset of exhumation in the Palaeocene of the western flank of the Eastern Cordillera and northern Magdalena Basin. Exhumation ages in the hanging wall of the Bituima Fault, further to the south, ranges from 41 to 21 Ma (Gómez et al. 2003; Parra et al. 2009b; Caballero et al. 2013) and from 50 to 35 Ma for the western limb of the Guaduas Syncline (Parra et al. 2009b). Although thermochronometry does not reveal early Palaeogene onset of cooling, provenance signatures and sedimentation rates record those early Palaeogene phases of deformation, as shown in this study. In order to test the hypothesis proposed in our study, it is necessary to carry out detrital thermochronology analysis in Eocene and younger rocks in the Guaduas Syncline and Checua area.

- An east-verging thrust belt (Central Cordillera) bounding a syn-orogenic basin formed by flexural subsidence. In the foreland basin model, it is expected that the frontal segment of the Central Cordillera supplied sediments to the basin, with an expected dominance of Jurassic zircon age populations, feldspar and volcanic lithic fragments (Fig. 9a). However, heavy mineral analysis and detrital zircon age populations (see data above) indicate that Cretaceous intrusive rocks and Permo-Triassic metamorphic rocks, presently exposed adjacent to the Romeral palaeo-suture, were the dominant source rocks (Fig. 9a). Therefore, the western margin of the Central Cordillera was exposed, whereas the eastern flank of the Central Cordillera was not exposed at this time. Our data support the model of an eastwards-tilting Central Cordillera with an apparent absence of bounding thrusts (Pindell et al. 2005) as result of Caribbean subduction. The model proposed by Pindell et al. (2005) describes the syn-orogenic basin as a 'sag-like' structure with a negative flexure mechanism. The irregular eastwards variation of thickness and sedimentation rates in the Palaeocene period (Fig. 8: Tables 1 & 2) supports neither a positive flexural foreland basin nor a single elongated depocentre expected in the negative flexural model.
- A single magmatic arc. Saylor et al. (2012) supported the foreland basin model for the Palaeocene time because they interpreted that Palaeocene volcanic zircons reported in the Paz de Rio area were supplied from the Central Cordillera and transported by an east- to northeastward-directed drainage. In this paper, we document that early Palaeogene volcanism affected both the Central Cordillera and the intraplate settings (see Bayona et al. 2012). We propose that volcanic activity may be generated as far as in the eastern domain, as indicated by the mixture of Palaeocene volcanic zircons with zircon populations originated in the craton (Fig. 6), and the presence of volcanic rocks in the Usme area. Our palaeogeographical model considers that Palaeocene sediments in the axial zone were supplied by the Central Cordillera and sedimentary cover of inverted structures, following a northeastwards-directed drainage pattern, as documented by palaeocurrent measurements (see the data in Bayona et al. 2008; Saylor et al. 2011) (Fig. 8).

Similar provenance patterns and regular trend of sedimentation rates in the axial domain allows the interpretation that the Boyaca–Soapaga Fault System was not active during Palaeocene–early Eocene time (Fig. 9c). Thermochronological data from samples in the hanging wall of those structures (Mora *et al.* 2010; Saylor *et al.* 2012) document exhumation in late Eocene–Oligocene time, supporting the interpretation of a continuous basin in the axial zone. Why this fault system did not reactivate, in contrast to the faults bordering the extensional basin, needs further investigation.

## Conclusions

The spatial and temporal variation of sedimentation rates and provenance signatures was used to decipher early phases of deformation in a multiphase orogen, such as the Eastern Cordillera of Colombia. Maastrichtian-lower Eocene strata in the western, axial and eastern domains of the central segment of the Eastern Cordillera support a model of eastwards migration of intraplate fault reactivation (positive inversion structures). Diachronous intraplate fault reactivation, development of syn-orogenic basins and intraplate magmatism are associated with the convergence and subduction of the buoyant Caribbean Plate beneath the South American margin (Fig. 9). The reactivation of former Cretaceous faults limited the eastwards tilt of the Central Cordillera, and compressive stresses transmitted to intraplate settings reactivated faults as far in as the proximal Llanos Basin.

Fault reactivation in Palaeocene time produced three independent syn-orogenic basins separated by low-amplitude uplifts with exposure of the Cretaceous sedimentary cover that supplied sediments to an axial depocentre. The Central Cordillera supplied sediments to the western and axial domains, as recorded by detrital zircon population ages younger than 360 Ma, metamorphic rock fragments and garnet-epidote-hornblende heavy mineral assemblage. In late Palaeocene-early Eocene time, depositional systems of the axial depocentre were bordered by structures associated with the inversion of extensional faults, which correspond to the border of the Cretaceous extensional basin. Cretaceous rocks exposed at both margins by inverted structures supplied sediments to the axial depocentre. Maastrichtian-early Eocene rates of sedimentation recorded in sections to the west of the axial depocentre have values similar to the rates recorded in the western depocentre.

In addition to the Cretaceous sedimentary cover, another source area that supplied sediments to the eastern depocentre was a currently buried basement-cored high in the southern Llanos Basin, as indicated by the dominance of a zircon population older than 1.2 Ga (Fig. 6), the immature composition of upper Palaeocene sandstones, and a record of unstable titanite and hornblende fragments. Intraplate magmatism reached areas as far as basementcored uplifts in the Llanos Basin, as indicated by the presence of Palaeocene volcanic zircons with sediments supplied from such intraplate uplifts.

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