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Invited review

Sedimentary record of Andean mountain building

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ABSTRACT

Integration of regional stratigraphic relationships with data on sediment accumulation, provenance, paleodrainage, and deformation timing enables a reconstruction of Mesozoic-Cenozoic subduction-related mountain building along the western margin of South America. Sedimentary basins evolved in a wide range of structural settings on both flanks of the Andean magmatic arc, with strong signatures of retroarc crustal shortening, flexure, and rapid accumulation in long-lived foreland and hinterland basins. Extensional basins also formed during pre-Andean backarc extension and locally in selected forearc, arc, and retroarc zones during Late Cretaceous-Cenozoic Andean orogenesis. Major transitions in topography and sediment routing are recovered through provenance studies, particularly detrital zircon U-Pb geochronological applications, which distinguish three principal sediment source regions—the South American craton, Andean magmatic arc, and retroarc fold-thrust belt. Following the cessation of Late Triassic–Early Cretaceous extensional and/or postextensional neutral-stress conditions, a Late Cretaceous-early Paleocene inception of Andean shortening was chronicled in retroarc regions along the western margin by rapid flexural subsidence, a wholesale reversal in drainage patterns, and provenance switch from eastern cratonic sources to Andean sources. An enigmatic Paleogene hiatus in the Andean foreland succession recorded diminished accumulation and/or regional unconformity development, contemporaneous with a phase of limited shortening or neutral to locally extensional conditions. Seemingly contradictory temporal fluctuations in tectonic regimes, defined by contrasting (possibly cyclical) phases of shortening, neutral, and extensional conditions, can be linked to the degree of mechanical coupling along the subduction plate boundary. Along-strike variations in Late Cretaceous-Cenozoic deformation and crustal thickening demonstrate contrasting high-shortening versus low-shortening modes of Andean orogenesis, in which the central Andes are distinguished by large-magnitude east-west shortening (> 150–300 km) and corresponding cratonward advance of the fold-thrust belt and foreland basin system, several times that of the northern and southern Andes. These temporal and spatial changes in shortening and overall tectonic regime can be related to variable plate coupling during first-order shifts in plate convergence, second-order cycles of slab shallowing and steepening, and second-order cycles of shortening, lithospheric removal and local partial extensional collapse in highly shortened and thickened segments of the orogen.

1. Introduction

Growth of the Andes mountains (Fig. 1) shapes erosion, sediment delivery, river courses, drainage networks, orographic barriers, climate, biodiversity, and ocean circulation patterns across South America and its periphery. Tectonic uplift of the Andes and coupled subsidence of sedimentary basins fundamentally control the distribution of topography and relief in western South America, generating not only the foremost sources of sediment but also the sinks where detrital material is accommodated over geological timescales. Mesozoic-Cenozoic patterns of sediment dispersal, sediment accumulation, river genesis, and drainage reorganization reflect processes of continental landscape evolution associated with protracted subduction and lithospheric deformation along a convergent plate boundary. In particular, the origination and advance of the Andean retroarc fold-thrust belt and foreland basin system (Fig. 1) played a pivotal role in the transition from an extensional or neutral tectonic regime with a westward (Pacific) draining pre-Andean landscape to a contractional regime with an overall eastward (Atlantic) draining landscape.

Despite the applicability of these concepts to the Andes and Andean-type ocean-continent convergent plate boundaries in general, questions persist over (1) the inception of Andean mountain building (e.g., Dalziel, 1986; Coney and Evenchick, 1994; Jordan et al., 2001a; DeCelles and Horton, 2003), (2) spatial variations and possible pulses or lulls in deformation (Rutland, 1971; Gansser, 1973; Aguirre, 1976; Mégard, 1984; Noblet et al., 1996; Horton, 2018), (3) temporal shifts in

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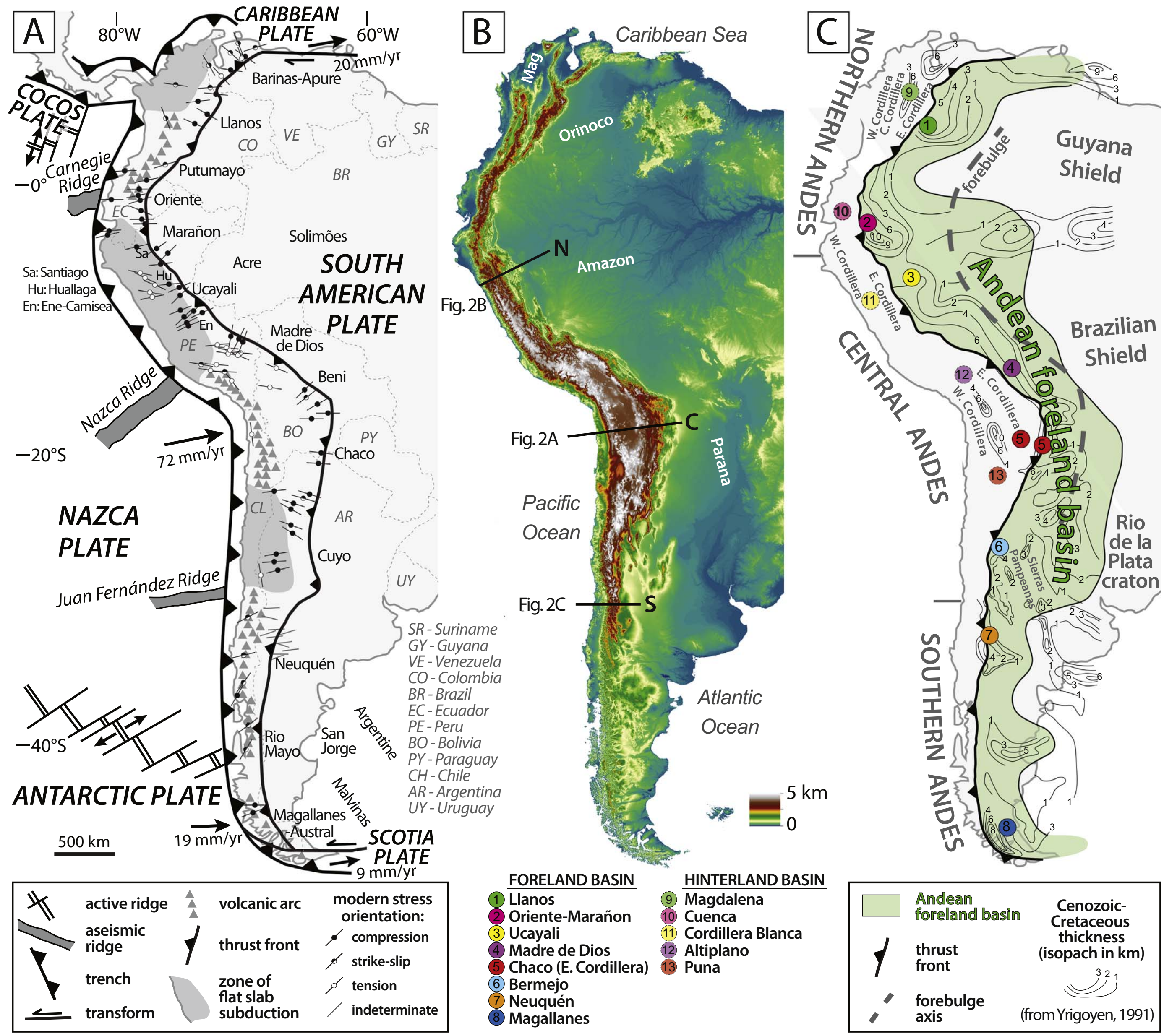

Fig. 1. Maps of (A) tectonic framework, (B) topography, and (C) sedimentary basin configuration of South America. (A) Map of plate boundaries, Andean magmatic arc (including the northern, central, and southern volcanic zones), regions of flat slab subduction, modern stress orientations from earthquake focal mechanisms, eastern front of Andean fold-thrust belt, and key segments of the retroarc foreland basin system. Plate velocities are shown relative to stable South American plate (DeMets et al., 2010). (B) DEM topographic map showing the Andes mountains and adjacent foreland region, including the Amazon, Parana, Orinoco, and Magdalena (Mag) river systems. (C) Map of Andean retroarc basins, showing isopach thicknesses (in km) of Cretaceous-Cenozoic basin fill, forebulge axis (from Chase et al., 2009), and locations of 13 sites (8 foreland basins, 5 hinterland basins) considered in this synthesis.

tectonic regime (Mpodozis and Ramos, 1990; Ramos, 2010; Mpodozis and Cornejo, 2012; Charrier et al., 2015; Horton and Fuentes, 2016), and (4) the history of paleodrainage and sediment accommodation in the basins of western South America (Potter, 1997; Lundberg et al., 1998; Hoorn et al., 2010; Roddaz et al., 2010; Horton et al., 2015a; Anderson et al., 2016). These uncertainties commonly arise from insufficient stratigraphic age control, conflicting basin structural configurations, contrasting modes of sediment accommodation, condensed stratigraphic intervals and unconformities of unclear origin, and/or limited correlation across diverse basin systems. Andean retroarc foreland and hinterland basin systems contain long-lived, chiefly nonmarine clastic successions that reflect sediment accommodation governed by a range of mechanical and thermal processes (Jordan and Alonso, 1987; Sempere et al., 1990; Jordan, 1995; Jacques, 2003; DeCelles, 2012; Horton, 2012; Carlotto, 2013). These synorogenic successions, typically 4–10 km thick, offer windows into the stages of erosional exhumation in the fold-thrust belt and magmatic arc, relative timing of Andean uplift, establishment of major sediment sources, basin subsidence and sediment accumulation patterns, key shifts in landscape development, and the establishment of topographic barriers and large river systems.

This paper integrates a range of stratigraphic, sedimentologic, and geochronological datasets in order to evaluate the Mesozoic-Cenozoic sedimentary record of mountain building in South America. A continental-scale framework spanning the northern, central, and southern Andes (Fig. 1) sets the stage for addressing several interrelated questions.

- When did Andean mountain building commence?
- How did synorogenic basins evolve, and in what structural configurations?

- What were the principal sources of sediment?
- Where were the major zones of sediment accommodation?
- What was the history of sediment accumulation and was it steady or unsteady?
- What were the foremost events in landscape evolution, with respect to drainage reorganization, growth of topographic divides, and genesis of large rivers?
- What were the spatial contrasts in these processes, in terms of along-strike (north-south) and across-strike (east-west) variations?
- Do temporal records indicate steady, continuous processes or intermittent behavior involving punctuated episodes or cyclicity?

This Andean-wide synthesis focuses on the sedimentary record and is necessarily incomplete. Analogous syntheses of Andean structural, magmatic, thermochronologic, paleoelevational, and climatic records are likely to generate complementary insights into these respective topics. Key structural issues include the geometry, style, and magnitude of Andean deformation, as well as the role of inheritance at local and regional scales (e.g., Kley and Monaldi, 1998; Kley et al., 1999; McQuarrie, 2002a; Ramos et al., 2004; Oncken et al., 2006; Giambiagi et al., 2012; McGroder et al., 2015; Perez et al., 2016b). Magmatic records highlight changes in Andean arc activity in relationship to slab dynamics and interactions with continental lithosphere (e.g., Coira et al., 1982; James and Sacks, 1999; Haschke et al., 2006; Kay et al., 1999, 2005; Trumbull et al., 2006; Ramos and Folguera, 2009; Mamani et al., 2010; Folguera and Ramos, 2011; Ducea et al., 2015). Thermochronological data provide timing of exhumational cooling for selected Andean regions (e.g., Thomson, 2002; Barnes et al., 2006; Gillis et al., 2006; Ege et al., 2007; Parra et al., 2009a, 2009b; Mora et al., 2010a, 2010b; Lease et al., 2016; Savignano et al., 2016). Under appropriate conditions, stable isotopes of sedimentary, volcanic, and fossil materials offer broad constraints on past Andean elevations (e.g., Blisniuk and Stern, 2005; Garzione et al., 2006, 2017; Mulch et al., 2010; Insel et al., 2010; Leier et al., 2013; Saylor and Horton, 2014; Canavan et al., 2014; Carrapa et al., 2014; Anderson et al., 2015; Quade et al., 2015; Mulch, 2016; Rohrmann et al., 2016). Complex interactions and feedbacks among Andean tectonics, climate, and erosion are sensitive to variations in precipitation, seasonality, glaciation, slope, lithology, and other related factors in evolving landscapes (Masek et al., 1994; Horton, 1998, 1999; Montgomery et al., 2001; Barnes et al., 2006; Strecker et al., 2007, 2009; Mora et al., 2008; Ehlers and Poulsen, 2009; Thomson et al., 2010; Barnes and Pelletier, 2006; Cruz et al., 2010; Ramirez-Arias et al., 2012).

This review places emphasis on the stratigraphic framework, depositional systems, hiatuses, accumulation patterns, structural configurations, and sediment provenance of basins in western South America. By focusing on the Mesozoic-Cenozoic sedimentary record, this effort provides insights into the coupled history of sediment accumulation and Andean mountain building. It is anticipated that this synthesis will also provide a foundation for ongoing efforts to gauge the timing and mechanisms of surface uplift, informing our current understanding of: (1) the isostatic effects and interconnections among shortening, crustal thickening, magmatic addition, and erosion (Isacks, 1988; Whitman et al., 1996; Horton and DeCelles, 1997; Pope and Willett, 1998; Horton, 1999; James and Sacks, 1999; Pelletier et al., 2010); (2) the potential dynamic processes associated with plate interface stresses, flat slab subduction, asthenospheric slab windows, deep mantle processes, and lithospheric removal (Lamb and Hoke, 1997; Beck and Zandt, 2002; Garzione et al., 2006, 2017; Barnes and Ehlers, 2009; Kay and Coira, 2009; Ramos and Folguera, 2009; Capitanio et al., 2011; DeCelles et al., 2015a, 2015b; Folguera et al., 2015a; Faccenna et al., 2017); and (3) the broader interactions among orography, climate, and biodiversification (Montgomery et al., 2001; Strecker et al., 2007; McQuarrie et al., 2008a; Norton and Schlunegger, 2011; Barnes et al., 2012; Baker et al., 2014). Further, the Andean sedimentary archives described here can be compared with the distal records along the eastern (Atlantic) and northern (Caribbean) margins of South America to assess patterns of transcontinental sediment dispersal, including the onset of huge drainage systems such as the Amazon, Parana, Orinoco, and Magdalena rivers (e.g., Hoorn et al., 2010, 2017; Figueiredo et al., 2009; Sacek, 2014; Anderson et al., 2016).

2. Geologic framework

2.1. Current setting

As the type example of mountain building along an ocean-continent convergent plate boundary, the Andes are genetically related to Mesozoic-Cenozoic subduction of oceanic lithosphere of the Pacific Ocean basin beneath continental lithosphere of western South America (Figs. 1 and 2). Present interactions of the westward advancing South American plate with surrounding oceanic plates involve orthogonal to slightly oblique subduction of the Nazca and Antarctic plates and transform motion relative to the oceanic Caribbean and Scotia plates (Fig. 1A). The Peru-Chile trench defines the western plate boundary of South America, with an actively subducting Nazca/Antarctic oceanic slab of generally moderate (~30°) eastward dip. Zones of shallow to flat-slab subduction (< 10° dip) at 11°–6°N, 2°–15°S, and 27°–33°S correlate with spatial gaps in arc magmatism (between the northern, central, and southern volcanic zones) and the occurrence of oceanic aseismic ridges within the Nazca plate (Barazangi and Isacks, 1976; Cahill and Isacks, 1992; Gutscher et al., 2000; Ramos et al., 2002; Stern, 2004; Ramos and Folguera, 2009).

The Andes (Fig. 1B) constitute an ~8000 km long, 250–750 km wide contiguous topographic barrier with approximate mean elevations of 1200 m in the northern Andes (11°N–5°S), 2500 m in the central Andes (5°–33°S), and 800 m in the southern Andes (33°–55°S) (Gansser, 1973; Auboin et al., 1973; Mégard, 1987, 1989; Orme, 2007; Folguera et al., 2016). The broadly north-trending orogenic belt demarcates the continental drainage divide separating short river systems that flow westward into the Pacific Ocean from those that flow eastward into the Atlantic Ocean and Caribbean Sea, with an internal drainage system characterizing the ~4000 m high central Andean Altiplano-Puna hinterland plateau (14°–27°S) (Isacks, 1988; Allmendinger et al., 1997; Kennan, 2000; Montgomery et al., 2001).

The orogen is divided into tectonic provinces emblematic of Andean-type convergent margins (Figs. 1C and 2), including a forearc region, magmatic arc, retroarc fold-thrust belt, and foreland basin system that closely follow the western trace of the South American coastline (Jordan et al., 1983; Ramos, 1999). For most of the latitudinal extent of the Andes, a narrow western range with a steep forearc flank, generally referred to as the Western Cordillera, marks the active or formerly active magmatic arc. On the opposing eastern slope, a broad sector of rugged ranges referred to as the Eastern Cordillera and Subandean Zone comprise an east-directed retroarc fold-thrust belt exhibiting thin-skinned ramp-flat structures in the upper crust and basement-involved structures in the middle to lower crust (Fig. 2). The lowland provinces flanking the Andes are composed of variably named segments of the narrow forearc region in the west and multiple retroarc basins of the broader foreland basin system in the east (Fig. 1C).

2.2. Geologic history

A protracted history of subduction and arc magmatism along the western edge of South America commenced in Late Triassic-Jurassic time prior to and during breakup of Pangea and westward advance of the South American plate away from Africa (James, 1971; Coira et al., 1982; Aspden et al., 1987; Pankhurst et al., 2000; Pepper et al., 2016; Calderón et al., 2016). After Mesozoic extensional and subsequent neutral (postextensional) tectonic regimes along most segments of the western margin (Dalziel, 1981; Atherton et al., 1983; Uliana and Biddle, 1988; Mpodozis and Ramos, 1990; Salfity and Marquillas, 1994;

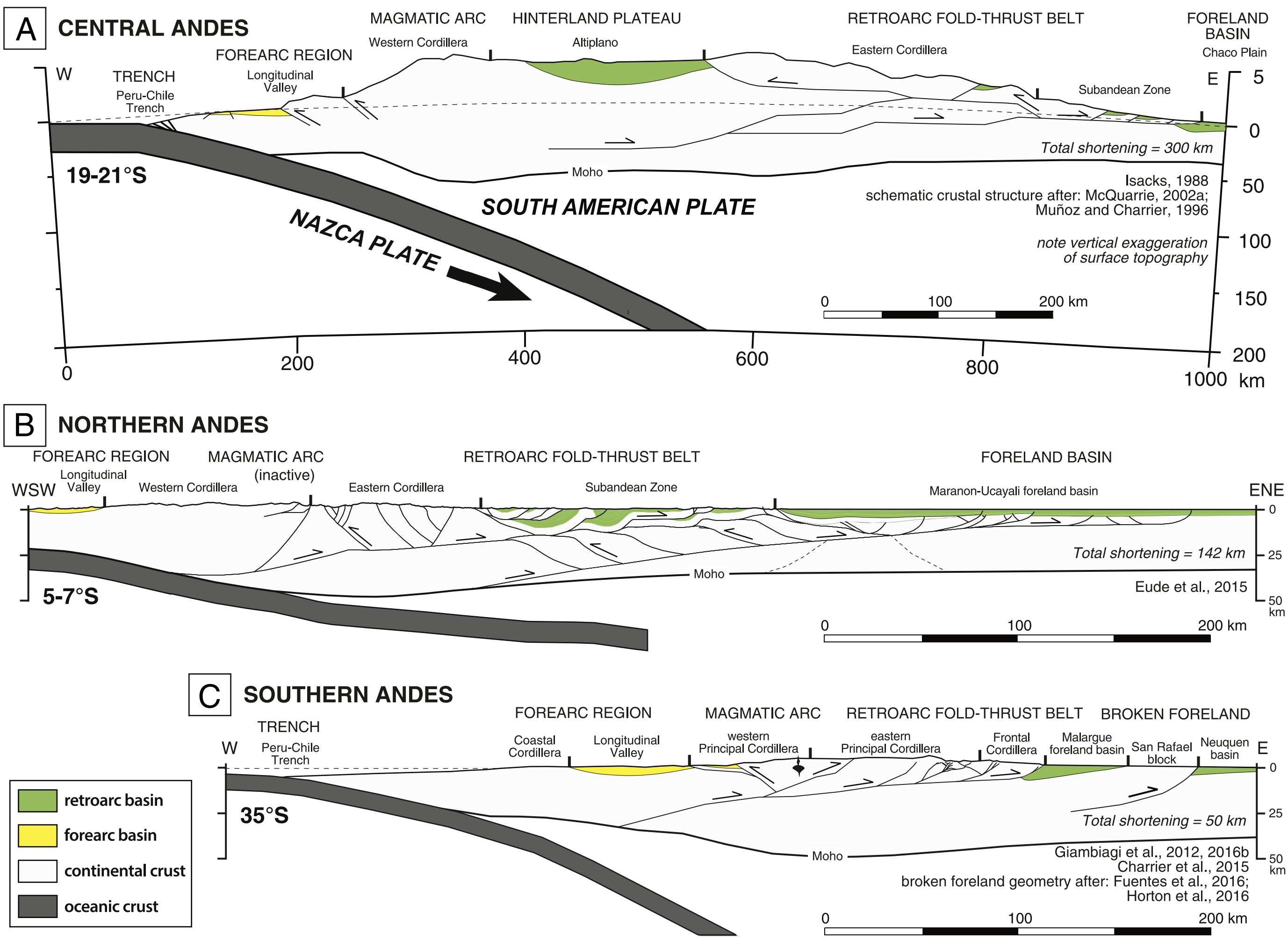

Fig. 2. Cross sections of the (A) central Andes, (B) northern Andes, and (C) southern Andes showing the tectonic configuration along the western margin of South America, including crustal thickness, structural geometries of the Andean fold-thrust belt, retroarc basin configurations, and geometry of the subducted Nazca slab. Map locations of cross sections shown in Fig. 1B. Note the difference in scale and vertical exaggeration of surface topography for the wide, highly shortened central Andean profile.

Jaillard et al., 1990; Cooper et al., 1995), a transition to regional shortening led to construction of the Andean orogenic belt (Wilson, 1991; Ramos, 2010; Horton et al., 2010, 2016; Ghiglione et al., 2016a). Plate tectonic reconstructions show a late Mesozoic increase in absolute westward motion (mantle or hotspot reference frame) of the South American continent (Seton et al., 2012; Maloney et al., 2013; Müller et al., 2016; Wright et al., 2016) and considerable Cenozoic variations in relative convergence rate and direction (Pilger Jr., 1984; Pardo-Casas and Molnar, 1987; Somoza, 1998; Silver et al., 1998; Lonsdale, 2005; Somoza and Ghidella, 2012).

Although the Mesozoic-Cenozoic history is broadly portrayed in terms of pre-Andean extensional or neutral conditions followed by protracted regional compression, additional shifts in tectonic regime likely represent temporal variations in mechanical coupling between the subducting and overriding plate (Sobolev and Babeyko, 2005; Ramos, 2010; Mpodozis and Cornejo, 2012; Horton and Fuentes, 2016; Horton, 2018). Modern observations of upper plate stresses expressed in earthquake focal mechanisms, GPS data, and Quaternary structures indicate Andean-wide compression (Fig. 1A) with local tension in high-elevation areas undergoing gravitational collapse (Suárez et al., 1983; Sébrier et al., 1985; Marrett et al., 1994; Schildgen et al., 2009; Schoenbohm and Strecker, 2009; Giovanni et al., 2010; Veloza et al., 2012; Assumpção et al., 2016; Giambiagi et al., 2016a; Heidbach et al., 2016) and strike-slip faulting principally in forearc regions characterized by strain partitioning associated with oblique plate convergence (Daly, 1989; Dewey and Lamb, 1992; Cembrano et al., 1996; Trenkamp et al., 2002; Rosenau et al., 2006; Nocquet et al., 2014).

Rather than accretion, subduction erosion appears to have been the prevailing process along the western plate interface. Subduction erosion of forearc crust has been inferred on the basis of the eastward advancing magmatic arc, absence of a significant accretionary prism or subduction complex, and coastal exposure of older magmatic arc and continental basement rocks (Rutland, 1971; von Huene and Scholl, 1991; Kay et al., 1999, 2005; Clift et al., 2003; Ramírez de Arellano et al., 2012). Despite claims of widespread accretion of far-traveled (exotic) terranes genetically linked to Andean deformation (e.g., Nur and Ben-Avraham, 1977, 1982; Moores et al., 2002), post-Paleozoic accretion of continental or oceanic fragments appears limited to the northern Andes (Dalziel and Forsythe, 1985; Mégard, 1987, 1989; Aleman and Ramos, 2000; Ramos, 2009). Accretion of materials largely of oceanic affinity (Irving, 1971; Goossens and Rose, 1973) included possible Jurassic-Cretaceous island arcs (Aspden et al., 1987; Feininger, 1987; Litherland et al., 1994), the Late Cretaceous Caribbean oceanic plateau or large igneous province (Burke, 1988; Kerr et al., 2003; Vallejo et al., 2009; Spikings et al., 2015), and the Late Cretaceous-Cenozoic Panama-Chocó intraoceanic arc (Duque-Caro, 1990; Montes et al., 2012, 2015). In addition, Mesozoic extension involved creation of backarc basins situated between the western edge of South America and fringing or marginal magmatic arcs, either with or without the genesis of an offshore oceanic spreading center. Following backarc extension and possible creation of a narrow strip of oceanic crust, a switch to regional shortening facilitated basin closure and accretion of arc and forearc materials to the continental margin (Dalziel et al., 1974; Bruhn and Dalziel, 1977; Dalziel, 1986; Mpodozis and Ramos, 1990, 2008; Coney and Evenchick, 1994; Benavides-Cáceres, 1999; Villagómez et al., 2011; Calderón et al., 2016).

2.3. Sedimentary record

A source-to-sink consideration of western South America distinguishes major erosional regions from zones of accumulation embodied by continental sedimentary basins. Although onshore basins are emphasized here, a potentially rich but less accessible sedimentary record exists offshore in Pacific sinks of the Peru-Chile trench and Atlantic sinks such as the Argentine and Malvinas basins (e.g., Galeazzi, 1998; Ghiglione et al., 2010, 2014, 2016b; Gruetzner et al., 2012; Eagles, 2016). The dominant sediment sources consist of the Andean magmatic arc, retroarc fold-thrust belt, and eastern cratonic regions such as the Guyana shield, Brazilian shield, and Rio de la Plata craton (Fig. 1C). The onshore sedimentary sinks are situated in forearc and retroarc structural settings, with retroarc foreland basins and hinterland basins (Figs. 1C and 2) containing the most robust stratigraphic archives of Mesozoic-Cenozoic sediment accommodation. Retroarc sedimentary basins generated during Andean shortening were governed by lithospheric flexure with the additional influence of dynamic subsidence and ponding in topographic lows of closed drainage systems.

The Andean foreland basin system includes a thick, eastward-tapering accumulation of clastic sedimentary fill, as defined by Cenozoic-Cretaceous isopach contours (Fig. 1C) and regional structural cross sections (Fig. 2). This low-elevation (< 500 m) basin system commonly exhibits a recognizable foredeep depocenter between the Andean thrust front and distal forebulge. The Andean forebulge is a broad basement high that is largely buried by younger basin fill with little or no topographic expression, but is delimited by gravity data, geoid anomalies, and sediment thickness values (Fig. 1C) (Watts et al., 1995; Horton and DeCelles, 1997; Ussami et al., 1999; Chase et al., 2009; DeCelles, 2012; Nivière et al., 2013; Folguera et al., 2015). An additional proximal zone of foreland accommodation is preserved in wedge-top and piggyback basins atop the frontal sectors of the fold-thrust belt, with higher-elevation hinterland basins positioned farther west between the magmatic arc and the frontal fold-thrust belt (Horton, 1998, 2005, 2012; Sobel et al., 2003; Gillis et al., 2006; Mosolf et al., 2011; Murray et al., 2010; Carrapa et al., 2012).

Similar to most contractional orogens, cratonward advance of the Andean thrust front has led to the structural incorporation of older components of the foreland basin system into the propagating fold-thrust belt. The retroarc foreland basin system spans the length of the Andes and incorporates many individually named basin segments (Fig. 1, basins 1–8) of Venezuela (Barinas-Apure), Colombia (Llanos, Putumayo), Ecuador (Oriente), Peru (Marañon, Santiago, Huallaga, Ucayali, Ene-Camisea), Brazil (Solimões, Acre), Bolivia (Madre de Dios, Beni, Chaco), and Argentina (Cuyo, Neuquén, San Jorge, Rio Mayo, Magallanes-Austral, and offshore Argentine and Malvinas basins). Although more restricted in their areal extent, current retroarc hinterland basins of the northern and central Andes include structurally compartmentalized basins (Fig. 1, basins 9–13) of the Magdalena Valley, Interandean Valley, Cordillera Blanca, and Altiplano-Puna plateau.

Andean basins encompass diverse structural geometries and tectonic styles, as depicted in representative regional cross sections (Fig. 2). Structural and geophysical evidence of significant shortening (> 150–300 km) and crustal thickening (> 50–70 km deep Moho) in the central Andes (Fig. 2A) contrasts with limited shortening (< 50–150 km) and moderate crustal thickening (< 50 km deep Moho) in the northern and southern Andes (Fig. 2B and C) (Kley, 1996; Ramos et al., 1996; Baby et al., 1997; Kley and Monaldi, 1998; Kley et al., 1999; McQuarrie, 2002a, 2002b; McQuarrie et al., 2008b, 2005; Mora et al., 2006, 2013; Oncken et al., 2006, 2013; Espurt et al., 2008; Klepeis et al., 2010; Fosdick et al., 2011; Giambiagi et al., 2012, 2016b; Orts et al., 2012, 2015; Eichelberger et al., 2013, 2015; Ghiglione et al., 2014; Betka et al., 2015; Eude et al., 2015; Perez et al., 2016b). The basins also span regions of contrasting slab configuration, with a moderately dipping subduction zone for the central and southern profiles (Fig. 2A and C) versus a flat-slab geometry for the northern profile (Fig. 2B).

3. Basin genesis during plate convergence

Although Andean stratigraphic records attest to sediment accumulation in forearc, hinterland, and retroarc foreland settings (Fig. 2), the specific structural settings, basin geometries, and subsidence mechanisms reflect a wide range of tectonic processes along ocean-continent convergent plate margins (Fig. 3). Several structural configurations and modes of sediment accommodation operate in forearc and retroarc basins, including compressional, tensional, strike-slip, and neutral stress regimes. Accurate reconstruction of Andean orogenesis largely hinges on accurate delineation of the temporal and spatial histories of the sedimentary basins and their associated tectonic regimes.

3.1. Plate interface and forearc processes

Before considering the better known retroarc regions, it is instructive to outline the activities along the plate interface and forearc region. Several processes are possible near the plate interface during subduction of oceanic lithosphere (Fig. 3A–C). First, along continental margins with a backarc basin floored by either attenuated continental crust or newly created oceanic crust, initial shortening may result in thrust emplacement of a fringing or off-edge magmatic arc (Fig. 3A). This is best expressed in the southernmost Andes, where initial Andean shortening induced thrust closure of the Early Cretaceous Rocas Verdes basin, with eastward emplacement of an ophiolite sequence and consolidation of early forearc, arc, and backarc assemblages (Dalziel, 1986; Wilson, 1991; Fosdick et al., 2011; Ghiglione et al., 2014, 2016a; Calderón et al., 2016). A similar case has been made for the northern Andes, with Early Cretaceous slab rollback and basin development followed by closure and thrust emplacement of transitional to oceanic crust (Peltetec unit; Spikings et al., 2015). Second, accretion of considerable oceanic material beyond minor continuous offscraping of material in the accretionary prism may occur through obduction and/or underplating to the forearc crust of buoyant seamounts, aseismic ridges, or oceanic plateaus (Fig. 3B). This mode of accretion has been invoked for pre-Late Cretaceous collision of assorted oceanic materials in the northern Andes of Ecuador and Colombia (Mégard, 1987; Aspden and Litherland, 1992; Litherland et al., 1994). Third, the collision of a relatively far-traveled oceanic plateau and/or island arc above an oceanward dipping subduction zone could induce basin closure along a passive continental margin (Fig. 3C). Although more speculative, emerging evidence from the northern Andes suggests such a scenario during Late Cretaceous accretion of the Caribbean oceanic plateau or large igneous province (Caribbean Large Igneous Province), prior to the establishment of a continent-dipping subduction zone (Kerr et al., 2002; Vallejo et al., 2017; Spikings et al., 2015).

During plate convergence, forearc basins are variably subjected to shortening, stasis, extension, or strike-slip modes of deformation (Fig. 3D–F). Forearc compression generates a basin related to thrust or reverse faults that induce flexural subsidence (Fig. 3D), as observed in the late Cenozoic northern and central Chilean forearc (Muñoz and Charrier, 1996; Charrier et al., 2007; Armijo et al., 2010; Farías et al., 2005, 2010). Neutral stress conditions in the forearc (Fig. 3E) promote ponding of sediment in a topographic low between the accretionary prism and magmatic arc, as expressed in the early Cenozoic record of central and northern Chile (Mpodozis and Cornejo, 2012; Horton et al., 2016). The forearc may also be governed by fault-induced subsidence during extension orthogonal to the plate margin (Fig. 3F), possibly driven by slab rollback. This situation has been proposed for the mid-

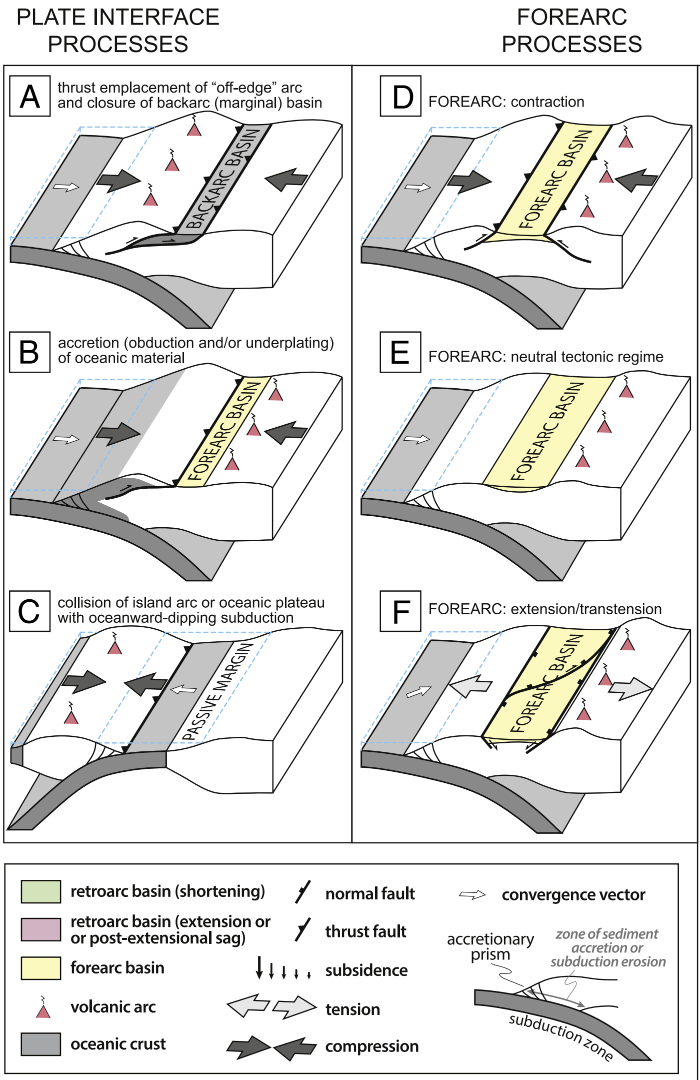

Fig. 3. Schematic block diagrams showing the range of tectonic processes and basin structural settings along an ocean-continent convergent margin, with emphasis on (A–C) plate interface, (D–F) forearc, and (G–J) retroarc processes; (A) thrust emplacement of fringing magmatic arc with closure of backarc basin; (B) accretion of oceanic material to continental margin by obduction and/or underplating; (C) collision of island arc or oceanic plateau along oceanward-dipping subduction zone; (D) forearc shortening with thrust-bounded forearc basin; (E) neutral-stress forearc basin; (F) forearc extension/transtension with fault-bounded forearc basin; (G) retroarc shortening with flexural foreland basin system consisting of wedge-top, foredeep, forebulge, and backbulge depozones; (H) retroarc shortening with flexure in broken foreland basin and piggyback to hinterland basins across fold-thrust belt; (I) neutral-stress retroarc regime with basin abandonment during thrust inactivity and no flexural accommodation; and (J) retroarc postextensional (postrift) thermal sag basin succeeding extensional (synrift) basin.

Cenozoic Abanico basin of central Chile (Charrier et al., 2002; Mpodozis and Cornejo, 2012) and segments of the Peruvian forearc (Noury et al., 2016, 2017; Alván et al., 2017). Although emphasis is placed on orthogonal plate motions, oblique convergence (Fig. 3F) generally results in strain partitioning and strike-slip deformation focused in the forearc and/or magmatic arc (e.g., Fitch, 1972). Late Cenozoic examples of such strike-slip faulting focused along westernmost arc to forearc regions include the Liquiñe-Ofqui fault in central Chile (Cembrano et al., 1996; Thomson, 2002; Rosenau et al., 2006) and the Guayaquil pull-apart basin of western Ecuador (Trenkamp et al., 2002; Witt et al., 2006; Nocquet et al., 2014).

3.2. Retroarc processes

Sediment accommodation in retroarc regions (Fig. 3G–J) is largely controlled by flexural subsidence during thrust loading, (2) dynamic subsidence related to mantle flow, (3) internal drainage in closed topographic basins, or (4) extensional fault motion and subsequent thermal subsidence (Lawton, 1994, 2008; DeCelles and Giles, 1996; Catuneanu, 2004; Miall et al., 2008; Aschoff and Steel, 2011; DeCelles, 2012; Horton, 2012; Sinclair, 2012).

Retroarc shortening and growth of an orogenic wedge induces regional flexure and genesis of a four-part foreland basin system defined by wedge-top, foredeep, forebulge, and backbulge depozones (Fig. 3G) (DeCelles and Giles, 1996). This framework is well developed in central Andean regions with substantial shortening and crustal thickening (Horton and DeCelles, 1997; Horton et al., 2001; DeCelles and Horton, 2003; Roddaz et al., 2005a, 2010). Variants of this configuration (Fig. 3H) involve (1) isolated hinterland basins between the magmatic arc and the fold-thrust belt (Beer et al., 1990; Horton et al., 2002; Carlotto et al., 2005; Murray et al., 2010; Horton, 2012; Carlotto, 2013), (2) narrow piggyback or intermontane basins within the fold-thrust belt (Noblet et al., 1988; Marocco et al., 1995; Horton, 1998, 2005; Mosolf et al., 2011; Suriano et al., 2015), and (3) a broken foreland basin in which large basement-cored uplifts partition the original foreland basin into separate entities (Jordan, 1995; Vergés et al., 2007; Ramos, 2009). These conditions are best expressed in the late Cenozoic record of west-central Argentina in the Sierras Pampeanas

(Pampean) zone of flat slab subduction (Fig. 1) (e.g., Bermejo and associated basins; Jordan et al., 1983, 2001b; Vergés et al., 2001; Ramos et al., 2002; Levina et al., 2014; Capaldi et al., 2017).

In the midst of contractional orogenesis, a potential hiatus in shortening may result in an interruption of foreland flexural subsidence, possible isostatic rebound in proximal basin sectors, and regional transport across an inactive basin (Fig. 3I) (e.g., Heller et al., 1988; Legarreta and Uliana, 1991). This abandonment scenario is exemplified in the Neuquén basin of Argentina (Horton and Fuentes, 2016; Horton et al., 2016) and Middle Magdalena Valley basin of Colombia (Gómez et al., 2003, 2005), with cessation of shortening and development of a regional unconformity (~5–20 Myr hiatus) across the basin. A recent example may be defined for the southernmost Andes, where limited shortening and sediment bypass since ~15 Ma across parts of the Magallanes-Austral basin has resulted in rapid accumulation restricted to the distal offshore Argentine and Malvinas basins (Ghiglione et al., 2016b). Finally, in extensional to neutral retroarc systems, initial (synrift) normal faulting may create isolated subbasins succeeded by later (postrift) sag basins during postextensional thermal subsidence (Fig. 3J). Many such precursor (pre-Andean) basins are entailed in the Late Triassic–Early Cretaceous history of synextensional accumulation east of the magmatic arc prior to regional Andean shortening (e.g., Llanos, Neuquén, and Magallanes-Austral basins; Cooper et al., 1995; Ramos and Aleman, 2000; Horton et al., 2010, 2016).

4. Retroarc stratigraphic successions

Retroarc stratigraphic records along the length of the Andean orogenic belt (Fig. 1) reveal fundamental components of Cretaceous-Cenozoic basin evolution. A review of well-exposed successions from the central and southern Andes (Fig. 4) provides two representative field case studies (Section 4.1). These examples lay the groundwork for a generalized regional compilation (Section 4.2) of the stratigraphic nomenclature, unit thicknesses, dominant lithologies, depositional conditions, and available chronologic data, which highlight temporal and spatial patterns across multiple retroarc foreland (Fig. 5A) and hinterland (Fig. 5B) basins.

4.1. Representative basins

Mesozoic-Cenozoic stratigraphic columns from two localities in the southern Andes (Neuquén basin, Argentina) and central Andes (Eastern Cordillera, Bolivia) illustrate several salient features of Andean retroarc basins. In the southern Andes, deposits of the northern Neuquén basin in the Malargüe fold-thrust belt of west-central Argentina (Fig. 1, basin 7) span the Late Triassic-Jurassic through Neogene (Fig. 4A), making it one of the longest duration records in the Andes. From a basal angular unconformity on Permo-Triassic intrusive and extrusive igneous rocks (Choiyoi Group), the ~7 km thick succession (Fig. 4B) consists of principally marine Mesozoic facies (shale, sandstone, and subordinate evaporite and carbonate of the Precuyo, Cuyo, Lotena, Mendoza, and Rayoso Groups) capped by an upward-coarsening Upper Cretaceous–Cenozoic section of nonmarine shale, sandstone, and conglomerate (Neuquén and Malargüe Groups, Pircala, Coihueco, Agua de la Piedra, Loma Fiera, and Tristeza Formations). A post-Early Cretaceous unconformity at the base of the Neuquén Group marks a shift to coarse-grained nonmarine sedimentation. A major Paleogene hiatus (~40–20 Ma) is represented by a highly condensed section (< 2–20 m thick) or disconformity demarcated by a distinctive regional conglomerate (termed the "Rodados Lustrosos") composed of resistant, well-rounded, polished clasts with highly weathered, oxidized, and desert varnished surfaces (Groeber, 1951; Horton and Fuentes, 2016). In lower stratigraphic levels, a thick Lower Jurassic section is genetically associated with extensional subbasins bounded by normal faults (Giambiagi et al., 2012; Mescua et al., 2014). Upsection, a regionally extensive Middle Jurassic–Early Cretaceous interval is attributed to postextensional thermal subsidence (Manceda and Figueroa, 1995; Horton et al., 2016). Above this stratigraphic discontinuity, in upper stratigraphic levels, a thick Neogene clastic interval is capped by coarse-grained fluvial to alluvial fan facies containing growth strata and local angular unconformities generated by nearby fold-thrust structures (Giambiagi et al., 2008; Boll et al., 2014; Fuentes et al., 2016).

Collectively, the Neuquén stratigraphic succession chronicles a progressive history of initial Mesozoic extension, postextensional thermal subsidence, then a significant transition to Late Cretaceous-Cenozoic flexural subsidence in a foreland basin characterized by a single major Paleogene hiatus (Fig. 4A). A relatively low-magnitude of shortening (< 15–45 km) across the affiliated fold-thrust belt (Kozlowski et al., 1993; Manceda and Figueroa, 1995; Folguera et al., 2006; Turienzo, 2010; Giambiagi et al., 2012; Mescua et al., 2014; Folguera et al., 2015b; Fuentes et al., 2016) is consistent with a comparatively limited eastward migration of the Malargüe foreland basin system. Sedimentation ceased in Pliocene-Quaternary time as the basin was structurally partitioned by the activation of a basement-involved foreland uplift (San Rafael block; Fig. 2C), potentially linked to a shift in the dynamics of the subducting Nazca slab and attendant magmatic effects (Ramos and Folguera, 2009, 2011; Folguera et al., 2009).

In the central Andes of Bolivia, a ~3 km thick Upper Cretaceous-Cenozoic succession exposed in the Eastern Cordillera province of the Andean fold-thrust belt (Fig. 1, basin 5) rests unconformably on lower Paleozoic marine strata (Fig. 4C). The deposits of this succession transition upward from mixed marine and nonmarine clastic and carbonate facies into a thick coarse-grained nonmarine section, which is preserved across a series of Eastern Cordillera structures (Fig. 2A; Horton, 2005) and is particularly well exposed in the Camargo Syncline (Fig. 4D). The lowermost stratigraphic levels include regionally extensive marine Maastrichtian-lower Paleocene deposits spanning the Western Cordillera, Altiplano, and Eastern Cordillera (Welsink et al., 1995; Sempere et al., 1997; Mpodozis et al., 2005). A major Paleogene hiatus is defined by a poorly dated condensed interval (< 20–80 m thick) distinguished by a series of stacked well-developed paleosol horizons within nonmarine sandstone and mudrock host units. Above this discontinuity, a > 2.5 km thick upward coarsening clastic interval is marked by sandstone and conglomerate of distal fluvial to proximal megafan systems in an advancing foreland basin system (Horton and DeCelles, 1997, 2001).

The Eastern Cordillera succession of Bolivia registers a latest Cretaceous-Paleocene history of fine-grained regional sedimentation related to low-magnitude thermal, dynamic, or distal flexural subsidence followed by an important hiatus, then pronounced flexure in a foreland basin. Substantial east-west shortening of > 200–300 km across this part of the central Andes (Kley, 1996; Baby et al., 1997; Kley et al., 1997; Kley and Monaldi, 1998; McQuarrie, 2002a, 2002b; McQuarrie et al., 2005; Eichelberger et al., 2013, 2015; Anderson et al., 2017) requires that the fold-thrust belt and adjacent foreland basin of Bolivia have propagated dramatically eastward during Cenozoic shortening. Such large-scale horizontal advance of the Andean topographic load and resulting cross-sectional flexural subsidence profile toward the craton has several implications. First, this scenario suggests that the earliest evidence of shortening-induced sedimentation may be preserved only in the most-distal (backbulge) zones of the foreland basin. Second, major stratigraphic gaps could be the product of erosion or nondeposition in a forebulge setting. Third, the modern Chaco foreland basin (Figs. 1 and 2A) appears to be the final expression of a long-lived process and therefore contains only the youngest stage of the complete Andean orogenic history (e.g., Horton and DeCelles, 1997; DeCelles and Horton, 2003; Uba et al., 2006; DeCelles et al., 2011; Calle et al., 2018).

Comparison of the two representative retroarc basins of the southern and central Andes highlights several themes. Both basins are similarly characterized by upward coarsening successions displaying (1) a stratigraphic transition from marine to nonmarine depositional conditions, (2) an appreciable hiatus within the Paleogene nonmarine

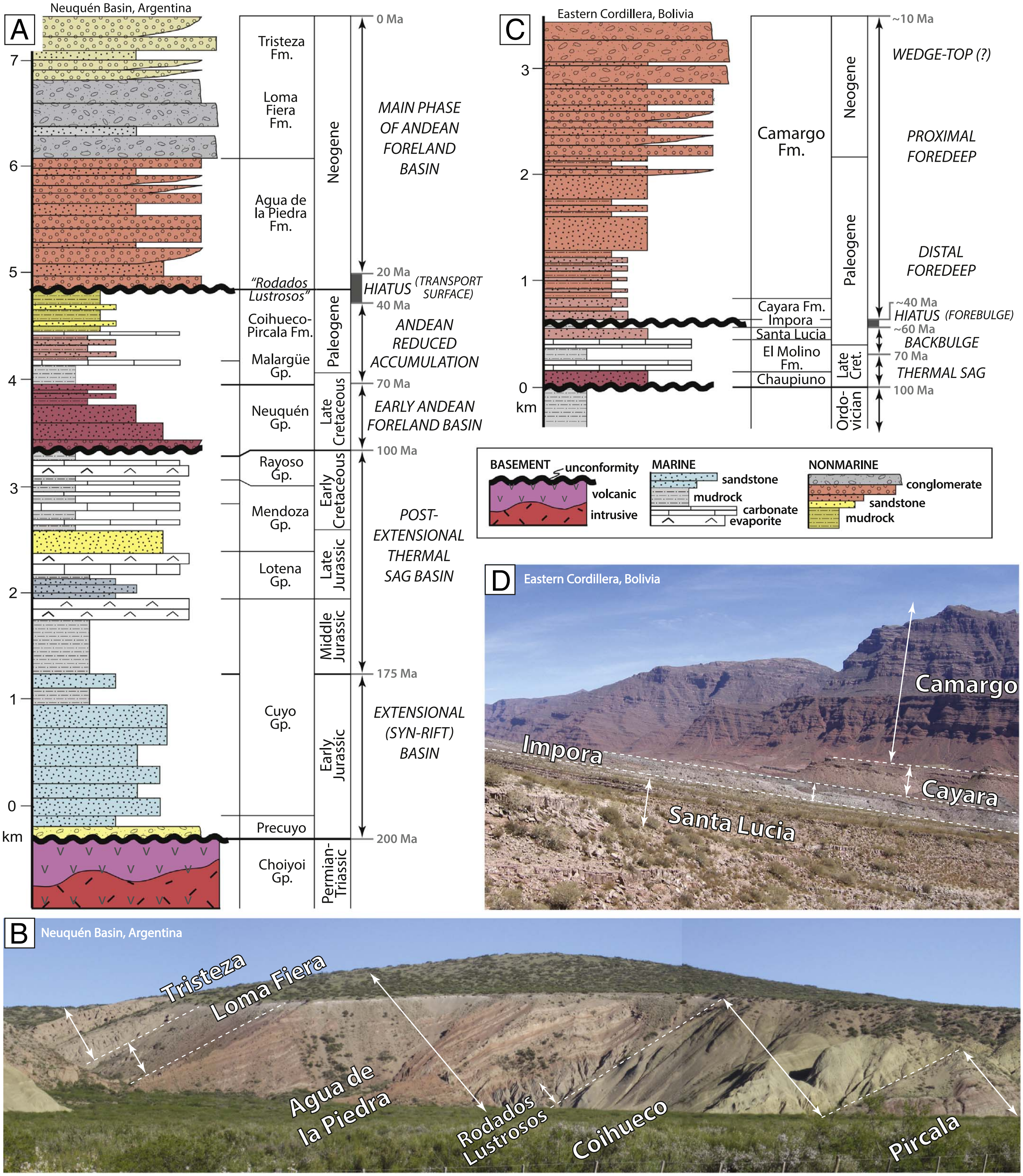

Fig. 4. Composite Mesozoic-Cenozoic stratigraphic sections and photographs of representative Andean retroarc basins showing stratigraphic nomenclature, lithology, thickness, age, and tectonic setting. (A) Stratigraphic section and (B) photograph of the northern Neuquén basin, Argentina (after Horton et al., 2016). (C) Stratigraphic section and (D) photograph of foreland basin fill in the Eastern Cordillera, Bolivia (after Horton and DeCelles, 2001; DeCelles and Horton, 2003). For both localities, note the basal unconformity and Paleogene stratigraphic hiatus in lower levels of foreland basin fill.

interval, and (3) a principally Neogene upper interval of proximal coarse-grained fluvial to alluvial fan facies closely associated with fold-thrust structures (growth strata). One prime difference is the presence or absence of a substantial precursor stratigraphic interval produced by earlier growth of extensional subbasins (synrift phase) and a post-extensional thermal sag (postrift phase). The details of such precursor records (generally referred to as pre-Andean), in terms of their presence, thickness, and spatial distribution, are critical in evaluating the

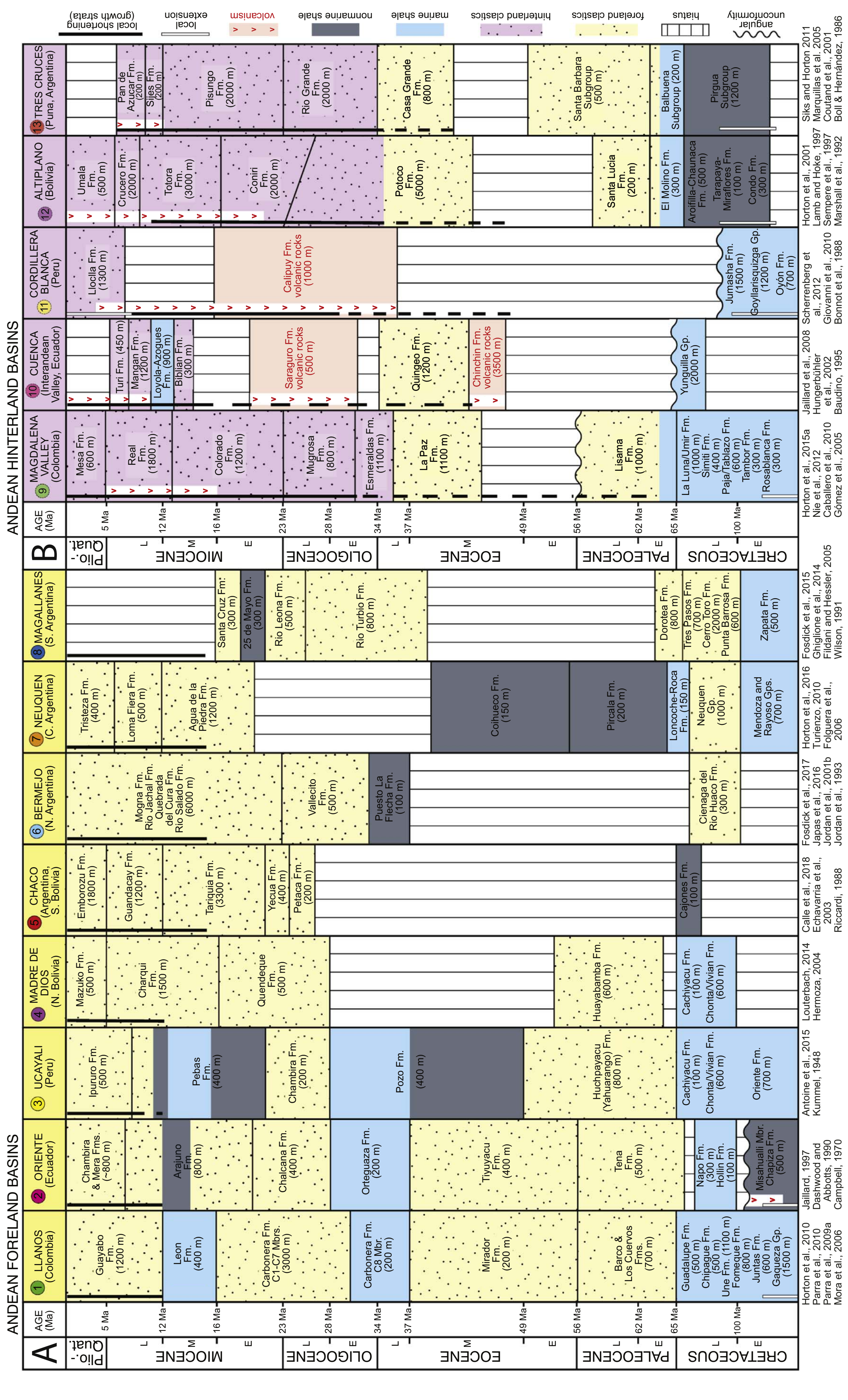

Fig. 5. Comparative chronostratigraphic diagram showing stratigraphic nomenclature, age, thickness, lithology, depositional setting, stratigraphic discontinuities (unconformities and hiatuses) and associated local deformation for (A) 8 Andean foreland basins and (B) 5 Andean hinterland basins. Map locations of 13 basin localities shown in Fig. 1.

role of inherited stratigraphic geometries in guiding later shortening (e.g., Stockmal et al., 2007; Sinclair, 2012; Chapman and DeCelles, 2015). A second major difference concerns the magnitude of horizontal shortening and lateral migration of the retroarc foreland basin. High-shortening segments of the orogenic belt are invariably susceptible to large-scale cratonward advance of deformation, potentially far beyond the original flexural wavelength of the initial flexural basin (e.g., DeCelles and DeCelles, 2001; Christophoul et al., 2003; Ford, 2004; Hilley and Strecker, 2004; Uba et al., 2009). In contrast, low-shortening regions with limited propagation of the topographic load result in relatively stationary forebulges (e.g., Fosdick et al., 2014; Folguera et al., 2015a, 2015b). Both of these topics—the role of precursor pre-Andean basins and the magnitude of Andean shortening—are considered essential to accurate interpretation of the retroarc stratigraphic record.

4.2. Regional stratigraphic framework

Compilation of the Mesozoic-Cenozoic stratigraphic archives from numerous sources enables a comparison of 13 representative retroarc basins along the length of the Andes (Fig. 5). Eight foreland segments include, from north to south, the Llanos, Oriente, Ucayali, Madre de Dios, Chaco, Bermejo, Neuquén, and Magallanes-Austral basins (Fig. 5A, basins 1–8) (Kummel, 1948; Tschopp, 1953; Campbell, 1970; Biddle et al., 1986; Riccardi, 1988; Dashwood and Abbotts, 1990; Jordan et al., 1993, 2001b; Jaillard, 1997; Echavarria et al., 2003; Hermoza, 2004; Hermoza et al., 2005; Folguera et al., 2006; Mora et al., 2006, 2010a; Parra et al., 2009a, 2010; Horton et al., 2010, 2016; Ghiglione et al., 2014; Louterbach, 2014; Antoine et al., 2016; Japas et al., 2016; Fosdick et al., 2015, 2017; Calle et al., 2018). Five hinterland zones in the northern and central Andes include age-equivalent strata of the Middle Magdalena Valley basin, Cuenca basin (within the Interandean Valley of Ecuador), Callejon de Huaylas basin (adjacent to the Cordillera Blanca), northern Altiplano, and Puna plateau (Fig. 5B, basins 9–13) (Morales, 1958; Boll and Hernández, 1986; Bonnot et al., 1988; Marshall et al., 1992; Baudino, 1995; Lamb and Hoke, 1997; Sempere et al., 1997; Coutand et al., 2001; Horton et al., 2001, 2015a; Hungerbühler et al., 2002; Gómez et al., 2005; Marquillas et al., 2005; Jaillard et al., 2008; Caballero et al., 2010, 2013; Giovanni et al., 2010; Nie et al., 2010, 2012; Moreno et al., 2011; Siks and Horton, 2011; Horton, 2012; Scherrenberg et al., 2012).

The geologic column presented for each basin summarizes the stratigraphic ages, nomenclature, thicknesses, stratigraphic discontinuities (unconformities and hiatuses), generalized lithologies, associated arc volcanism, and marine versus nonmarine depositional conditions of major stratigraphic units, facilitating the recognition of temporal and spatial patterns across 8 retroarc foreland basins (Fig. 5A) and 5 retroarc hinterland basins (Fig. 5B). Additional information includes the estimated timespans of extension, as recorded chiefly by basin-bounding normal faults (e.g., Marquillas et al., 2005; Mora et al., 2006, 2009; Giovanni et al., 2010; Scherrenberg et al., 2012; Tesón et al., 2013; Mescua et al., 2014), and the timing of local Andean shortening provided by growth strata affiliated with fold-thrust structures in hinterland and wedge-top depozones (e.g., Jordan et al., 1993; Zapata and Allmendinger, 1996; Gómez et al., 2005; Carrera and Muñoz, 2008; Siks and Horton, 2011; Orts et al., 2012; Parra et al., 2012; Perez and Horton, 2014; Horton et al., 2015b, 2016; Ramos et al., 2015; Echaurren et al., 2016).

This orogen-scale synthesis is bound to have minor errors in stratigraphic ages, owing to insufficient chronostratigraphic resolution or time-transgressive depositional histories. Further, the composite thickness values reported for individual basins are not basin-wide averages; rather, they symbolize thicknesses for particular localities within hinterland basins typically defined by narrow linear depocenters and within foreland basins that show eastward thinning, as depicted in regional isopach contours (Fig. 1C; Yrigoyen, 1991). To facilitate comparison, the geologic columns also include interpretations of basin tectonic setting in terms of foreland versus hinterland position.

The Lower Cretaceous to Quaternary composite stratigraphic thickness ranges from 3.5 to 14 km, with an average of about 7 km for all basins, roughly 6 km and 9 km in foreland and hinterland regions, respectively (Fig. 5). Consistent with regional isopach data (Fig. 1C), the thickest successions are observed in the northernmost Andes (Llanos and Magdalena basins), central Andes (Chaco and Altiplano basins), and southernmost Andes (Magallanes-Austral basin).

In terms of lithologies and depositional systems, the successions consistently show overall upward coarsening trends, with marine facies prevalent in Cretaceous units but restricted to rare, narrow Cenozoic intervals. The major shift from marine to nonmarine facies principally takes place in Maastrichtian to lower Paleocene strata, broadly contemporaneous with a long-term drop in eustatic sea level at ~72–61 Ma (Miller et al., 2005; Haq, 2014). Brief marine incursions in retroarc foreland zones have been shown for the late Eocene-Oligocene and early-middle Miocene in the northern Andes (Fig. 5A; Llanos, Oriente, and Ucayali basins; Santos et al., 2008; Roddaz et al., 2005a, 2010; Jaramillo et al., 2017) and late Oligocene-early Miocene in the southern Andes (Patagoniense seaway) (Malumián, 2002; Cuitiño and Scasso, 2010; Parras et al., 2012; Encinas et al., 2014, 2016), with additional debated proposals for major Miocene incursions from the north (Pebas seaway) and south (Parana seaway) that may have reached the central Andean foreland (e.g., Räsänen et al., 1995; Hoorn et al., 1995, 2010; Hovikoski et al., 2005; Lundberg et al., 1998; Hernández et al., 2005; Latrubesse et al., 2007; Uba et al., 2007).

Perhaps most intriguing is the presence of significant stratigraphic gaps in multiple retroarc foreland settings (Fig. 5A). Hiatuses spanning roughly 10 to 30 Myr are particularly evident within the Paleogene basin record, where they are marked by non-angular disconformities or highly condensed intervals. Even where discontinuities are absent, the preserved Paleogene stratigraphic section is uniformly low in thickness, typically < 500–1000 m. These observations all point to a period of limited accommodation within Andean retroarc regions, a key issue that is elaborated later in this article.

Comparisons of individual foreland basin localities with their nearest hinterland basins provide a valuable measure of the degree of shared sedimentary history for individual cross-strike orogen-perpendicular transects. For the central Andes at 14°–25°S, a considerable discrepancy is observed in which the onset of major foreland accumulation in the Chaco Plain to Madre de Dios basin was delayed until the late Oligocene, some 20–40 Myr after initial flexural accommodation farther west in the corresponding Altiplano and Puna hinterland plateaus (Fig. 2A). Given the extreme shortening in this part of the central Andes, this delay in accommodation may be ascribed to the large-scale eastward advance of the fold-thrust belt and foreland basin, coupled with uplift and erosional removal of older foreland basin deposits. This example underscores the challenge in recognizing the true inception of shortening and flexural accommodation in high-shortening segments of a contractional orogen.

5. Foreland basin unconformities

Accurate detection and interpretation of stratigraphic hiatuses in retroarc foreland basins (Miall, 2016) is fundamental to defining a complete history of basin evolution and Andean orogenesis. Alternative modes of unconformity genesis (Fig. 6) can be variably linked to forebulge dynamics, tectonic quiescence, or local fold-thrust deformation, with implications for subsidence mechanisms and associated deformation at local and regional scales.

Forebulges represent the flexural response to thrust-induced topographic loading, as influenced by the orogenic load geometry and mechanical properties of the flexed foreland lithosphere, notably the effective elastic thickness. They may form positive topographic features susceptible to surface erosion, or broad-wavelength zones of sediment bypass or limited accumulation. The genesis of forebulge stratigraphic

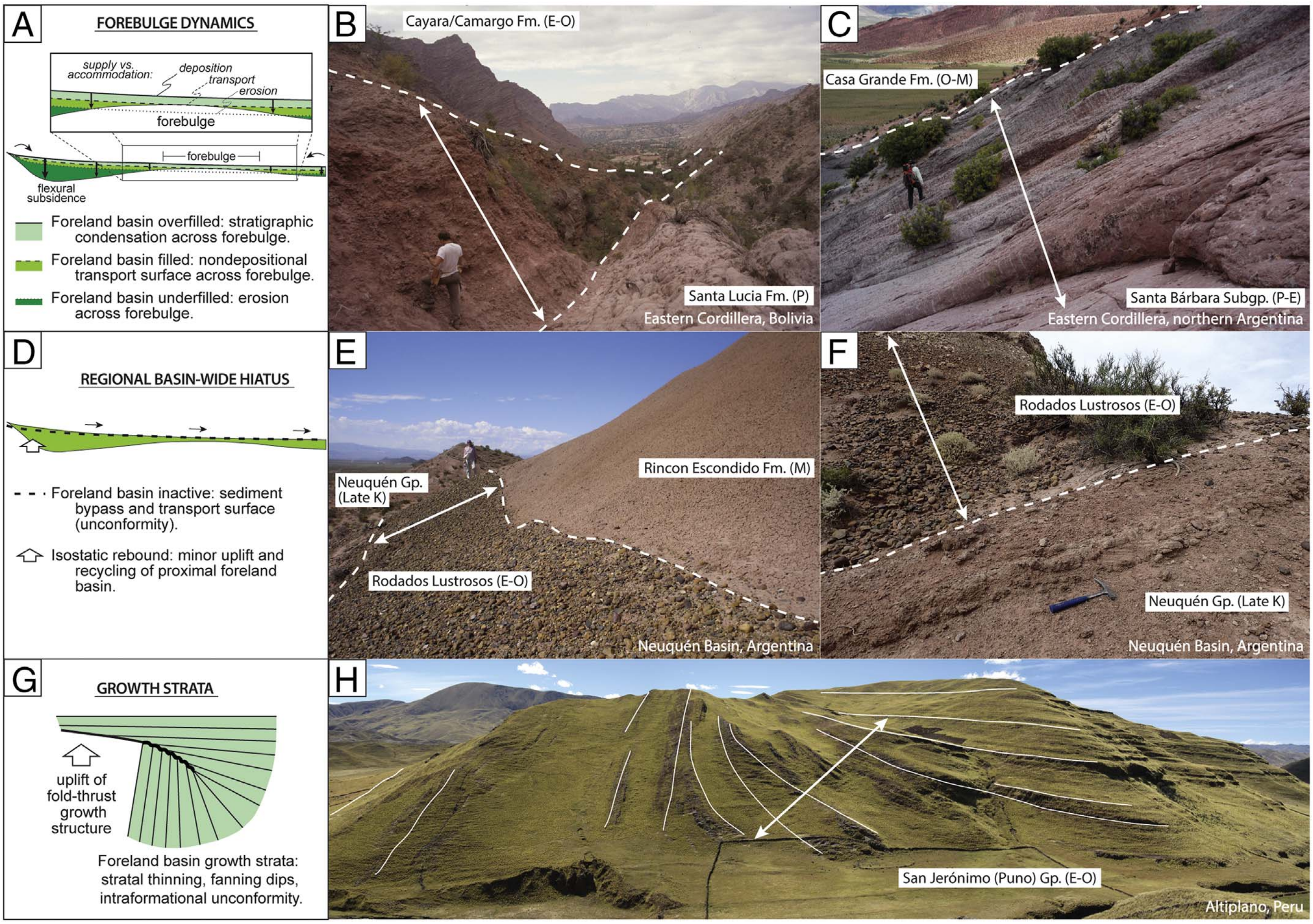

Fig. 6. Diagrams and photographs depicting the genesis of unconformities and condensed stratigraphic sections in Andean retroarc foreland basins. (A) Variations in forebulge dynamics dictate basin configuration (underfilled, filled, overfilled) and forebulge process (erosion, transport, deposition), with examples from (B) Eastern Cordillera, Bolivia (see Horton, 2005) and (C) Eastern Cordillera, northern Argentina (see DeCelles et al., 2011; Siks and Horton, 2011). (D) Generation of a regional hiatus across foreland basin, with examples from (E) Neuquén basin (from Garrido et al., 2012) and (F) northern Neuquén basin, Argentina (see Horton et al., 2016). (G) Shortening-related growth strata (after Riba, 1976; Anadón et al., 1986) and example from (H) Altiplano basin, southern Peru (reflected image; see original in Perez and Horton, 2014; Horton et al., 2015b). Age abbreviations: Cretaceous (K); Paleocene (P); Eocene (E); Oligocene (O); and Miocene (M).

discontinuities is regulated by the ratio between sediment supply and accommodation generation (Fig. 6A). This value dictates foreland basin geometry in terms of underfilled (supply > accommodation), filled (supply = accommodation), or overfilled (supply < accommodation) basin configurations (e.g., Flemings and Jordan, 1989; Crampton and Allen, 1995; Currie, 1997; DeCelles, 2012).

For underfilled basins, erosion of an exposed forebulge produces an unconformity. For filled basins, sediment bypass results in a disconformity or highly condensed stratigraphic interval over the forebulge crest. For overfilled basins, low accommodation over the buried forebulge yields a condensed interval. In high-shortening segments of the central Andes (Madre de Dios, Chaco, Altiplano, Puna basins), long-lived stratigraphic discontinuities up to 5–20 Myr are marked by stacked paleosol horizons (or supersols) up to 20–100 m thick and have been attributed to forebulge genesis and migration (e.g., Horton et al., 2001; DeCelles and Horton, 2003; Roddaz et al., 2005b; DeCelles et al., 2011; Louterbach, 2014). These paleosols are distinguished by the preservation of pedogenic features (glaebules, nodules, root traces, soil horizonation, diagnostic trace fossils), the destruction of primary sedimentary structures, and the stratigraphic position between thin, fine-grained deposits of a distal foreland basin (backbulge) below and thick coarse-grained basin fill of proximal foredeep deposits above (Fig. 6B and C).

In addition, major stratigraphic hiatuses in foreland basins may also denote periods of large-scale basin abandonment during thrust inactivity and the absence of flexural accommodation. Although there remains no strong evidence to support proposals of episodic thrusting on short ($< 10^5 - 10^6$ yr) timescales, cases of protracted inactivity of longer duration ($> 10^6 - 10^7$ yr) may signify cessation of thrust loading and flexural subsidence (Fig. 6D). Rather than sediment accumulation, tectonic quiescence should promote regional sediment bypass with development of an abandonment or transport surface, locally marked by thin accumulations of relict gravel lags composed of resistant clasts (Fig. 6E and F). This mechanism has been interpreted for ~5–20 Myr hiatuses within foreland basin successions of the northern Andes (Magdalena Valley and Putumayo basins; Gómez et al., 2003; Londono et al., 2012) and southern Andes (Neuquén basin; Horton et al., 2016; Horton and Fuentes, 2016).

In the structurally dismembered sectors of foreland basins, fold-thrust deformation generates growth strata and local unconformities in wedge-top and piggyback basins. These widely recognized synorogenic features of contractional basins are defined by an upsection reduction in stratal dip, thinning of beds toward the local structure, and common intraformational angular discordances (Fig. 6G and H) (Riba, 1976; Anadón et al., 1986). Growth strata are restricted to proximal basin fill within the wavelength of individual structures (typically < 5–10 km horizontal distances) and unequivocally demonstrate syndepositional shortening. These features are preferable to less-direct measures of deformation timing because they can help place exact age constraints on shortening, as utilized in the aforementioned local and regional syntheses (Figs. 4 and 5).

6. Retroarc sediment accumulation

6.1. Sediment accumulation histories

A synthesis of Mesozoic-Cenozoic sediment accumulation histories for Andean retroarc regions (Fig. 7) facilitates an orogen-scale comparison of the onset of major sedimentation, phases of basin subsidence, temporal shifts in accumulation rates, and potential spatial variability. Accumulation histories within foreland basins have proven instrumental in the assessment of the onset and pace of contractional orogenesis and tectonic subsidence (e.g., Heller et al., 1986; Allen et al., 2000; Xie and Heller, 2009). In the Andes, insufficient age control from dominantly nonmarine, fossil-poor intervals has limited broader comparison and correlation across parts of the Cenozoic stratigraphic record. Magnetic polarity stratigraphy has provided a chronostratigraphic framework on < 5–10 Myr timescales (e.g., Reynolds et al., 1990, 2001; Jordan, 1995; Echavarria et al., 2003), but reconstructions of longer duration must rely on a combination of ever-improving depositional age constraints, notably from U-Pb geochronology of tuffs and interbedded detrital materials (e.g., Fildani et al., 2003; Blisniuk et al., 2005; DeCelles et al., 2007; Barbeau et al., 2009; Fosdick et al., 2011; Horton et al., 2015b; Ghiglione et al., 2015; Cuitiño et al., 2016).

Here, a series of time versus thickness diagrams (Fig. 7) display the sediment accumulation curves for 43 localities across all 13 basins (8 retroarc and 5 hinterland basins; Fig. 1) on the basis of stratigraphic ages and thickness information from multiple published studies (Natland et al., 1974; Biddle et al., 1986; Baby et al., 1995; Manceda and Figueroa, 1995; Steinmann, 1997; Thomas et al., 1995; Allmendinger et al., 1997; Marksteiner and Aleman, 1997; Cardozo and Jordan, 2001; Echavarria et al., 2003; Navarro et al., 2005; Gómez

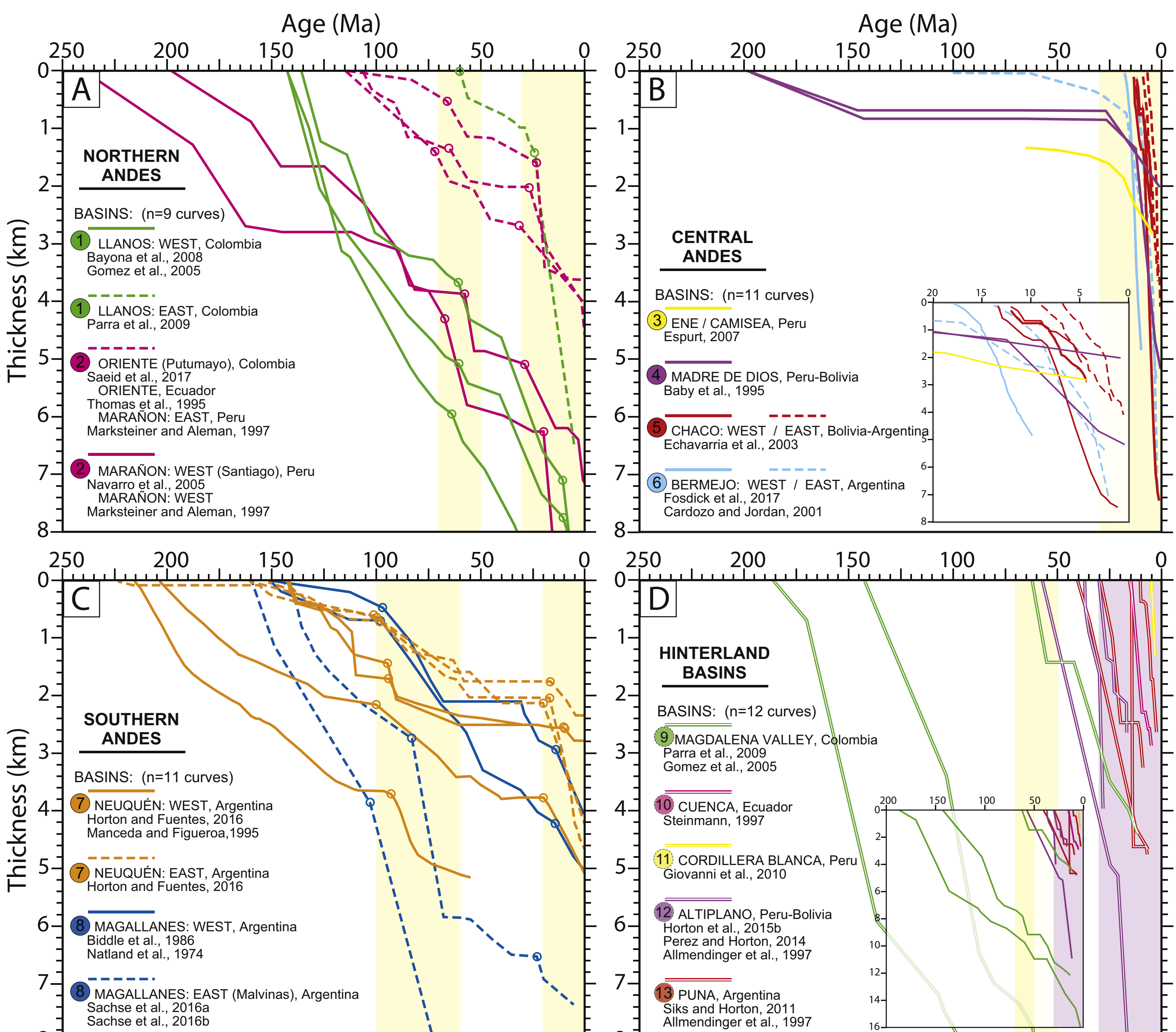

Fig. 7. Time versus thickness diagrams showing undecompacted sediment accumulation histories for Andean retroarc basins, including (A) northern Andes (Llanos, Oriente, and Marañon basins), (B) central Andes (Ene-Camisea, Madre de Dios, Chaco, and Bermejo basins), (C) southern Andes (Neuquén and Magallanes-Austral basins), and (D) Andean hinterland basins (Magdalena Valley, Cuenca, Cordillera Blanca, Altiplano, and Puna basins). Each curve displays stratigraphic thickness for an individual basin locality (solid curves for western localities; dashed curves for eastern localities), with inflection points (open circles) showing a shift to rapid accumulation. Color shading denotes phases of rapid accumulation in foreland (yellow) and hinterland (purple) settings. Map locations of 13 basin localities shown in Fig. 1.

et al., 2005; Espurt, 2007; Bayona et al., 2008; Parra et al., 2009a; Giovanni et al., 2010; Siks and Horton, 2011; Perez and Horton, 2014; Horton et al., 2015a, 2015b; Horton and Fuentes, 2016; Sachse et al., 2016a, 2016b; Fosdick et al., 2017; Saeid et al., 2017).

As in most contractional mountain belts, significant diversity in structural geometry, shortening magnitude, crustal thickness, and overall lithospheric architecture (Fig. 2) ensures spatial variations in the amount of topographic loading and associated flexural subsidence. For the Andes, spatial variations in the mechanical properties of the South American plate, including pre-conditioning through earlier attenuation and inheritance of regional heterogeneities (faults, sutures, basement fabrics), are further important in guiding subsidence patterns (e.g., Watts et al., 1995; Ussami et al., 1999; Mora et al., 2009; Carrera and Muñoz, 2013; Fosdick et al., 2014). Pre-Andean extension in particular has left an imprint on structural and thermal regimes by creating Mesozoic normal faults suitable for later contractional reactivation (e.g., Grier et al., 1991; Kley et al., 1997; Kley and Monaldi, 2002; Deeken et al., 2006; Insel et al., 2012; Sánchez et al., 2012) and inducing thermal weakening of large regions, possibly making them more susceptible to later shortening (Isacks, 1988; Wdowinski and Bock, 1994; Pope and Willett, 1998). Nevertheless, despite these potential complications, several discernible trends over the past 250 Myr are apparent in different Andean segments along the length of the orogen (Fig. 7).

6.2. Mesozoic sedimentation

A very long record of basin accumulation is defined in the northern and southern Andes, where Mesozoic sedimentation commenced at rapid rates and was followed by progressively diminished rates. This pattern is expressed as concave-upward sediment accumulation curves in time versus thickness plots (Fig. 7A and C) and is attributable to Mesozoic backarc extension (initial fault-controlled synrift subsidence) and subsequent postextensional sag basin conditions (postrift thermal subsidence). The precise onset of accumulation ranges between roughly 250 and 140 Ma, consistent with the following timing disparities demonstrated by previous studies: (1) Triassic extension in the Marañon and Madre de Dios basins (Jaillard et al., 1990; Baby et al., 1997; Mathalone and Montoya, 1995; Sempere et al., 2002; Macellari and Hermoza, 2009; Perez et al., 2016a); (2) Late Triassic-Early Jurassic extension in the Neuquén basin (Manceda and Figueroa, 1995; Vergani et al., 1995; Franzese and Spalletti, 2001; Howell et al., 2005); (3) Late Jurassic-earliest Cretaceous extension in the Llanos, Magdalena, Oriente, and Magallanes-Austral basins (Biddle et al., 1986; Cooper et al., 1995; Baby et al., 2004; Mora et al., 2006, 2009; Sarmiento-Rojas et al., 2006; Sachse et al., 2016a, 2016b; Malkowski et al., 2016); and (4) Early to mid-Cretaceous extension in the Salta rift of northernmost Argentina, near the Puna and Bermejo basins (Grier et al., 1991; Salfity and Marquillas, 1994; Marquillas et al., 2005; Deeken et al., 2006; Insel et al., 2012). These diverse phases of extension were uniformly succeeded by a long period of principally neutral stress conditions and slow thermal subsidence dictated by lithospheric cooling (Ramos and Aleman, 2000; Jaillard et al., 2000).

After diminished Mesozoic subsidence, an abrupt Late Cretaceous-early Paleocene increase in accumulation rates can be linked to the inception of large-scale Andean shortening. For individual sediment accumulation curves, a convex-upward profile with an inflection point in time versus thickness plots defines the onset of amplified accumulation (Fig. 7). The shift to rapid accumulation commenced at 70–60 Ma in the northern Andes (Fig. 7A), in agreement with interpretations of Maastrichtian-early Paleocene initial shortening in the Central Cordillera of Colombia (Cooper et al., 1995; Gómez et al., 2003, 2005; Horton et al., 2010; Mora et al., 2010b; Villagómez and Spikings, 2013; Reyes-Harker et al., 2015) and its along-strike continuation to the south, the Eastern Cordillera of Ecuador (Barragán, 1999; Aleman and Ramos, 2000; Ruiz, 2002; Baby et al., 2004; Vallejo, 2007). In the southern Andes, accelerated accumulation commenced at ~100 Ma and persisted until roughly 60 Ma (Fig. 7C). These ages match independent estimates for shortening in the Magallanes fold-thrust belt derived from cross-cutting structural relationships, metamorphic conditions, and thermochronological cooling histories (Dalziel et al., 1974; Dalziel, 1981, 1986; Bruhn and Dalziel, 1977; Nelson et al., 1980; Kohn et al., 1995). The ~100 Ma onset of shortening also coincides with a well-documented shift in deposystems within the Magallanes-Austral basin that has been long ascribed to earliest Andean shortening (Natland et al., 1974; Winn and Dott, 1979; Dott et al., 1982; Biddle et al., 1986; Wilson, 1991).

6.3. Cenozoic sedimentation

Early Andean flexural subsidence was superseded in Paleogene time by a poorly understood phase of limited accumulation that may be the consequence of a lull in Andean shortening. Although this phase is not uniform in magnitude or duration, available Paleogene foreland accumulation histories consistently indicate a reduction in accumulation between roughly 60 and 20 Ma, as defined by the reduced slope of undecompacted sediment accumulation curves in time versus thickness plots (Fig. 7). In the southern Andes, reduced Paleogene accumulation at roughly 60–20 Ma may represent a neutral tectonic regime across the retroarc region (Horton and Fuentes, 2016; Horton et al., 2016), partially contemporaneous with late Eocene–early Miocene extension in hinterland to forearc provinces (Suárez and Emparan, 1995; Godoy et al., 1999; Jordan et al., 2001a; Charrier et al., 2002; Ramos and Folguera, 2005; Burns et al., 2006; Folguera et al., 2010; Rojas Vera et al., 2010; García Morabito and Ramos, 2012). In the northern Andes, reduced accumulation at roughly 50–30 Ma is discernible across the Llanos foreland basin and parts of the Magdalena Valley hinterland basin (Fig. 7A). Although more speculative, the middle Eocene–Oligocene history of the northern Andes may also reflect, in part, diminished shortening or neutral stress conditions, in accord with reports of retroarc quiescence in Colombia (Gómez et al., 2003; Roddaz et al., 2010; Londono et al., 2012; Mora et al., 2013) and a low-elevation, low-exhumation hinterland region in Ecuador (Delfaud et al., 1999; Spikings et al., 2000; Aleman and Ramos, 2000; Christophoul et al., 2002; Vallejo et al., 2016).

In contrast to foreland segments of the northern and southern Andes, a Mesozoic to Paleogene record is absent from several hinterland basins and the central Andean foreland. In the central Andes, this can be attributed to large-magnitude shortening and cratonward advance of the orogenic wedge, such that the present foreland localities (Fig. 5A) were originally beyond the flexural wavelength of the early fold-thrust belt. In hinterland areas, the missing stratigraphic record (Fig. 5B) potentially signifies a conversion from an actively eroding sector of the fold-thrust belt to a zone of net accommodation. This transition is recognized as shift from external to internal drainage, which may be the result of a broadening (widening) orogen and associated orographic effects, such as aridification of a hinterland plateau (Schwab, 1985; Horton et al., 2002; Sobel et al., 2003; Strecker et al., 2007, 2009; Hilley and Coutand, 2010).

In what is commonly considered to be the main phase of Andean mountain building, accumulation histories across all retroarc segments of the northern, central, and southern Andes show rapid sedimentation over the past 20–30 Myr (Fig. 7). The highest rates of accumulation are observed in the central Andean foreland (Fig. 7B) and hinterland basins of the northern and central Andes (Fig. 7D), where 3–8 km of Neogene fill was deposited. This record denotes major flexural subsidence in foreland settings, in agreement with structural evidence for increased shortening rates (e.g., Elger et al., 2005; Oncken et al., 2006, 2013; Mora et al., 2008, 2013). Enhanced accommodation in hinterland settings can be linked genetically to a combination of flexure, local fault-induced subsidence, and topographic infilling of closed drainages within hinterland basins (Jordan and Alonso, 1987; Kennan et al., 1995; Marocco et al., 1995; Hungerbühler et al., 2002; Carlotto et al., 2005; Horton, 2005, 2012; Carlotto, 2013).

7. Detrital zircon provenance and the inception of Andean orogenesis

7.1. Synthesis of U-Pb geochronological results

A compilation of detrital zircon U-Pb geochronological results plots (Fig. 8) reveals the key signatures indicative of the activation and progression of Andean orogenesis. Recent technological advances in ICPMS (inductively coupled plasma mass spectrometry) methodologies have fueled an explosion of U-Pb geochronological applications to sediment provenance problems (Gehrels et al., 2008; Gehrels, 2014). The approach is remarkably well suited for Andean sedimentary basins because the major sediment source regions—the Andean magmatic arc, retroarc fold-thrust belt and South American craton—yield distinctive provenance signatures or fingerprints and can be readily distinguished from one another (e.g., Horton et al., 2010; Nie et al., 2012; Capaldi et al., 2017). For the Andes, recent provenance studies have focused much effort on determining the erosional exhumation history of sediment source regions of the evolving retroarc fold-thrust belt and magmatic arc, with greater attention to the younger, generally Neogene, history of Andean orogenesis (e.g., Sagripanti et al., 2011; Decou et al., 2013; Perez and Horton, 2014; Levina et al., 2014; Anderson et al., 2016; Ramos et al., 2015; Echaurren et al., 2016; Amidon et al., 2017; Streit et al., 2017; Witt et al., 2017). The motivation here is to identify the decisive switch from pre-orogenic to early orogenic provenance signatures, thereby informing our understanding of initial mountain building, regional paleodrainage, and growth of topographic barriers (e.g., Baker et al., 2014; Hoorn et al., 2010; Rodríguez Tribaldos et al., 2017).

Given this objective, detrital zircon U-Pb ages have been compiled chiefly from Cretaceous and Paleogene sandstone samples in order to delineate substantial shifts in regional sediment routing. It is asserted here that an orogen-scale reversal in “sedimentary polarity” (e.g., Dewey and Bird, 1970; Coney and Evenchick, 1994), from continental interior (cratonic) source regions to Andean source regions, occurred in all sections of the Andean retroarc basin system. Within individual basin segments, this solitary reversal event symbolizes a wholesale shift from westward drainage to eastward drainage, consistent with establishment of major orogenic source regions and a sizeable Andean topographic barrier. A summary of ~13,000 U-Pb ages from ~150 detrital zircon samples is presented for 7 regional basin transects to enable along-strike and across strike comparisons, allowing for elucidation of north-south variations in sedimentary polarity and tracking of the time-transgressive west to east progression of basin evolution during Andean deformation. Meaningful comparison of such a large dataset requires samples of comparable stratigraphic age from similar locations to be pooled into composite age probability functions, affording rapid visual comparison of U-Pb age distributions from multiple samples. In most cases, a single composite curve represents results from about 2–6 samples, depending on data availability. A total of 34 age probability curves are depicted in 17 plots (Fig. 8), with each plot comparing two composite samples: one older sample showing a pre-Andean provenance signature and one younger sample showing an early Andean provenance signature. For each sample pair, the respective depositional/stratigraphic ages for the pre-Andean and Andean samples provide a maximum and minimum estimate for the timing of sedimentary routing reversal for each locality.

7.2. Northern Andean provenance

For the northern Andes (Fig. 1), U-Pb results for detrital zircon sample suites are compiled for three regional retroarc basin transects, including (1) the Magdalena Valley, Eastern Cordillera, and Llanos basins of Colombia (Fig. 8A), (2) the Western Cordillera, Cuenca, and Oriente basins of Ecuador (Fig. 8B), and (3) the Western Cordillera, Eastern Cordillera, and Marañon (Santiago) basins of northern Peru (Fig. 8C) (Vallejo, 2007; Horton et al., 2010, 2016; Nie et al., 2010, 2012; Saylor et al., 2011; Caballero et al., 2013; Silva et al., 2013; George et al., 2015; Jackson et al., 2015; George and Horton, 2017; Horton, 2017; Gutierrez et al., 2017).

All northern Andean transects consistently show a sharp contrast between pre-Andean and early Andean samples. Pre-Andean age populations for principally Upper Cretaceous stratigraphic units are dominated by mostly pre-450 Ma ages, including major age groups at 500–700, 900–1100, 1350–1450, 1500–1600, and 1750–1850 Ma. This age distribution demonstrates erosional derivation from Precambrian-Ordovician cratonic basement rocks of the northern Andean foreland and Guyana (Guiana) shield (e.g., Cordani et al., 2000, 2005; Cordani and Teixeira, 2007; Chew et al., 2007; Cardona et al., 2010a; Ibañez-Mejia et al., 2011; Kroonenberg and De Roever, 2010). In contrast, early Andean samples of younger, principally Paleogene strata show substantially reduced proportions of pre-450 Ma basement ages and the abrupt introduction of distinctive Late Paleozoic-Cenozoic ages (post-300 Ma) representative of Andean sediment source regions. The Andean signatures are clearly linked to Cretaceous-Paleogene magmatic arc rocks and Permian-Jurassic intrusive rocks exposed in the retroarc fold-thrust belt (Aspden et al., 1987; Cardona et al., 2010b; Horton et al., 2010; Villagómez et al., 2011; Bustamante et al., 2016). Although there was no longer direct input from the craton, it is important to note that cratonic signatures were reduced but not eliminated from Paleogene strata, because these old zircon grains were recycled from pre-Andean basin fill within the fold-thrust belt (Ruiz, 2002; Gombojav and Winkler, 2008; Horton et al., 2010; Bande et al., 2012).

From these data, the timing of sedimentary polarity reversal can be estimated for several regions. In Colombia, the reversal is assigned to the early Paleocene (~60 Ma) in western localities (Magdalena Valley and Eastern Cordillera basins) and to the Oligocene (~30 Ma) in the eastern foreland (Llanos basin). This ~30 Myr discrepancy can be attributed to eastward advance of deformation, with a shift from the Central Cordillera to Eastern Cordillera (e.g., Dengo and Covey, 1993; Cooper et al., 1995; Parra et al., 2009a, 2009b; Mora et al., 2010b, 2013). In Ecuador, a Maastrichtian (~70–65 Ma) reversal is uniformly recognized across eastern to western regions. The consistency in this signal across the Ecuadorian Andes likely reflects the narrow extent and limited shortening of this segment of the orogen (Campbell, 1970, 1974; Baldock, 1982; Baby et al., 2004, 2013).

7.3. Central Andean provenance

U-Pb ages for retroarc basins of the central Andes (Fig. 1) are presented for a southern Peru transect from the Altiplano to Madre de Dios basin (Fig. 8D) and for a southern Bolivia transect spanning the Eastern Cordillera and Chaco foreland basin (Fig. 8E) (Perez and Horton, 2014; Horton et al., 2015b; Perez et al., 2016b; Horton, 2017; Calle et al., 2018).

Pre-Andean signatures defined for Cretaceous strata of southern Peru and Bolivia are characterized by Precambrian-Ordovician ages broadly defined by 450–700 Ma, 1000–1600 Ma, and 1700–2000 Ma groups (Fig. 8D and E). These ages are representative of basement rocks of the southern Peruvian foreland and adjacent parts of the Brazilian shield, including the Sunsas, Rondonia-San Ignacio, and Rio Negro-Juruena provinces (Teixeira et al., 1989; Cordani and Teixeira, 2007; Bettencourt et al., 2010; Bahlburg et al., 2011). The age distributions for overlying Cenozoic strata show several spatial complexities. For southern Peru, the Cenozoic units show a major departure from the pre-Andean strata, with a reduction in basement cratonic ages and a sharp appearance of Andean age components. In the west (Altiplano), this shift is indicated by the introduction of a post-50 Ma unimodal peak derived from the adjacent Western Cordillera magmatic arc (Perez and

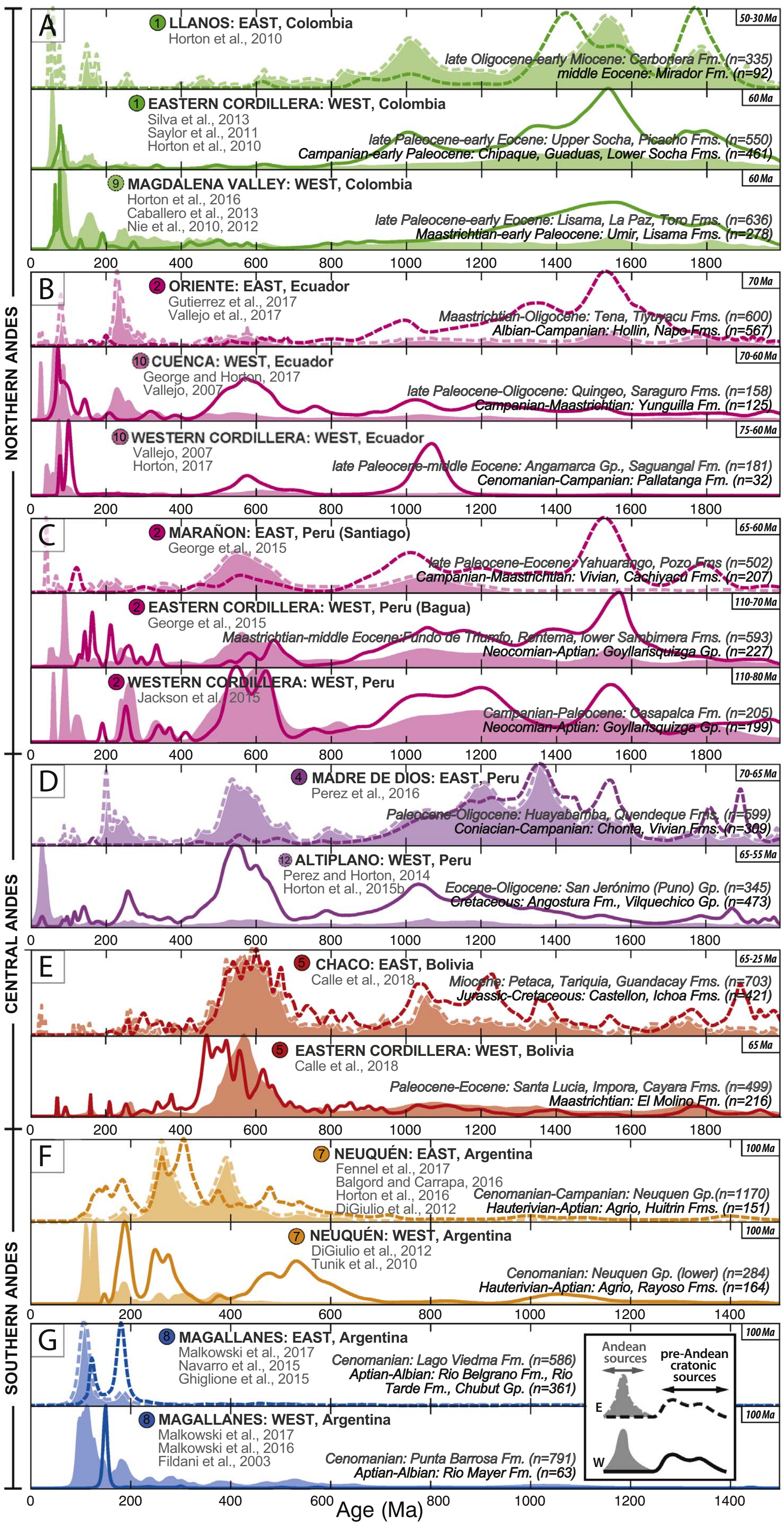

Fig. 8. Comparative plots showing detrital zircon U-Pb age distributions for composite samples from the Andean retroarc basin system, organized by region and age. For each of the 7 regional basin transects (A–G), results are arranged by location, with western localities (solid curves) plotted below eastern localities (dashed curves). For each of the 17 plots, a pair of composite age probability curves allows comparison of the provenance signatures before (open area under curve) versus after (shading under curve) the reversal in sedimentary polarity. The estimated timing of paleodrainage reversal is depicted in the upper right of each plot. Northern Andes: (A) Colombia (Magdalena Valley, Eastern Cordillera, and Llanos basins); (B) Ecuador (Western Cordillera, Cuenca, and Oriente basins); (C) northern Peru (Western Cordillera, Eastern Cordillera, and Marañon (Santiago) basins). Central Andes: (D) southern Peru (Altiplano, and Madre de Dios basins) and (E) southern Bolivia (Eastern Cordillera, Chaco). Southern Andes: (F) central Argentina (western and eastern Neuquén basin) and (G) southern Argentina (western and eastern Magallanes-Austral basin).

Horton, 2014; Horton et al., 2015b). In the east (Madre de Dios basin), however, the shift is manifest as the appearance of 200–300 and 450–750 Ma age groups characteristic of the fold-thrust belt, with no significant appearance of Andean arc detritus. Unlike southern Peru, Cenozoic samples for Bolivia show a limited departure from the pre-Andean age populations, sharing many of the cratonic basement age groups (Fig. 8E).

The contrasting trends for southern Peru and Bolivia may arise from the large magnitude of shortening and basin migration across the central Andes. Whereas the westernmost basin localities (Altiplano) are sensitive to proximal contributions from the Western Cordillera magmatic arc along the western basin margin, localities farther east (Eastern Cordillera and Madre de Dios foreland basin) may have been solely fed by a retroarc fold-thrust belt that attained sufficient topographic expression to prevent delivery of magmatic arc detritus. Large-scale shortening, crustal thickening, and development of a huge salient ("Bolivian orocline") across this wide segment of the fold-thrust belt may have effectively isolated the foreland basin from magmatic arc sources during early uplift of the central Andes. This disconnection apparently persisted for much of Cenozoic time, as evidenced by contrasting hinterland versus foreland provenance patterns and an absence of significant arc detritus in Chaco foreland basin fill (Horton et al., 2002; Pepper et al., 2016; Calle et al., 2018).

7.4. Southern Andean provenance

For retroarc regions of the southern Andes (Fig. 1), detrital zircon U-Pb results are presented for transects across the western and eastern segments of the Neuquén basin (Fig. 8F) and Magallanes-Austral basin (Fig. 8G) (Fildani et al., 2003; Tunik et al., 2010; Di Giulio et al., 2012; Navarro et al., 2015; Balgord and Carrapa, 2016; Horton et al., 2016; Horton and Fuentes, 2016; Malkowski et al., 2016, 2017; Fennell et al., 2017).

In both basins, a provenance change of Cenomanian age is defined by a switch from assorted eastern bedrock sources to western Andean sources of the magmatic arc and fold-thrust belt. Rather than crystalline Precambrian basement of the distal Rio de la Plata craton, the pre-Andean signatures of Lower Cretaceous strata are dominated by eastern contributions from a Jurassic (Chon Aike) igneous province of the Deseado Massif, Permo-Triassic (Choiyoi) igneous province, and subordinate lower Paleozoic igneous-metamorphic rocks (Pankhurst et al., 2000, 2006; Varela et al., 2011). A comparison with overlying strata reveals a pronounced shift, with the introduction of a nearly unimodal 150–100 Ma age group denoting initial contributions from the Andean magmatic arc in the west (Tunik et al., 2010; Di Giulio et al., 2012). This signature, however, is absent from the distal easternmost segments of the Neuquén basin, where Permo-Triassic and lower Paleozoic ages are instead observed.

The reversal in sedimentary polarity can be pinpointed to ~100 Ma judging from depositional age constraints for immediately underlying and overlying stratigraphic units. This moment is interpreted as the onset of Andean shortening and foreland basin genesis. For the Neuquén and Magallanes-Austral basins, the preponderance of magmatic arc signatures expressed in clastic fill during early Andean shortening is consistent with many past stratigraphic and provenance studies of the southern Andes (e.g., Natland et al., 1974; Winn and Dott, 1979; Dott et al., 1982; Biddle et al., 1986; Wilson, 1991; Fildani and Hessler, 2005; Fildani et al., 2008; Romans et al., 2010; Fosdick et al., 2014; Malkowski et al., 2017). As in the central Andes, the explanation for the relative contributions from the magmatic arc may relate to the magnitude of shortening. However, for the southern Andes, the generally low magnitude of shortening—typically < 50 km in the Neuquén region (Manceda and Figueroa, 1995; Ramos et al., 1996; Giambiagi et al., 2012, 2016b; Mescua et al., 2014; Rojas Vera et al., 2014; Sánchez et al., 2015; Fuentes et al., 2016) and < 100 km in the Magallanes fold-thrust belt (Alvarez-Marrón et al., 1993; Klepeis et al., 2010; Fosdick et al., 2011; Ghiglione et al., 2014; Poblete et al., 2014; Betka et al., 2015)—may account for a persistent source-to-sink connection between the magmatic arc and foreland basin.

8. Andean tectonics and basin evolution

Sedimentary basins of western South America are sensitive to the evolution of topography and sediment source regions, making them well suited to address issues of Andean uplift, provenance, and paleodrainage. The integration of diverse datasets provides the underpinnings of a reconstruction of Andean tectonics, with an emphasis on retroarc basins. The generalized five-step reconstruction of the Mesozoic-Cenozoic tectonic history of the Andean orogenic belt and associated basins (Fig. 9) offers a synthesis in accord with available data regarding the stratigraphic framework and deformation timing of retroarc foreland and hinterland basins (Figs. 4 and 5), genesis of unconformities and thin condensed sections (Fig. 6), history of sediment accumulation (Fig. 7), and major shifts in sediment provenance (Fig. 8).

The reconstruction represents a source-to-sink evaluation of sediment accumulation at a regional scale over timeframes exceeding 10^6–10^7 yr. As such, the preserved stratigraphic record, which spans up to 200 Myr, is the net product of sediment supply (from upland weathering, erosion, and transport) and the generation of sediment accommodation by subsidence over a range of basin settings. The competition between sediment supply and accommodation is particularly relevant because (1) the foreland and hinterland stratigraphic records require the presence of appreciable hiatuses, condensed intervals, and disconformities (Fig. 6), and (2) the associated sediment accumulation histories reveal Andean-wide reductions in accumulation rates during Paleogene time (Fig. 7). This situation is comparable to more-localized forebulge dynamics, where underfilled, filled, and overfilled basin conditions (Fig. 6A) illustrate the interplay between sediment supply and accommodation (e.g., Flemings and Jordan, 1989; Crampton and Allen, 1995; Currie, 1997; DeCelles, 2012).

Although principally geared toward the southern and northern Andes, the reconstruction shares components with the history of the highly shortened and thickened central Andes. A central goal is to offer a practical conceptual basis for Andean evolution that emphasizes elements that should be transportable to other convergent systems. Specifically, most of the critical stages outlined here can be securely identified using independent datasets bearing on sediment accumulation, sediment provenance, structural setting, and chronostratigraphic framework.

Key themes include:

- preorogenic extension and postextensional processes;
- the onset of crustal shortening and flexure;
- early orogenic provenance switch and paleodrainage reversal;
- variable magnitude of shortening and foreland basin advance;
- stratigraphic hiatuses and their implications;
- variable rates of sediment accumulation;
- contrasting tectonic regimes and modes of basin genesis; and
- ultimate drivers of orogenesis.

8.1. Pre-Andean extension and thermal subsidence

Mesozoic extension generated a discontinuous series of marginal and backarc basins along most of the western margin of South America (Dalziel, 1986; Mpodozis and Ramos, 1990; Coney and Evenchick, 1994). Although the precise timing is variable along strike, most extension occurred in the Late Triassic to Early Cretaceous window, prior to late Early Cretaceous (~130–120 Ma) breakup of Gondwana. Basin geometries included compartmentalized, fault-controlled half-graben and full-graben (rift) basins succeeded by larger, integrated post-extensional sag basins (Fig. 9A, step 1). Initial short-term, fault-controlled subsidence along normal faults was followed by thermal

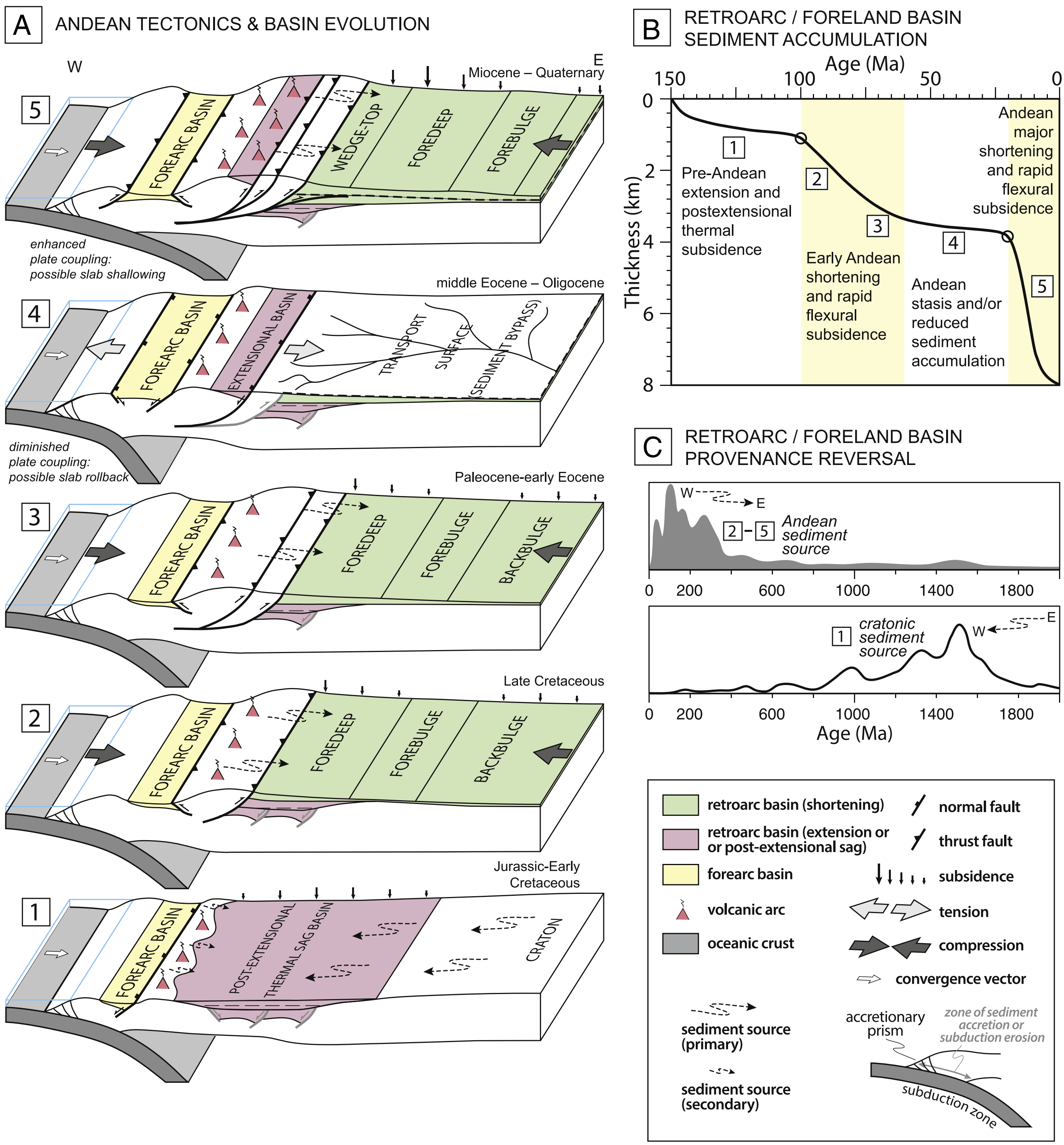

Fig. 9. A generalized five-step reconstruction of the Mesozoic-Cenozoic tectonic history of the Andean orogenic belt and associated basins, including (A) schematic block diagrams (in a similar format as Fig. 3) emphasizing basin evolution, sediment dispersal, arc magmatism and overall tectonic regime (steps 1–5), (B) the associated sediment accumulation history (steps 1–5), and (C) a major shift in sediment provenance recorded by schematic U-Pb age distributions before (step 1) and after (steps 2–5) a wholesale reversal in sedimentary

subsidence driven by long-term, post-attenuation lithospheric cooling (Fig. 9B). Preserved stratigraphic successions of sufficient duration chronicled synrift and postrift phases of sediment accumulation (Fig. 7).

Although a subduction-related magmatic arc occupied far western marginal positions from Late Triassic (or even late Paleozoic) to Early Cretaceous time, sediment delivery to postextensional sag basins was vastly dominated by clastic materials derived from cratonic source regions in the east (e.g., Biddle et al., 1986; Uliana and Biddle, 1988; Di Giulio et al., 2012; Naipauer and Ramos, 2016). Localized examples of arc provenance are present in intraarc and proximal arc positions, but igneous centers were commonly surrounded by marine conditions and were not sites of significant topography (e.g., Vergara et al., 1995; Boekhout et al., 2012). This westward sedimentary polarity (Fig. 9C) is well expressed in detrital zircon provenance data along the length of the Andean plate margin (Fig. 8).

8.2. Onset of Andean orogenesis and reversal in sedimentary polarity

The sedimentary record of initial Andean shortening is distinctly defined by the onset of rapid sediment accommodation and a large-scale sediment routing reversal. Growth of initial topography during Late Cretaceous-Paleocene shortening provided the earliest thrust loads and the resulting flexural genesis of a retroarc foreland basin (Fig. 9A, step 2). Shortening-induced early Andean topography far exceeded the local volcanic edifices of the magmatic arc that had formerly existed along the westernmost plate margin in a fringing off-edge position. Many fold-thrust structures were guided by precursor extensional structures, with common reactivation of inherited pre-Andean structural and stratigraphic inhomogeneities (e.g., Mora et al., 2006; Giambiagi et al., 2008; Savignano et al., 2016). The initial flexural wavelength of the earliest foreland basin commonly matches the areal distribution of the precursor basin, consistent with the interpretation of retroarc foreland basins as successor basins to earlier extensional backarc or postextensional sag basins (e.g., Hsü, 1997; Fosdick et al., 2014). The onset of flexural subsidence is registered by an inflection point in time versus thickness plots (Fig. 9B), signaling an increase in accommodation generation. It seems probable that some degree of enduring thermal subsidence produced a modest amount of additional accommodation space (e.g., Starck, 2011). The switch to rapid accumulation is accompanied by a reversal in sediment provenance, marked by the abrupt appearance of Andean detritus (Fig. 9C).

The contemporaneous shift in accumulation (Fig. 7) and sedimentary polarity (Fig. 8) occurred at ~100 Ma in the southern Andes, and ~70–60 Ma in the remainder of the orogen. This age discrepancy may underscore significant along-strike variability in initial mountain building or may be an artifact of selective erosional removal of the earliest retroarc foreland basin fill due to large-scale shortening and cratonward growth of the orogenic wedge. The latter option is particularly attractive for the central Andes, where there is no definitive pre-Cenozoic record of foreland basin conditions, despite evidence for Late Cretaceous shortening directly to the west in the present forearc of northern Chile and Peru (e.g., Chong, 1977; Cobbing et al., 1981; Vicente, 1990; Benavides-Cáceres, 1999; Arriagada et al., 2006; Bascuñán et al., 2016).

8.3. Early Andean growth of foreland basin

Continued shortening, crustal thickening, and topographic buildup promoted eastward advance of the growing orogenic wedge (Fig. 9A, step 3). Eastward migration of the early Andean foreland basin resulted in systematically younger ages of the provenance switch from cratonic to Andean sources. For example, in the northern Andes of Colombia, stratigraphic sections of the Magdalena Valley, Eastern Cordillera, and Llanos basins (Fig. 8A) show a sedimentary polarity reversal of Maastrichtiaan-early Paleocene age in the west versus an Oligocene age in east (Gómez et al., 2003; Parra et al., 2009a; Horton et al., 2010, 2015a; Nie et al., 2010, 2012; Saylor et al., 2011).

The activation and propagation of new thrust faults facilitated eastward migration of the Andean topographic load and associated cross-sectional (east-west) flexural subsidence profile. Whereas rapid foredeep subsidence persisted in proximal zones (Fig. 9B), the eastward-advancing distal forebulge and backbulge zones underwent limited accumulation, or possibly erosion. This distal basin record is crucial for understanding the central Andes, where some major stratigraphic hiatuses (Figs. 4C and 5A) represent past forebulge positions (Fig. 6A–C) and facilitate protracted tracking of foreland basin migration (e.g., Horton et al., 2001; DeCelles and Horton, 2003). In terms of sediment provenance, in the northern and southern Andes, the magmatic arc remained a major source of sediment, but with progressively greater contributions from the fold-thrust belt. In the central Andes, the exceptional width of the orogenic wedge (Fig. 2) likely accounted for the absence of magmatic arc detritus throughout the Cenozoic succession in the Madre de Dios and Chaco foreland basins (Perez et al., 2016b; Calle et al., 2018).

8.4. Possible pause in Andean shortening with foreland abandonment

A reduction or pause in Andean shortening may have driven limited accumulation or large-scale abandonment across the retroarc foreland basin (Fig. 9A, step 4). In retroarc regions, particularly in the Neuquén basin, an apparently neutral tectonic regime was manifest as a late Eocene–early Miocene cessation of shortening in the frontal fold-thrust belt, with little to no accommodation (Fig. 9B) and a protracted stratigraphic discontinuity in the form of a highly condensed section or unconformity (Fig. 6D–F). In the corresponding magmatic arc to forearc regions, an extensional tectonic regime controlled the focused growth of extensional basins, commonly reactivating early Andean thrust faults (e.g., Charrier et al., 2002; Iannelli et al., 2017).

The transition zone between the central and southern Andes provides the clearest record of these processes, with a major foreland hiatus in the Neuquén basin contemporaneous with normal faulting and extensional basin formation in the adjacent hinterland to forearc regions to the west, including the Cura-Mallin and Abanico basins of westernmost Argentina and Chile (Suárez and Emparan, 1995; Jordan et al., 2001a; Charrier et al., 2002, 2007; Burns et al., 2006; Folguera et al., 2010; Horton and Fuentes, 2016; Horton et al., 2016; Horton, 2018). Nevertheless, a clear Paleogene reduction in accommodation (Fig. 7) is also expressed in retroarc regions of the northern Andes. For the high-shortening segments of the central Andes, such a neutral to extensional phase may have been precluded or overprinted by enhanced coupling during a contemporaneous phase of flat-slab subduction (James and Sacks, 1999; Kay et al., 2005; Ramos and Folguera, 2009; Mamani et al., 2010).

8.5. Main phase of Andean shortening and basin subsidence

The most widely recognized stage of Andean orogenesis involves Neogene shortening and flexural subsidence across nearly all retroarc foreland localities (Fig. 9A, step 5). A sharp increase in sediment accumulation can be linked to cratonward propagation of the locus of thrust loading (Figs. 7 and 9B). Advance of the orogenic wedge also caused continued structural dismemberment of older parts of the foreland basin, and in some cases, growth of compartmentalized hinterland basins such as the Altiplano-Puna plateau (Allmendinger et al., 1997; Horton et al., 2002; DeCelles and Horton, 2003; Sobel et al., 2003; Murray et al., 2010) and the Magdalena Valley basin (Dengo and Covey, 1993; Cooper et al., 1995; Gómez et al., 2005; Moreno et al., 2011; Wolaver et al., 2015). In places where shortening jumped abruptly eastward, the original foreland region was structurally partitioned into a broken foreland basin (Fig. 3H). Well-preserved growth stratal relationships (Fig. 6G and H) document the precise timing of individual structures within the fold-thrust belt and/or adjacent to basement-involved foreland block uplifts (e.g., Zapata and Allmendinger, 1996; Vergés et al., 2007).

For regions that experienced earlier extension, Neogene shortening commonly resulted in tectonic inversion of Paleogene extensional basins (e.g., Godoy et al., 1999; Folguera et al., 2006; Rojas Vera et al., 2014). This main phase of Andean shortening is also the period for which additional tectonic processes have been identified, including possible lithospheric removal, rapid plateau uplift, and dynamic interactions among climate, erosion and fold-thrust activity (Masek et al., 1994; Horton, 1999; Montgomery et al., 2001; Strecker et al., 2007; McQuarrie et al., 2008a; Norton and Schlunegger, 2011; Barnes et al., 2012; Whipple and Gasparini, 2014; Armijo et al., 2015; DeCelles et al., 2015a; Schoenbohm and Carrapa, 2015).

9. Discussion and further research

This orogen-scale synthesis highlighting foreland and hinterland sedimentation yields several insights and new avenues for research on the Andes and other ocean-continent convergent margin orogenic systems. Given the inherent difficulty in directly dating crustal faults, limited resolution of most cross-cutting structural relationships, and the lack of Andean-wide metamorphism, the stratigraphic record provides the principal opportunities for dating Andean orogenesis. Such dating takes the form of growth strata coupled to individual structures (Figs. 5 and 6), the onset of rapid flexural subsidence (Fig. 7), the wholesale reversal in sedimentary polarity during initial topographic growth (Fig. 8), and the introduction of new sediment sources during progressive deformation and orogenic growth (Fig. 9).

Andean-type systems display a wide range of sedimentary basins in diverse structural settings with varied modes of subsidence (Fig. 3). A common temporal sequence for Andean retroarc basins (Fig. 9) entails pre-Andean fault-related extension (synrift phase), postextensional thermal subsidence (sag phase), initial flexural subsidence during Andean shortening, and a later diversified history reflecting contrasting retroarc conditions. Whereas many low-shortening segments of the southern Andes recorded a stratigraphic hiatus associated with a neutral tectonic regime (Fig. 3I) prior to final Andean deformation (Fig. 3G), high-shortening segments of the central Andes recorded continuous advance of the orogenic wedge and the genesis of hinterland basins and/or a broken foreland basin (Fig. 3H). Not surprisingly, the sedimentary record proves critical in the consideration of the matters outlined below.

9.1. Evaluating paleodrainage and topographic divides

Erection of an integrated topographic barrier along the length of the Andes profoundly affected sediment dispersal, climate, and life across South America (e.g., Hoorn et al., 2010, 2017; Mora et al., 2010a; Baker et al., 2014). Such a barrier helped govern biological separation events, species distributions, and biodiversification (Gentry, 1982; Albert et al., 2006; Antonelli et al., 2009), as well as orographic climate changes such as forearc aridification and focusing of rainfall in retroarc regions (e.g., Montgomery et al., 2001; Strecker et al., 2007; Mora et al., 2008; Norton and Schlunegger, 2011). Future orogen-scale syntheses will benefit from the unification of geological records and expanded stable isotopic estimates of paleoelevation (e.g., Gregory-Wodzicki, 2000; Blisniuk and Stern, 2005; Saylor and Horton, 2014; Anderson et al., 2015; Quade et al., 2015; Lamb, 2016; Garzione et al., 2017), along with careful consideration of the isotopic effects of climate change and water sources (Ehlers and Poulsen, 2009; Insel et al., 2010; Poulsen et al., 2010; Jeffery et al., 2012).

Here, the focus is on the ubiquitous provenance switch from cratonic sources to early Andean sources (Fig. 8), which represents a defining moment in South American landscape evolution (Fig. 9). This continental-scale reversal in sedimentary polarity reflects an abrupt influx of synorogenic sediment during earliest Andean shortening and foreland basin development. For individual segments of the Andes, the reversal appears to be a solitary event corresponding to the establishment of a topographic divide that persisted throughout the Andean orogeny. However, this simplified view requires further consideration, particularly in terms of (1) the north-south continuity of the topographic divide and (2) the east-west drainage configuration of the Andean foreland and interior South America.

Spatial variations in the provenance reversal suggest along-strike (north-south) differences in the establishment of a divide. From the provenance results summarized here (Fig. 8), a topographic divide was in place by ~100 Ma in the southern Andes and by ~70–60 Ma in the central and northern Andes. The prevalence of magmatic arc detritus in foreland basin fill is consistent with a topographic divide roughly corresponding to the trace of the magmatic arc. A notable exception, however, is the central Andes, where a dearth of arc detritus in the foreland points to a topographic divide within the retroarc fold-thrust belt, probably near the Altiplano-Eastern Cordillera boundary (Lamb et al., 1997; Horton et al., 2002; Calle et al., 2018). This scenario can be attributed to the relatively large magnitude of shortening (> 150–300 km), crustal thickening, and orogenic advance (McQuarrie, 2002a, 2002b; DeCelles and Horton, 2003; McQuarrie et al., 2005).

The past position of the continental drainage divide was undoubtedly influenced by patterns of deformation and magmatism, as regulated by erosion on opposing forearc and retroarc flanks. Although the divide may have remained relatively fixed along the magmatic arc, possible gaps or portals have been proposed for low-shortening segments of the southern Andes (Patagonia incursion at 40°–45°S; Nielsen, 2005; Bechis et al., 2014; Encinas et al., 2014, 2016) and northern Andes (Guayaquil Gap or Marañon Portal at ~0°–5°S; Grabert, 1971; Hoorn et al., 1995, 2010; Lundberg et al., 1998; Potter, 1997; Mapes, 2009; Sacek, 2014). These localities may have experienced Paleogene extension, which could have further reduced Andean topography (see Section 6.3). From the available evidence, it is proposed that a continuous topographic divide from roughly 5°S to 40°S was established in latest Cretaceous-Paleocene time, coincident with the magmatic arc or high hinterland topography of the retroarc fold-thrust-belt. However, final unification of the northernmost and southernmost Andean segments and completion of a single integrated barrier from > 10°N to > 50°S, remains a topic for future research.

A further complexity concerns the evolution of transcontinental river drainage systems. Although the provenance reversal defines the rapid elimination of a westward (Pacific) routing pre-Andean landscape, the newly established eastward routing Andean rivers did not necessarily form a fully transcontinental eastward (Atlantic) draining landscape. This critical point is underscored by provenance data showing an across-strike (west to east) spatial progression in the timing of drainage reversal. In the northern Andes, the reversal in sedimentary polarity occurred at ~60 Ma in the west and ~30 Ma in the east. This time-transgressive behavior indicates that east-flowing rivers adjacent to the early Andes were restricted to proximal basin segments and did not cross the entire Andean foreland.

In evaluating possible explanations, it is appealing to emphasize the competition between sediment supply and accommodation, such that a transition from underfilled to overfilled foreland basin conditions (Fig. 6A) may account for the onset of transcontinental drainage. This transition could be the product of increased Andean denudation and sediment delivery (e.g., Mora et al., 2010a; Roddaz et al., 2010), river deflection by basement arches (e.g., Hoorn et al., 1995; Costa et al., 2001; Anderson et al., 2016), or regional subsidence patterns driven by subducted slab dynamics and mantle convection (Shephard et al., 2010; Sacek, 2014). Regardless of the precise trigger of transcontinental drainage, it will be vital in future studies to track the course of past rivers (e.g., Amazon, Parana, Orinoco, and Magdalena rivers) and identify the precise arrival of Andean detritus in distal stratigraphic records of the Andean foreland, South American interior, and Atlantic margin (e.g., Hoorn et al., 2017; van Soelen et al., 2017).

9.2. Recognizing and interpreting stratigraphic hiatuses

Identification of angular unconformities and growth strata has been fundamental to defining Andean deformation (Steinmann, 1929; Newell, 1949; Mégard et al., 1984; Noble et al., 1990; Zapata and Allmendinger, 1996; Gómez et al., 2005; Perez and Horton, 2014; Horton et al., 2015b; Echaurren et al., 2016; Fennell et al., 2017). However, many stratigraphic discontinuities appear to be marked by thin, condensed stratigraphic sections or non-angular unconformities (Figs. 4 and 5). An important challenge in Andean retroarc basins is the detection of such discontinuities, in which commonly Upper Cretaceous–Paleocene distal facies are capped concordantly by Cenozoic nonmarine strata of uncertain age. Documenting the existence and duration of these discontinuities (Fig. 6) may be enabled through the recognition of thick pedogenic intervals, presence of an ultrastable conglomeratic lag (representing a transport surface), or by absolute dating of deposits directly above and below the discontinuity (e.g., del Papa et al., 2010; Horton and Fuentes, 2016; Horton et al., 2016; Miall, 2016).

There appear to be three principal ways in which disconformities and condensed sections form in foreland basins, as dictated by the ratio of sediment supply to accommodation generation. First, cessation of retroarc subsidence may define a break between a precursor thermal sag and an initial flexural basin (e.g., Sempere et al., 2002; Gianni et al., 2015; Navarro et al., 2015; Calle et al., 2018). This process would be registered as a hiatus at the very base of the foreland basin succession. Second, in forebulge settings, variations in the ratio of sediment supply to accommodation generation alternately induce erosion (underfilled basin), sediment bypass (filled basin), or limited accumulation of a condensed interval (overfilled basin) (Fig. 6A; Currie, 1997; DeCelles, 2012). Forebulge unconformities would be most common in distal foreland records, within the lower levels of foreland successions. Third, in the midst of an orogeny, the development of an abandonment surface representing sustained exposure and/or transport across the foreland (Fig. 6D) could arise from a shift to a neutral tectonic regime with no flexural accommodation (e.g., Heller et al., 1988; Legarreta and Uliana, 1991). This situation may occur across proximal to distal zones and therefore characterize any part of a foreland basin succession.

The Paleogene stratigraphic record of the Andean foreland and hinterland is noteworthy for the prevalence of prolonged discontinuities (Figs. 4 and 5) and reduced accumulation (Fig. 7). Although these patterns could be related to diminished postrift thermal subsidence, evidence of earlier Andean shortening suggests that the Paleogene history of reduced accommodation was more likely linked to forebulge dynamics or a phase of tectonic quiescence. Most interpretations of a forebulge unconformity center on the central Andean record of southern Peru, Bolivia, and northernmost Argentina (Horton and DeCelles, 1997; Horton et al., 2001; DeCelles and Horton, 2003; Louterbach et al., 2017), in accord with high shortening and large-scale orogenic advance. Low-shortening regions of the northern and southern Andes appear more consistent with a phase of neutral foreland conditions and no shortening-induced flexure, contemporaneous with extension in hinterland, magmatic arc, and forearc regions (e.g., Jordan et al., 2001a; Ramos and Folguera, 2005; Burns et al., 2006; Charrier et al., 2002; Folguera et al., 2010; Horton and Fuentes, 2016; Horton, 2018).

Improved age control and chronostratigraphic correlations are required to fully identify stratigraphic discontinuities and uncover any temporal and spatial trends. For example, are the retroarc hiatuses contemporaneous, time transgressive, or nonsystematic? What are the spatial limits of the hiatuses, and can they be traced laterally into zones of greater accommodation? In terms of their genesis, are the apparent contrasts among the northern, central, and southern Andes a response to variations in shortening magnitude, shortening duration, or other regional or plate-scale factors? Ultimately, the detection of these stratigraphic discontinuities and accurate discrimination among the variable modes of genesis will improve interpretations of the existence, duration, and driving mechanisms of potential tectonic pulses or cycles.

9.3. Identifying variations in deformation mode and magnitude

Although dominated by upper crustal shortening, evolution of the Andes included phases of neutral to extensional deformation. The central Andes are commonly cited as a type example of Andean processes, but this segment of the orogen is representative of only the most intense shortening and surface uplift. In reality, the Andes show highly variable amounts of shortening, crustal thickening, surface uplift, and exhumation, with contrasting modes of deformation reflected in modern stress regimes and basin structural settings (Figs. 1–3). It is proposed here that the Andes can be divided into two endmember categories. High-shortening segments of the Andes sustain > 150–300 km of margin-perpendicular shortening, major crustal thickening (with zones in excess of 50 km thick), and exhibit fluctuations (possibly cyclical) in deformation and magmatism, including rapid orogenic wedge propagation, high-flux magmatic events, lithospheric removal, and extensional collapse (see DeCelles et al., 2009, 2015b; Carrapa and DeCelles, 2015; Ducea et al., 2015; and references therein). In contrast, low-shortening segments of the Andes undergo mostly < 50–100 km of shortening, with commonly mixed modes of deformation consisting of phases (possibly cyclical) of extension or neutral stress conditions (Folguera et al., 2006, 2015b; Ramos et al., 2014; Giambiagi et al., 2015).

Both high- and low-shortening modes involve stratigraphic hiatuses, although potentially of contrasting origin. High-shortening systems exhibit conspicuous cratonward advance of deformation (> 100 km), with a highly migratory forebulge and associated regional stratigraphic hiatus that becomes progressively younger toward the craton (Fig. 3G). In such high-shortening systems, the forebulge hiatus spans 10–20 Myr, assuming typical rates of shortening advance and continental crustal mechanical properties (DeCelles and DeCelles, 2001; Horton et al., 2001; DeCelles and Horton, 2003; DeCelles, 2012). Low-shortening systems have limited advance, a less-mobile forebulge, and better preserved records of mixed-mode deformation incorporating shortening, neutral and extensional conditions (Figs. 3I and 9A).

In assessing Andean tectonic regimes, the Paleogene history is crucial, with evidence of neutral to extensional conditions best represented in the southern Andes. Are the southern Andes unique in experiencing Cenozoic retroarc extension, or has a comparable record been obscured in the northern and central Andes by later shortening, erosion, or overprinting magmatism? Given widespread evidence of inversion tectonics, in which former normal faults are reactivated as thrust or reverse faults (e.g., Cooper et al., 1995; Manceda and Figueroa, 1995; Godoy et al., 1999; Cobbold et al., 2001; Charrier et al., 2002; Kley and Monaldi, 2002; Mora et al., 2006, 2010b; Giambiagi et al., 2008; Mescua et al., 2014; Navarrete et al., 2015; Perez et al., 2016a; Fuentes et al., 2016), could a complex record of Paleogene extension be archived in fold-thrust structures that reactivated older normal fault systems?

Improved delineation of shortening magnitude and discrimination of tectonic regimes will provide insights into key processes involved in the growth of large orogenic belts. Are the central Andes unique in lacking a history of Andean extension, or has that record been obscured by later shortening and synorogenic sedimentation? Uncertainties in the estimates of regional shortening center on the role of early Andean (Late Cretaceous-early Paleogene) shortening in the arc and forearc regions, where the record may be masked by magmatic overprints, buried by younger sedimentary basin fill, and/or removed by tectonic erosion along the subduction interface (e.g., Horton et al., 2001; McQuarrie, 2002a; McQuarrie et al., 2005). Specifically, what are the effects of such cryptic, early Andean shortening on estimates of total shortening and along-strike discrepancies in shortening magnitude?

9.4. Comparing retroarc and forearc basin records

A cross-orogen comparison of retroarc versus forearc stratigraphic and structural records will advance understanding of the spatial scales and regional continuity of Andean orogenic processes. The present synthesis focuses on the retroarc sedimentary archives of foreland and hinterland basins, mostly by virtue of the long-lived, nearly continuous stratigraphic records accessible in such locations. Several scenarios are possible for forearc deformation and basin development (Fig. 3D–F). In fact, a relatively common theme emerges from published literature in which contrasting modes of shortening, extension, neutral, and strike-slip conditions are reported over a range of spatial and temporal scales. Moreover, this debate persists even for the modern record. For example, sharply conflicting interpretations of active shortening and active extension have been advanced in overlapping field areas in the Peru-Chile forearc (e.g., Rutland, 1971; Thornburg and Kulm, 1981; Moberly et al., 1982; Macharé et al., 1986; Sebrier et al., 1988; Hartley et al., 2000; Audin et al., 2008; León et al., 2008; Armijo et al., 2010; Hall et al., 2012; Giambiagi et al., 2016a; Alván et al., 2017; Benavente et al., 2017).

Several considerations may help explain these contradictions in Andean forearc regions. First, modern earthquakes and the Quaternary record of deformation suggest that any active extension of the uppermost crust is of very limited magnitude and concentrated along coastal regions in association with large slip events along the subduction interface (Delouis et al., 1998; González et al., 2003; Loveless et al., 2005, 2010). Second, subduction erosion and tectonic underplating are probably irregular processes affecting selected forearc regions during focused periods, subject to favorable mechanical conditions (von Huene and Ranero, 2003; Clift et al., 2003; Clift and Hartley, 2007). Third, although of limited regional extent, episodic accretion of island arcs, fringing marginal arcs, and/or oceanic materials (including oceanic sediments, ridges, and plateaus) may have exerted a strong control on forearc deformation (Fig. 3A–C) (Kerr et al., 2003; Pindell et al., 2005; Luzieux et al., 2006; Vallejo et al., 2017; Spikings et al., 2015). Fourth, the magnitude of forearc shortening and extension appears to be uniformly low (Isacks, 1988; Muñoz and Charrier, 1996; Wörner et al., 2002; Victor et al., 2004; Farías et al., 2005; Jordan et al., 2010; Noury et al., 2016), possibly allowing for a greater influence of local effects.

It is unclear whether, and to what degree, Andean retroarc regions have been affected by forearc-focused processes such as subduction erosion, tectonic underplating, and arc/oceanic accretion. More generally, do forearc and retroarc regions share similar accumulation histories and comparable shifts in tectonic regime, or do these zones operate somewhat or fully independently? The record of Andean extension is vital. Normal faulting within hinterland regions (including segments of the Altiplano-Puna plateau and its margins) has been shown to be contemporaneous with shortening in the Andean retroarc fold-thrust belt (e.g., Dalmayrac and Molnar, 1981; Suárez et al., 1983; Sebrier et al., 1988; Schoenbohm and Strecker, 2009). Such extension, however, is confined to the highest segments of the Andes (as apparent from earthquake focal mechanisms; Fig. 1A), and appears to reflect extension or partial collapse of highly thickened crust, as modulated by the viscosity and strength of continental crust and lithosphere (Molnar and Lyon-Caen, 1988; Pope and Willett, 1998; Flesch and Kreemer, 2010). In contrast, appreciable extension of forearc regions cannot be linked to extensional collapse of thick crust, and therefore seems incongruent with retroarc shortening (Ramos, 2009; Charrier et al., 2015; Giambiagi et al., 2015, 2016a) and more likely to be emblematic of minor extension owing to local mechanical coupling and/or underplating processes (e.g., Delouis et al., 1998; Adam and Reuther, 2000; González et al., 2003; Farías et al., 2005).

Nevertheless, these interpretations may not be unique, and a cross-orogen comparison with a renewed emphasis on the forearc record should provide answers to several key questions. Fundamentally, do opposing flanks of the Andes operate in unison under comparable tectonic regimes, or do they operate independently and irregularly? Further, what processes govern the evolution of forearc and retroarc structures, with possible out-of-phase behavior expressed in spatially discrete patterns of shortening and extension?

9.5. Assessing the ultimate drivers of contractional orogenesis

The Andes are the product of subduction-related mountain building (Figs. 1–3), yet protracted periods of subduction occurred along the western margin of South America without large-scale shortening of the overriding plate. Such periods of neutral or extensional tectonic regimes are commonly ascribed to rollback of the subducting slab (Dickinson and Seely, 1979; Dewey, 1980; Royden, 1993), which has been expressed as a trench rollback velocity and correlated with orogenic phases in the Andes (Daly, 1989; Russo and Silver, 1996; Ramos, 1999, 2009, 2010). The emerging evidence in stratigraphic and structural records for temporal and spatial fluctuations among shortening, neutral, and extensional conditions (Figs. 4 and 5) introduces a critical element in the consideration of the driving forces behind Andean orogenesis.

The first-order drivers of mountain building are linked to plate dynamics, including relative convergence rates (Pilger Jr., 1984; Somoza, 1998; Pardo-Casas and Molnar, 1987; Daly, 1989; Lonsdale, 2005), absolute convergence rates (Coney and Evenchick, 1994; Silver et al., 1998; Sobolev and Babeyko, 2005; Somoza and Zaffarana, 2008; Maloney et al., 2013), the direction or obliquity of convergence (Somoza, 1998; O'Driscoll et al., 2012), and the width and age of the subducting slab (Schellart, 2008; Capitanio et al., 2011). Other studies have focused on the precise interactions at the subduction interface, in which the degree of mechanical coupling is regulated by plate strength, frictional forces, shear heating, pore fluid pressures, overall thermal structure, and the mitigating effects of lubrication from subducted trench sediments (Jarrard, 1986; von Huene and Scholl, 1991; von Huene and Ranero, 2003; Lamb and Davis, 2003; Vietor and Echtler, 2006; Luo and Liu, 2009; Armijo et al., 2015). Alternatively, processes operating at great depth, including lower mantle resistance to down-dip penetration of the subducting slab, could play an important role in Andean mountain building (Husson et al., 2012; Faccenna et al., 2017). In turn, all of these factors will interact with the stresses associated with the growing topography of the orogen (Coblentz and Richardson, 1996; Iaffaldano and Bunge, 2009; Meade and Conrad, 2008).

Whereas the aforementioned processes operate at plate scales, second-order processes may drive orogenesis at smaller, more-regional scales (Horton, 2018). Flat-slab subduction can enhance plate coupling and drive cratonward advance of deformation (e.g., Jordan et al., 1983; Gutscher et al., 2000; Ramos et al., 2002; Ramos and Folguera, 2009; Folguera and Ramos, 2011). Slab breakoff events may drive not only magmatic processes but also isostatic rebound and associated surface uplift (e.g., Haschke et al., 2002, 2006). Inherited rheological properties of the overriding plate, as related to the age, composition, thickness, strength, and thermal structure of South American crust and lithosphere, may largely reflect pre-Andean processes and guide Andean deformation (Isacks, 1988; Wdowinski and Bock, 1994; Whitman et al., 1996; Allmendinger et al., 1997; Kley et al., 1999; Haschke et al., 2006; Mamani et al., 2008; Luo and Liu, 2009; Giambiagi et al., 2012).

Several motivating questions are as follows. How do variable tectonic regimes and modes of basin development compare with plate convergence parameters? What are the ultimate sources of regional compression in subduction-related mountain building? How are lithospheric-scale stresses manifest in upper crustal structures and associated basins? What controls the spatial and temporal variations in Andean deformation and basin genesis? Further insights from the detrital sedimentary record will be paramount in addressing these questions.

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