



# Unconformity development in retroarc foreland basins: implications for the geodynamics of Andean-type margins

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**Abstract:** Unconformities in foreland basins may be generated by tectonic processes that operate in the basin, the adjacent fold–thrust belt or the broader convergent margin. Foreland basin unconformities represent shifts from high accommodation to non-depositional or erosional conditions in which the interruption of subsidence precludes the net accumulation of sediment. This study explores the genesis of long-duration unconformities (>1–20 myr) and condensed stratigraphic sections by considering modern and ancient examples from the Andes of western South America. These case studies highlight the potential geodynamic mechanisms of accommodation reduction and hiatus development in Andean-type retroarc foreland settings, including: (1) shortening-induced uplift in the frontal thrust belt and proximal foreland; (2) the growth and advance of a broad, low-relief flexural forebulge; (3) the uplift of intraforeland basement blocks; (4) tectonic quiescence with regional isostatic rebound; (5) the end of thrust loading and flexural subsidence during oblique convergence; (6) diminished accommodation or sediment supply due to changes in sea-level, climate, erosion or transport; (7) basinwide uplift during flat-slab subduction; and (8) dynamic uplift associated with slab window formation, slab break-off, elevated intraplate (in-plane) stress, or related mantle process. These contrasting mechanisms can be distinguished on the basis of the spatial distribution, structural context, stratigraphic position, palaeoenvironmental conditions, and duration of unconformities and condensed sections.

**Thematic collection:** This article is part of the Fold-and-thrust belts collection available at: <https://www.lyellcollection.org/cc/fold-and-thrust-belts>

**Received** 31 December 2020; **revised** 30 September 2021; **accepted** 12 October 2021

Unconformities in foreland basins can be attributed to tectonic, climatic, eustatic and internally driven (autogenic) processes. These forcing mechanisms determine zones of erosion, deposition, sediment bypass, sediment starvation and stratigraphic condensation in the fold–thrust belt, foreland basin and adjacent craton (Fig. 1). This paper considers long-duration (>1–20 myr) unconformities and condensed stratigraphic intervals in foreland settings, explores potential modes of hiatus development and evaluates the geodynamic processes affiliated with modern and ancient examples from the Andes and its retroarc foreland basin system.

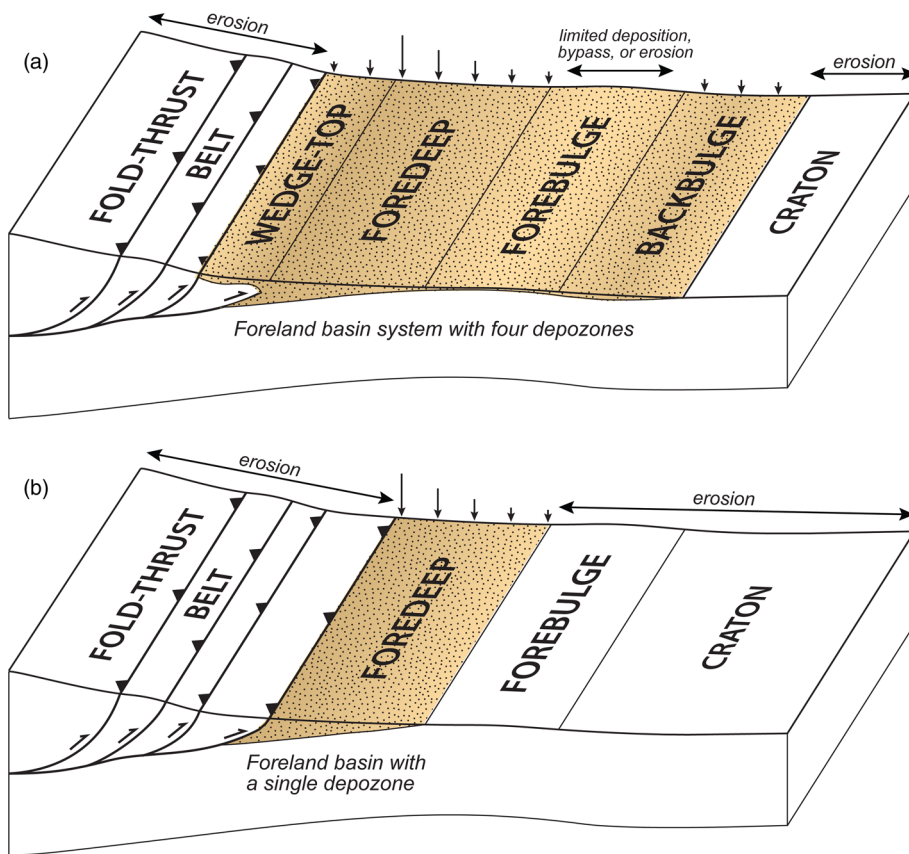
For the purpose of this study, an unconformity is defined as a 3D (planar or non-planar) surface representative of a substantial temporal hiatus resulting from non-deposition or erosion. A condensed section (or zone of stratigraphic condensation) is defined as a concordant stratigraphic interval of limited thickness (generally <5–100 m) produced by slow sediment accumulation with possible intermittent non-deposition and/or minor erosion. Unconformities and condensed sections may form stratigraphic discontinuities between diverse facies in marine or non-marine depositional systems, and may be confined to proximal or distal basin margins or may occur as basinwide features (e.g. Blackwelder 1909; Barrell 1917; Wheeler 1958; Vail *et al.* 1984; Loutit *et al.* 1988; Shanmugam 1988; Clari *et al.* 1995; Miall 2016).

The foreland unconformities and condensed intervals discussed here span >1–20 myr and may be diachronous (time-transgressive). In other systems, the terms paraconformity or diastem may be suitable for a disconformity (non-angular unconformity) or condensed section of shorter duration. To avoid the confusion that may accompany the interpretation of unconformities – such as the debate over the genesis of the first unconformity identified by James

Hutton (e.g. Tomkeieff 1962; Young and Caldwell 2009; Jutras *et al.* 2011) – this study attempts to delineate the specific temporal and spatial framework for the reported stratigraphic discontinuities. For simplicity, the term hiatus (rather than lacuna or vacuity; Wheeler 1958) is used to refer to the period of time demarcated by a particular unconformity or condensed interval.

Foreland basins develop in contractional orogenic systems along convergent plate boundaries, in peripheral and pro-wedge/retro-wedge settings within continental collision zones and in retroarc settings associated with Andean-type subduction margins. Foreland basins record long-term (>10–100 myr) rapid sediment accommodation, principally in response to regional isostatic (flexural) subsidence due to thrust loading and crustal thickening in the adjacent orogenic wedges (Price 1973; Dickinson 1974; Beaumont 1981; Jordan 1981). Foreland basins are also affected by far-field dynamic processes resulting from mantle flow and mechanical coupling between the subducting/underthrusting slab and the overriding plate (Royden 1993; DeCelles and Giles 1996; Liu *et al.* 2011). Most foreland basins can be categorized into two end-member geometries (Jordan 1995; Sinclair 1997; DeCelles 2012): (1) an ‘overfilled’ basin commonly composed of aggradational wedge-top, foredeep, forebulge and backbulge depozones (Fig. 1a); or (2) an ‘underfilled’ basin defined by a single aggradational foredeep depozone bordered distally by a degradational forebulge and craton (Fig. 1b).

Foreland basin unconformities are most readily formed near the erosive basin margins, along both the distal cratonic margin and the structurally disrupted proximal basin margin adjacent to the bounding fold–thrust belt (Fig. 1). Given the long-term cratonward advance of orogenic wedges and genetically linked foreland basins (Bally *et al.* 1966; Dewey and Bird 1970; Coney 1973), these basin



**Fig. 1.** Schematic block diagrams showing the principal zones of deposition and erosion within (a) a complete (overfilled) four-component foreland basin system consisting of wedge-top, foredeep, forebulge and backbulge depozones and (b) an underfilled foreland basin with a single foredeep depozone. Erosion is focused in the flanking thrust belt and craton regions, with potential erosion, sediment bypass or severely limited sediment accumulation across the forebulge region. The relative magnitude of sediment accommodation is denoted by spatial variations in basin-fill thickness (stippled pattern) and flexural subsidence (vertical arrows).

margin unconformities are preferentially concentrated in the lower and upper stratigraphic levels of foreland basin successions (Fig. 2).

In lowermost stratigraphic levels, a basal foreland unconformity may denote a basinwide stratigraphic boundary between pre-orogenic deposits below and synorogenic deposits above. The progressive advance of the thrust-induced flexural load and corresponding foreland subsidence profile (i.e. the flexural wave) generates a diachronous (time-transgressive) hiatus marked by this regional basal contact in which the overlying synorogenic deposits onlap the exposed distal basin margin and become progressively younger toward the craton (Fig. 2) (Cant and Stockmal 1989; Sinclair *et al.* 1991; Crampton and Allen 1995; DeCelles and Giles 1996; Gupta and Allen 2000). An additional basinward younging pattern is recorded in overfilled foreland basins, where a diachronous condensed interval or disconformity symbolizes the cratonward advance of the forebulge (Fig. 2a). In upper stratigraphic levels, local unconformities expressed in the proximal foreland basin register shortening and erosion in the frontal sector of the orogenic wedge. Such wedge-top deposits are distinguished by angular unconformities and growth stratal geometries produced during coarse sediment accumulation in proximity to active fold–thrust structures (e.g. Riba 1976; Anadón *et al.* 1986; Suppe *et al.* 1992; Jordan *et al.* 1993; Horton and DeCelles 1997; Ghiglione *et al.* 2010).

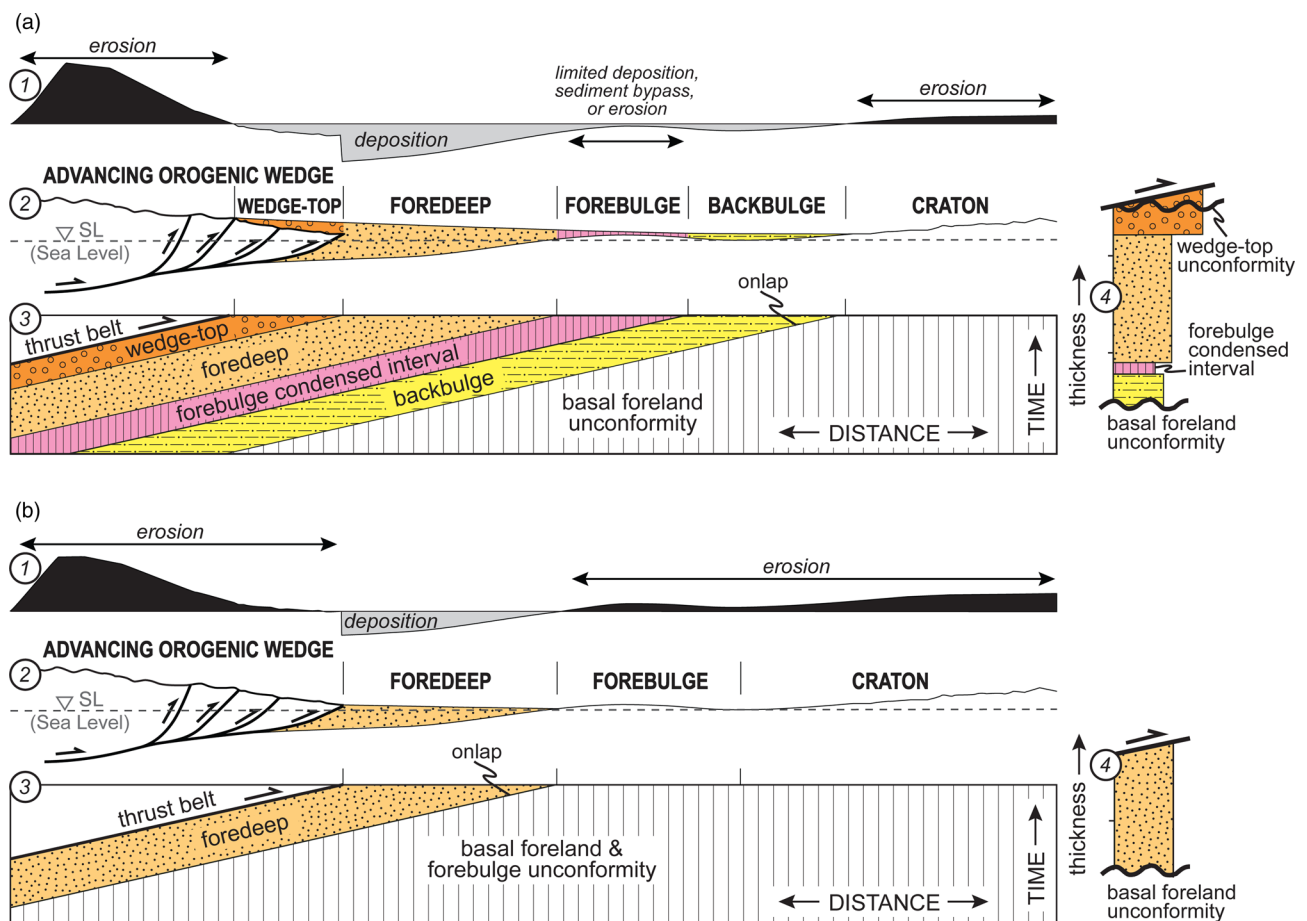
Idealized chronostratigraphic cross-sections or Wheeler diagrams (Wheeler 1958; Miall 2016) highlight the long-term diachronous record of foreland depozones, unconformities and condensed stratigraphic intervals (Fig. 2). In overfilled basin systems (Figs 1a and 2a), a fully developed vertical stratigraphic succession (as per Walther's law) consists of a basal foreland unconformity overlain successively by distal backbulge deposits, a condensed interval (zone of stratigraphic condensation) or a disconformity marking the forebulge, a thick upward-coarsening foredeep section and capping wedge-top deposits with growth strata and structurally controlled discordances. By contrast, underfilled basins (Figs 1b and 2b) are characterized by a basal angular or non-angular unconformity,

which represents protracted erosion across the distal forebulge and craton, overlain by a thick continuous foredeep succession.

Although these end-member cases provide useful templates, they fail to capture the complexity of foreland basins with unconformities that are not confined to the proximal and distal basin margins. This paper explores possible mechanisms for the regional generation of long-duration stratigraphic discontinuities within retroarc foreland basins, with implications for the geodynamics of Andean-type convergent plate boundaries. The motivation is to describe these possibilities and critically assess ancient and modern examples from the Andean foreland basin of western South America. Whereas others have focused on the stratigraphic records of hinterland deformation (e.g. Steinmann 1929; Mégard *et al.* 1984; Noblet *et al.* 1996; Horton 2012), the neotectonics of the modern foreland (e.g. Proyecto Multinacional Andino 2009; Vellozo *et al.* 2012; Folguera *et al.* 2015a; Costa *et al.* 2020) or the long-term evolution of the foreland basin system (e.g. Jordan and Alonso 1987; Cooper *et al.* 1995; Jordan *et al.* 2001a; DeCelles and Horton 2003; Gómez *et al.* 2005; Bayona *et al.* 2008, 2020; Horton *et al.* 2010, 2020; Roddaz *et al.* 2010; Horton 2018a, b), this study highlights the Andean-type geodynamic processes capable of producing foreland stratigraphic hiatuses of considerable duration and spatial extent.

### Mechanisms of unconformity development

Summarized here are potential unconformity generation mechanisms for retroarc foreland basins (Fig. 3). Each is distinct, but they are not mutually exclusive; several different processes may affect a single region and may operate in synchronicity or in temporal succession. The separate options are unified in that each serves to interrupt an otherwise continuous process of rapid sedimentation and accommodation generation (Figs 1 and 2). Of the eight mechanisms discussed, the first three (Fig. 3a–c) create unconformities or condensed stratigraphic intervals in relationship to specific



**Fig. 2.** Idealized diagrams for (a) a complete (overfilled) foreland basin system (Fig. 1a) and (b) an underfilled foreland basin (Fig. 1b) showing (1) the erosion/deposition profile along (2) a regional cross-section (with foreland basin subsidence profile) spanning the thrust belt to craton, with the long-term basin evolution represented by (3) an idealized time-stratigraphic cross-section (Wheeler diagram) and (4) the corresponding schematic stratigraphic succession. Progressive cratonward advance of the orogenic belt and foreland basin (i.e. the flexural wave) results in time-stratigraphic charts showing diachronous (time-transgressive) stratigraphic boundaries (between the deposits of adjacent depozones) and depositional onlap toward the craton (with an increase in hiatus duration toward the craton). Hiatuses (vertical ruled pattern) in the time-stratigraphic charts represent unconformities (non-angular unconformities) and condensed intervals in the stratigraphic record. The upward-coarsening foreland stratigraphic succession reflects successive vertical stacking of laterally adjacent, distal to proximal depozones (per Walther's law) during orogenic advance.

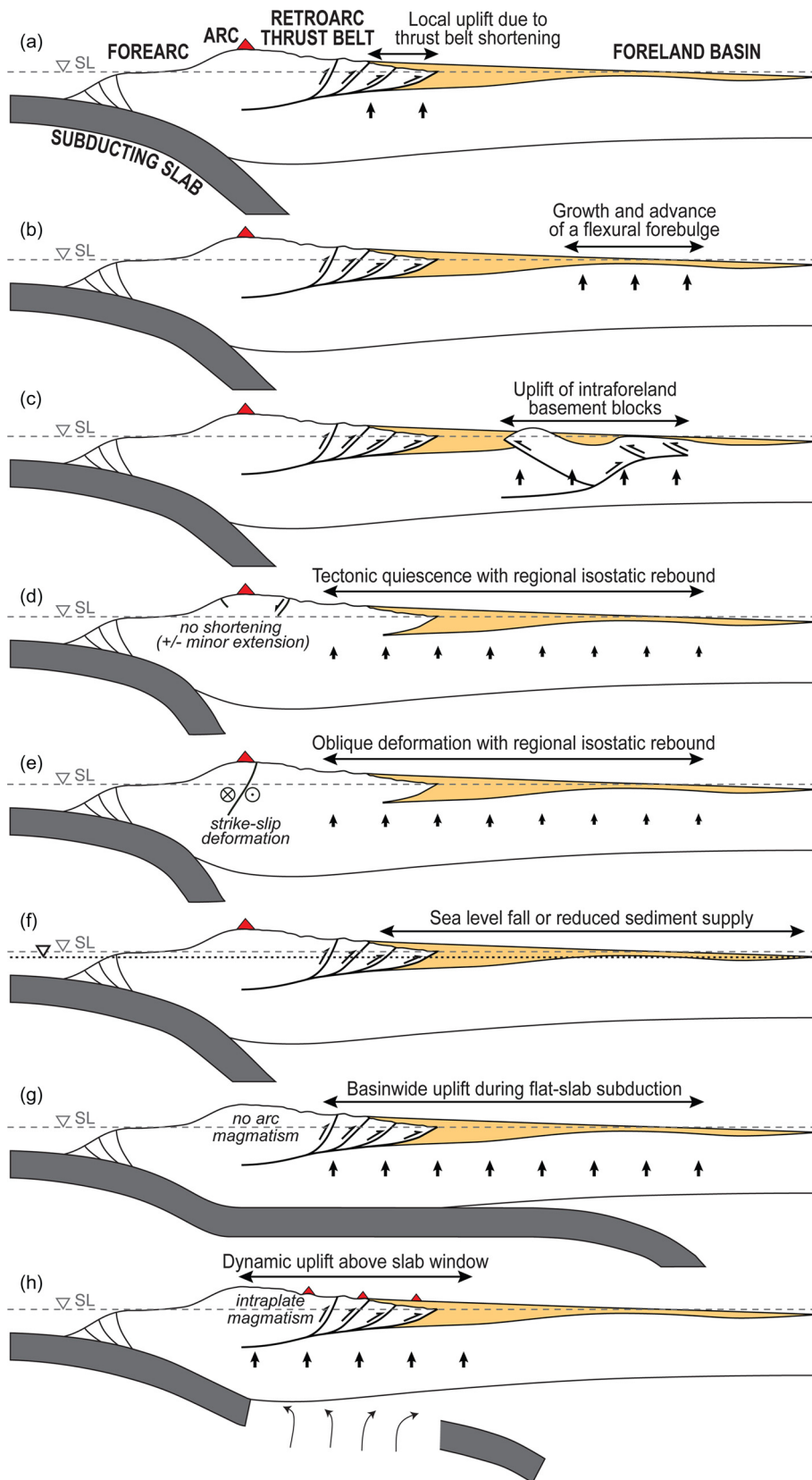
structures or broad structural highs, whereas the others (Fig. 3d–h) involve regional geodynamic processes that may affect large expanses of the foreland basin and its margins. The proposed mechanisms have been identified in previous syntheses and some may apply to non-foreland settings (e.g. *Vail et al.* 1984; *Weimer* 1984; *Loutit et al.* 1988; *Shanmugam* 1988; *Ettensohn* 1994; *Crampton and Allen* 1995; *DeCelles* 2012; *Miall* 2016; *George et al.* 2020).

Unconformities and condensed sections may occur throughout a foreland basin succession and need not be restricted to specific stratigraphic levels. This paper considers long-duration discontinuities that span >1–20 myr. In practice, insufficient age control will pose challenges to the precise quantification of hiatuses. In particular, condensed intervals – where several metres of section represent long periods of time – may contain a series of internal surfaces formed by multiple phases of abandonment or erosion, with the common presence of non-marine pedogenic facies or gravel lag deposits. The following mechanisms (Fig. 3) are regarded as viable explanations for lengthy hiatuses that may be embodied in the stratigraphic record as: (1) an unconformity defined by a single erosive or non-erosive surface; (2) an unconformity capped by thin regolith or gravel lag deposits; or (3) a condensed section potentially rich in palaeosols or other facies indicative of reduced accommodation.

### Shortening in the frontal thrust belt and proximal foreland

Uplift in the frontal fold–thrust belt can produce unconformities in the proximal, wedge-top sector of a foreland basin (Fig. 3a). Progressive syndepositional tilting of basin fill results in growth strata recognized by an up-section decrease in dip, abrupt lateral thickness variations (thinning onto the synchronous fold–thrust structure) and internal angular unconformities (*Biot* 1937; *Riba* 1976; *Anadón et al.* 1986; *Suppe et al.* 1992). The spatial extent of a wedge-top unconformity and growth stratal package is governed by the wavelength of the adjacent structure and is therefore limited to a short horizontal distance (typically <5–10 km) from the syndepositional thrust fault and related fold (e.g. *DeCelles* 1994; *Ghiglione et al.* 2002; *Perez and Horton* 2014). This constraint on the thrust front position through time provides crucial insights for the reconstruction of thrust belt development (e.g. *Jordan et al.* 1993, 2001a; *DeCelles and Giles* 1996; *DeCelles and Horton* 2003; *Ghiglione and Ramos* 2005; *DeCelles et al.* 2011).

The proximity to active structures ensures that wedge-top unconformities and associated growth strata in clastic systems are confined to coarse facies near the mountain front (Figs 1 and 2a). Given this restricted spatial extent, any proximal stratigraphic discordances are expected to pass rapidly into correlative conformable sections in the laterally adjacent foredeep (Fig. 3a).



**Fig. 3.** Schematic cross-sections showing potential geodynamic mechanisms for the generation of unconformities and condensed stratigraphic intervals in a foreland basin, including: (a) localized shortening-related uplift in the frontal thrust belt and proximal foreland; (b) growth and advance of a broad, low-relief flexural forebulge in the distal foreland; (c) uplift of intraforeland basement blocks along crustal-scale reverse faults; (d) tectonic quiescence with regional isostatic rebound; (e) cessation of thrust loading and flexural subsidence during oblique convergence; (f) diminished accommodation or diminished supply due to changes in sea-level, climate, erosion or sediment transport; (g) basinwide uplift caused by increased interplate coupling during flat-slab subduction; and (h) dynamic uplift associated with slab window formation, slab break-off, elevated intraplate (in-plane) stress or related mantle processes.

### **Growth and advance of a broad, low-relief flexural forebulge**

An advancing orogenic wedge may yield a widespread unconformity that denotes a diachronous hiatus induced by the cratonward migration of the flexural wave during progressive topographic loading (Figs 2 and 3b). This long-term process may be manifest not only as a basal foreland unconformity separating pre-orogenic from

synorogenic deposits (e.g. Sinclair *et al.* 1991; Crampton and Allen 1995; Caballero *et al.* 2020), but also as a discrete condensed interval or disconformity indicative of an advancing broad-wavelength, low-amplitude forebulge (Fig. 2a) (Plint *et al.* 1993; DeCelles and Horton 2003; Fuentes *et al.* 2009; DeCelles 2012). The continuous advance of a flexural forebulge at a fixed rate is considered unlikely. Unsteady processes involving irregular advances or punctuated ‘jumps’ in forebulge location are deemed

more probable due to the preferential reactivation of pre-existing structures, stratigraphic anisotropies or inherited basement fabrics (Waschbusch and Royden 1992a; Bayona and Thomas 2003, 2006; DeCelles 2012; Chapman and DeCelles 2015). Moreover, spatial variations in plate strength will influence forebulge migration, with a stronger plate (i.e. higher flexural rigidity and effective elastic thickness) toward the craton fostering a greater flexural wavelength and broader forebulge through time (e.g. Stockmal *et al.* 1986; Waschbusch and Royden 1992b; Fosdick *et al.* 2014).

The regional basin architecture in terms of underfilled/overfilled geometry (Jordan 1995) and the site-specific geomorphic and palaeoenvironmental conditions – most crucially, the local ratio of accommodation to sediment supply (Horton 2018a) – will determine whether the broad, low-relief forebulge will involve (1) erosion, (2) sediment bypass or (3) modest sediment accumulation and, in turn, result in, respectively, (1) an unconformity defined by a single erosional surface, (2) an unconformity potentially marked by relict detritus in the form of a thin gravel lag or (3) a condensed depositional interval routinely demarcated in non-marine clastic systems by pedogenic facies. Preservation of the predicted forebulge stratigraphic record within an overfilled four-component basin system (Figs 1a, 2a and 3b) requires considerable lateral mobility of the fold–thrust belt and therefore will be limited to orogenic systems with large-magnitude shortening and crustal thickening (e.g. DeCelles and DeCelles 2001; Christophoul *et al.* 2003; DeCelles 2012).

### ***Uplift of intraforeland basement blocks***

Intraforeland basement block uplifts disconnected from the thin-skinned fold–thrust belt can produce unconformities and condensed intervals over large swaths of a foreland basin (Fig. 3c). Because many intraforeland uplifts are formed by independent structures deeply rooted in the lower crust, the spatial scale of associated stratigraphic discontinuities can be large (i.e. several tens of kilometres) (e.g. Tweto 1980; McQueen and Beaumont 1989; Erslev 2005; Rudolph *et al.* 2015; Folguera *et al.* 2015c). The interplay between the relative rates of rock uplift, regional flexural subsidence and sediment supply will dictate whether an individual block will be emergent and build positive topography or will remain buried beneath foreland basin fill (e.g. Dickinson *et al.* 1988; Ramos *et al.* 2002; Yonkee and Weil 2015; Lawton 2019). Specifically, an isolated non-emergent block would be represented by a buried structural high that recorded diminished subsidence relative to the adjacent proximal and distal sectors of the foreland basin (Fig. 3c). By contrast, an emergent block that was later buried would prompt an erosional unconformity with a spatial distribution matching the dimensions of the block uplift. However, a block that remained a positive topographic feature for an extended period and never returned to a net accumulation situation would leave no preserved basin-fill record.

Sufficient subsurface control can enable the identification of both emergent and non-emergent basement block uplifts within a single foreland basin (e.g. Balkwill *et al.* 1995; Lalami *et al.* 2020). Although a basement block uplift in the distal foreland may be difficult to distinguish from a flexural forebulge (e.g. Ziegler *et al.* 2002), most intraforeland block uplifts exhibit a lower wavelength (width), a higher structural relief and can be geometrically linked to crustal-scale structures of moderate to large displacement (several to tens of kilometres) (Fig. 3c) (e.g. Bayona and Thomas 2003, 2006).

### ***Tectonic quiescence with regional isostatic rebound***

A prolonged pause in crustal shortening and flexural loading may promote isostatic rebound and the development of an unconformity across the foreland basin (Fig. 3d). During a period of tectonic quiescence, continued erosion would decrease the orogenic topography, causing flexural unloading and minor isostatic uplift

that would diminish in magnitude toward the craton. This process would yield an unconformity characterized by erosion in the proximal foreland and sediment bypass or severely reduced accommodation in the medial to distal foreland (e.g. Heller *et al.* 1988; Cant and Stockmal 1989; Legarreta and Uliana 1991; Ross *et al.* 2005; Morin *et al.* 2019).

Alternating short-term phases of thrust fault activity and quiescence have been invoked to explain foreland hiatuses (Flemings and Jordan 1990; Miall 1996; Catuneanu *et al.* 1997; Houston *et al.* 2000; Londono *et al.* 2012). Such episodic thrusting appears intuitively appealing because the seismogenic records of individual faults define cyclical patterns over periods <10 kyr (Avouac 2003; Allmendinger *et al.* 2009). However, over longer time frames (>1 myr), thrust faults and kinematically linked fold–thrust systems show sustained shortening without evidence of orogen-wide episodicity (Jordan *et al.* 1993, 2001a; Beaumont *et al.* 2000; Allmendinger and Judge 2014; Mouthereau *et al.* 2014; Anderson *et al.* 2018). Therefore, long pauses in regional shortening are not an intrinsic or autogenic component of contractional orogenic belts and require specific mechanisms of sufficient duration.

Rather than autogenic behaviour, long-term (>1 myr) cessation of shortening in a fold–thrust belt may be the expression of persistent tectonic quiescence due to lowered convergence rates and/or a shift to a neutral or modestly tensile stress regime (Fig. 3d). Commonly, these drivers may be genetically related to slab rollback (retrograde slab motion) during steepening of the subducting/underthrusting slab, potentially within a retreating plate boundary involving reduced mechanical interplate coupling between the converging plates (e.g. Dewey 1980; Royden 1993; Barberón *et al.* 2018; Horton 2018b; Fernández Paz *et al.* 2019).

### ***Cessation of thrust loading and flexural subsidence during oblique convergence***

A foreland basin unconformity may be generated by the termination of flexural subsidence during a shift from a contractional to largely strike-slip tectonic regime (Fig. 3e). The obliquity of plate convergence is an important control on the amount of strike-slip and contractional deformation in the overriding plate. Strain partitioning along a convergent margin during non-orthogonal plate motion (Fitch 1972; Jarrard 1986) is likely to regulate the relative proportions of strike-slip displacement and crustal thickening within the orogenic system, including structures in the forearc, magmatic arc and retroarc fold–thrust belt. A shift to highly oblique convergence has been recognized as a trigger for the termination of orogenic thickening, crustal loading and flexural subsidence in an adjacent foreland (e.g. Price 1994; Jaillard and Soler 1996; Simony and Carr 2011).

Although similar to the preceding option of a regional stratigraphic hiatus related to erosional unloading and minor flexural rebound (Fig. 3d), the distinction in this case (Fig. 3e) is that the convergent margin experiences not only a lessening in the rates of orthogonal (trench-normal) convergence, but also an increase in strike-slip deformation. Ultimately, the consequences within the foreland basin are similar, with the establishment of an extensive erosional unconformity that diminishes in magnitude toward the craton. A potential alternative is that an inboard jump in strike-slip deformation may compartmentalize the foreland basin into a series of smaller strike-slip basins (e.g. de Vicente *et al.* 2011).

### ***Diminished accommodation or sediment supply due to changes in sea-level, climate, erosion or transport***

A large decline in either accommodation or sediment supply may induce an unconformity across a foreland basin (Fig. 3f). Under appropriate conditions, a fall in relative sea-level could effectively

eliminate sediment accommodation across the basin. Basinwide erosion will only occur if the magnitude of the drop in regional base level is sufficiently large to exceed accommodation generation by flexural and dynamic mechanisms (Vail *et al.* 1984; Shanmugam 1988; Jordan and Flemings 1991; Dickinson *et al.* 1994; Clevis *et al.* 2004). In such situations, given the low topographic gradients across most foreland basins, the magnitude of erosion should be relatively uniform across the width of the basin (Fig. 3f). The resulting stratigraphic record is apt to be distinguished by widespread incision, with cutting of incised valleys several tens to possibly 100–200 m deep (e.g. Weimer 1984; Van Wagoner 1995; Plint *et al.* 2012).

Alternatively, in the absence of a base level change, the same effects could be accomplished by modifications in climate or erosional intensity that affect sediment supply, transport capacity and stream power. As examples, long phases of non-deposition, sediment bypass or erosion may be the respective outcomes of: (1) a sharply reduced sediment supply due to aridification or minimized weathering; (2) a shift to highly efficient sediment transport across the basin; or (3) intensified erosion that leads to the evacuation of foreland basin fill (e.g. Blisniuk *et al.* 2005; Clift 2010; Clift and Van Laningham 2010; Allen *et al.* 2013).

For orogenic wedges governed by critical taper mechanics (Dahlen and Suppe 1988), changes in climate and erosion can further regulate mass influx and outflux for the orogen and adjacent foreland, with potential shifts in the loci of crustal shortening and exhumation. Such orogenic self-organization would interact with internal (autogenic) and external agents affecting erosional/depositional mass budgets and basin sedimentation (Flemings and Jordan 1989; DeCelles and Mitra 1995; Horton 1999; Sobel *et al.* 2003; Sinclair and Naylor 2012; Ghiglione *et al.* 2019).

### ***Basinwide uplift during flat-slab subduction***

Basinwide uplift in retroarc regions may arise from the mechanical effects of a shift from normal or steep-angle subduction to flat-slab subduction (Fig. 3g). With greater compression, end-loading and/or basal traction in the overriding plate (Bird 1984; Axen *et al.* 2018), an accompanying reversal from subsidence to uplift in the foreland basin would eliminate accommodation and instigate pervasive erosion. This shift would be manifest as a regional erosional unconformity spanning the interval of flat-slab conditions, with a potentially younging trend toward the craton (reflecting the inboard advance of the subducted slab at depth). However, the geodynamic responses to subduction shallowing are complex and it is recognized that the precise effects of flat-slab subduction are varied and often unclear, with some studies proposing increased subsidence rather than uplift (e.g. Mitrovica *et al.* 1989; Liu *et al.* 2011; Flament *et al.* 2015; Gianni *et al.* 2018; Carrapa *et al.* 2019; Dávila *et al.* 2019; Saylor *et al.* 2020).

In the simplified model shown here (Fig. 3g), flat-slab subduction is considered to be responsible for increased mechanical coupling between the converging plates and the enhanced inboard transmission of compressional stress. However, similar effects across the foreland basin could be activated by long-distance material transport via crustal underthrusting, lower crustal flow, ductile injection or crustal translation above a deep décollement (e.g. Oldow *et al.* 1989; McQuarrie and Chase 2000; Erslev 2005). The broad wavelength of these processes suggests that they would not cause sharp angular discordances, but would lead to widespread exhumation of foreland basin fill.

### ***Dynamic uplift associated with slab window formation, slab break-off or elevated intraplate stress***

Dynamic uplift unrelated to thrust belt shortening and flexural loading may involve several potential geodynamic mechanisms for the origination of regional unconformities spanning foreland basins.

The formation of a slab window during the subduction of an actively spreading oceanic ridge can drive uplift of continental lithosphere above an expanding gap in the subducted slab (Fig. 3h) (Thorkelson 1996). Similarly, the tearing or break-off of a subducted slab may catalyse surface rebound and pronounced basinwide uplift (Davies and von Blanckenburg 1995; Andeweg and Cloetingh 1998; Gianni *et al.* 2018, 2019). Slab window formation, slab tearing and slab break-off promote uplift by way of heating and/or asthenospheric upwelling (Fig. 3h). These have a profound effect on magmatism, provoking a switch from subduction-related continental arc magmatism to anomalous intraplate (within-plate) magmatism with strong mantle/asthenospheric signatures.

At similar scales, the possible far-field transmission of intraplate (in-plane) compressional stress from a plate boundary may induce large-wavelength lithospheric folding or buckling (Ziegler *et al.* 2002; Kley and Voigt 2008; Cloetingh and Burov 2011; Lacombe and Bellahsen 2016). Alternatively, vertical motion associated with plume activity or mantle flow may yield broadly similar uplift patterns in the absence of elevated horizontal compressional stress (Burov and Cloetingh 2009; Faccenna and Becker 2020).

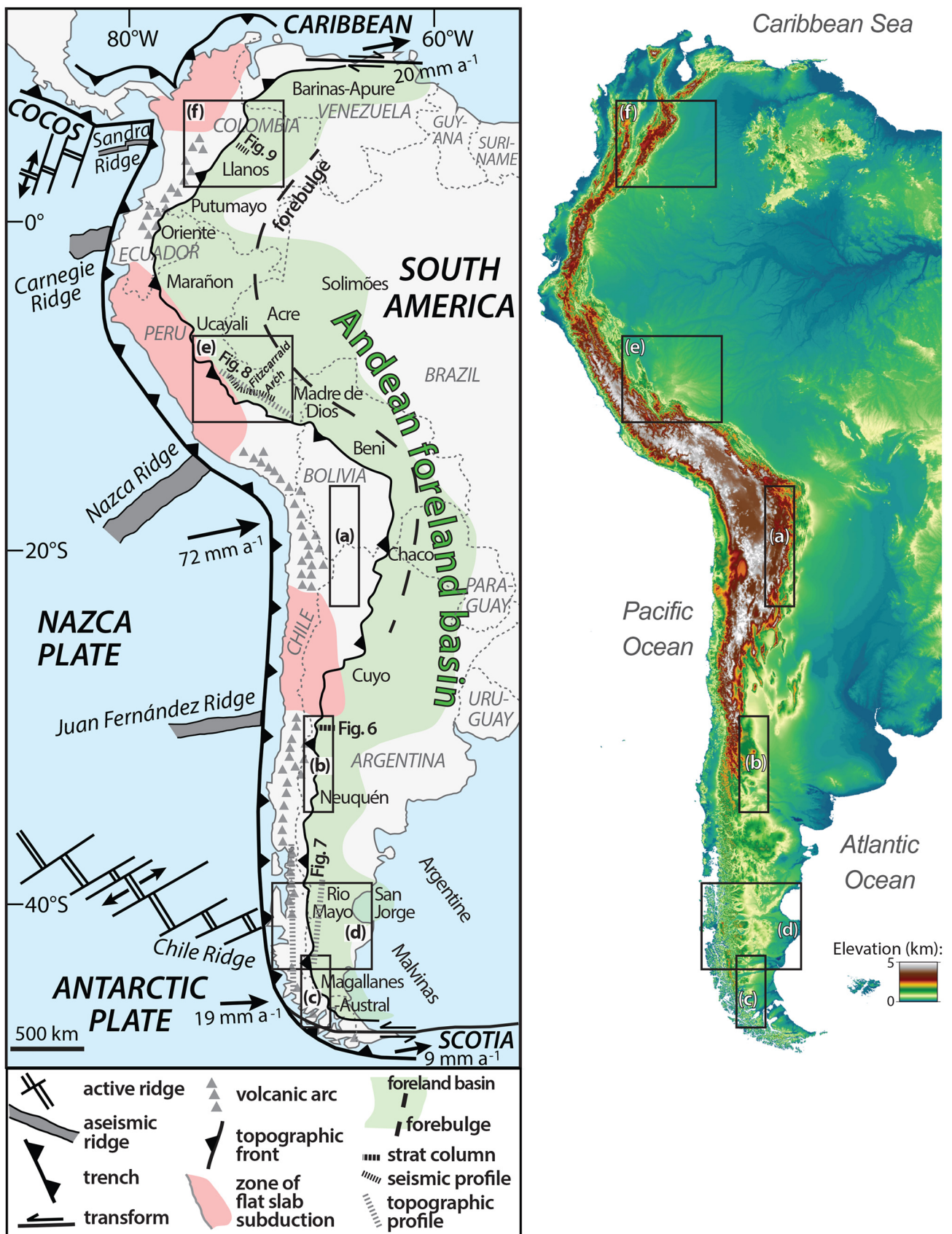
The aforementioned geodynamic mechanisms mostly reflect dynamic processes linked to subduction, plate interactions, mechanical coupling and/or mantle flow that result in large-scale basin abandonment and the creation of stratigraphic hiatuses. These factors could arrest foreland subsidence, prompt erosion across the basin and generate a regional unconformity.

### ***Case studies from the Andean foreland basin***

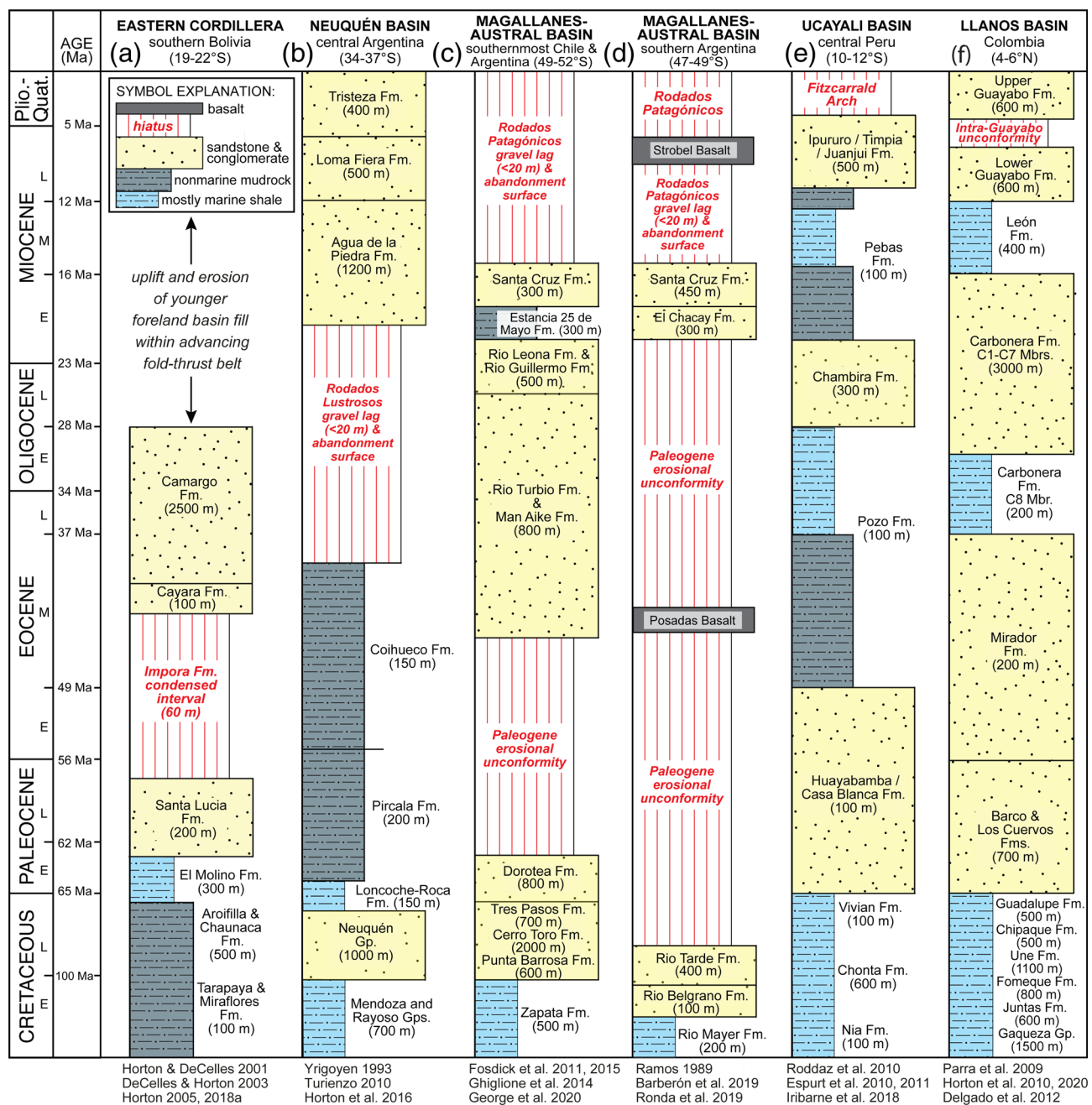
Geodynamic mechanisms of unconformity development are assessed for six examples from the Andean retroarc foreland basin (Fig. 4): three examples involving lower to mid-Cenozoic foreland basin fill, now uplifted in the Andean fold-thrust belt, and three examples from the Pliocene–Quaternary record of the modern foreland basin. These separate cases are unified by stratigraphic records with long-duration hiatuses (Fig. 5) that demonstrate the interruption of otherwise continuous accommodation generation.

The six situations span different time frames, deformation scenarios, climatic settings and geodynamic configurations. These include: (1) regions with the greatest shortening, thickest crust and highest mean elevation (in the central Andes of southern Peru, Bolivia and northern Argentina); (2) zones of low shortening, normal crustal thickness and subdued orogenic topography (the southern Andes, including Patagonia); and (3) areas involving the accretion of oceanic materials during transpressional reactivation of older extensional systems (the northern Andes of northern Peru, Ecuador and Colombia) (Kley *et al.* 1999; Aleman and Ramos 2000; Mora *et al.* 2010; Horton 2018b). The climatic conditions are markedly varied due to the zonal atmospheric circulation and rain shadow effects, which yield high-magnitude precipitation in retroarc sectors of the northern and central Andes, but sharply lower precipitation in the southern Andean foreland (Montgomery *et al.* 2001). Spatially irregular glaciation has disproportionately affected higher elevation Andean districts and large parts of the southern Andes (e.g. Bourgois *et al.* 2000; Ghiglione *et al.* 2019).

Andean plate tectonic configurations vary latitudinally (Fig. 4), as expressed in (1) restricted provinces of flat-slab subduction (e.g. the Colombian/Bucaramanga (2–8° N), Peruvian (5–15° S) and Chilean/Pampean (27–33° S) flat-slab segments), (2) the subduction of aseismic ridges (e.g. the Sandra (5–6° N), Carnegie (0–2° S), Nazca (14–16° S) and Juan Fernández (32–34° S) Ridges); and (3) the subduction of active oceanic spreading ridges and opening of asthenospheric slab windows (e.g. the Patagonian slab window adjacent to the Chile Ridge (45–48° S) along the Nazca–Antarctic plate boundary) (Jordan *et al.* 1983; Ramos 1999, 2005; Gutscher



**Fig. 4.** Simplified tectonic map of western South America and adjacent plates showing the subduction trench, active oceanic spreading ridges, aseismic ridges, zones of flat-slab subduction, Andean magmatic arc, Andean topographic front and modern Andean foreland basin and forebulge axis (modified from Horton 2018a, b; after Ramos and Folguera 2009). Plate velocities are shown relative to the South American plate (DeMets *et al.* 2010). The box outlines indicate the locations of the six case studies of unconformities and condensed stratigraphic intervals within three ancient segments of the Andean foreland basin (north–south strips; Fig. 5a–c) and three modern segments of the basin (squares; Fig. 5d–f), along with representative stratigraphic, topographic and seismic profiles (Figs 6–9).



**Fig. 5.** Chronostratigraphic columns for the six localities within the Andean foreland basin showing Cretaceous–Cenozoic rock units, ages, thicknesses and simplified lithologies. The stratigraphic discontinuities, representing temporal hiatuses (vertical ruled pattern), are expressed in the rock record as unconformities, condensed sections or abandonment surfaces with gravel lags. Site locations are shown in Figure 4.

*et al.* 2000; Lagabriele *et al.* 2004; Lonsdale 2005; Ramos and Folguera 2009; Wagner *et al.* 2017). Despite these spatial variations, modern earthquake focal mechanisms and global stress datasets show that shortening dominates most of the Andean retroarc fold-thrust belt, with low-magnitude extension restricted to elevated plateau regions and strike-slip deformation focused along the northern and southern extremities of the orogenic belt (Assumpção *et al.* 2016; Heidbach *et al.* 2018). Within the modern Andean foreland, a flexural forebulge is roughly parallel to the thrust front, consistent with a regional isostatic (flexural) response to the Andean topographic load (Horton and DeCelles 1997; Chase *et al.* 2009).

Current plate motions (DeMets *et al.* 2010) define nearly orthogonal Nazca–South America convergence along most of the western margin of the South American plate, excluding the northern and southern terminuses where oblique convergence and transform

motion occur at lower rates with the adjacent Caribbean, Antarctic and Scotia plates (Fig. 4). Past studies have recognized temporal variations in the relative convergence rate, absolute overriding plate velocity and convergence direction (e.g. Pilger 1984; Pardo-Casas and Molnar 1987; Somoza and Ghidella 2012). Whereas many studies have linked Andean orogenesis to Nazca–South America convergence and/or the absolute westward motion of South America (e.g. Coney and Evenchick 1994; Silver *et al.* 1998; Mpodozis and Cornejo 2012; Maloney *et al.* 2013; Horton 2018b), others suggest that more-oblique convergence, particularly in the early Cenozoic, may correlate with phases of diminished Andean shortening or tectonic quiescence (e.g. Jaillard and Soler 1996; Aleman and Ramos 2000; Carlotto 2013).

Six case studies from the Andean foreland basin (examples a–f) are presented individually in chronological and geographical order,

with ancient examples (examples a–c) presented from north to south (Figs 4 and 5a–c) and modern examples (examples d–f) presented from south to north (Figs 4 and 5d–f). For each locality, an overview of the chronostratigraphic relationships and description of the unconformities or condensed sections is followed by a discussion of the potential geodynamic mechanisms responsible for accommodation reduction and unconformity development.

### ***Paleogene condensed section with palaeosols, Eastern Cordillera, southern Bolivia and northern Argentina***

A widespread condensed section marked by distinctive palaeosols distinguishes the Paleogene foreland basin succession in the central Andes of Bolivia and northern Argentina (Fig. 4). The condensed stratigraphic zone (Fig. 5a) consists of a <20–100 m thick interval of stacked hypermature palaeosol horizons (or ‘supersols’) within the Impora Formation of Bolivia and the Maiz Gordo and Lumbraera Formations of northern Argentina (Horton and DeCelles 2001; DeCelles and Horton 2003; DeCelles *et al.* 2011; DeCelles 2012; Horton 2018a). These fine-grained pedogenic facies divide, with no angular discordance or erosional relief, an underlying *c.* 100–300 m thick Maastrichtian–Paleocene marine to distal non-marine section from an overlying Oligocene–Miocene, 2–6 km thick, upward-coarsening succession attributed to fluvial deposition in a foredeep setting. In some locations the condensed interval is instead expressed as a single disconformity with no presence of palaeosols. Although direct age control is hindered by thorough pedogenic alteration and a lack of primary volcanic layers, age constraints from the underlying and overlying units broadly limit the condensed zone to an early/middle Eocene to early Oligocene age (DeCelles and Horton 2003; Horton 2005; DeCelles *et al.* 2011; Siks and Horton 2011).

The roughly 15–20 myr condensed section (Fig. 5a) contains extremely mature palaeosols with soil horizonation (zones of leaching and accumulation, mottling, gleying and oxidation) and well-developed calcareous nodules, glaebules, peds, root traces and trace fossils, which overprint and mostly destroy the original sedimentary structures. These prominent pedogenic facies or ‘supersols’ are identified for up to 1000 km along strike, from 17 to 26° S (Fig. 4). The geological relationships require extremely low time-averaged rates of Paleogene sediment accumulation (<5 m/myr) over a large portion of the central Andean foreland basin, without syndepositional shortening, tilting or extensive erosion.

Potential explanations for the Paleogene condensed interval or equivalent disconformity in the central Andes at 17–26° S include: (1) climate change; (2) reduced accommodation during tectonic quiescence or oblique deformation; (3) regional uplift during flat-slab subduction; (4) intraforeland basement uplifts; and, the preferred interpretation, (5) forebulge growth and advance.

### ***Climate change (Fig. 3f)***

Elevated air temperatures and a humid climate during the Paleogene may have boosted weathering intensity and rainfall in the foreland basin (Sempere *et al.* 1997; Starck 2011), enhancing soil genesis and yielding the stacked palaeosols of the condensed section. However, such a change in climate would also have increased the supply of sediment, which is contradicted by a sharp decline in sediment accumulation during Paleogene basin evolution.

### ***Reduced accommodation during tectonic quiescence or oblique deformation (Fig. 3d, e)***

A potential Paleogene shutdown of shortening and/or transition to strike-slip deformation could explain a protracted pause in crustal loading and flexural subsidence. Although oblique Eocene deformation has been identified farther north in the hinterland districts of Peru (Carlotto 2013), a shift to strike-slip deformation is

not supported for the entire central Andean region, where subduction-related arc magmatism and retroarc shortening persisted (Mamani *et al.* 2010; Perez and Horton 2014; Horton *et al.* 2015; Garzione *et al.* 2017). For Bolivia and northern Argentina, thermochronological data indicate middle to late Eocene cooling in the hinterland, which points to erosional exhumation during sustained crustal shortening (Barnes *et al.* 2006; Gillis *et al.* 2006; Carrapa and DeCelles 2008; Rak *et al.* 2017).

### ***Regional uplift during flat-slab subduction (Fig. 3g)***

A roughly Eocene transition to shallow subduction may have promoted dynamic uplift that counteracted the effects of shortening-induced flexural subsidence. Evidence for slab shallowing derives from the time–space record of igneous activity, such that an inboard advance of arc magmatism can be attributed to the migration of the leading hinge of a shallow segment of the subducted slab (e.g. Sandeman *et al.* 1995; James and Sacks 1999; Ramos and Folguera 2011). The principal objection to this option is that dynamic uplift associated with flat-slab subduction would be unlikely to exceed the subsidence expected for thrust-generated topographic loads (see flexural models of Perez and Levine 2020).

### ***Intraforeland basement block uplift (Fig. 3c)***

Proposals for a Paleogene broken foreland basin in northern Argentina are based on local reports of minor dip discordances that are not connected to major fold–thrust structures (e.g. Montero-López *et al.* 2018; Payrola *et al.* 2020). The lack of a substantial up-section decrease in dip, the dominance of relatively fine-grained facies and the regional depositional continuity of Paleogene units (Boll and Hernández 1986; Jordan and Alonso 1987; Siks and Horton 2011; Starck 2011) preclude the large topographic barriers expected for a basin comparable with classic Laramide or Sierras Pampeanas broken foreland basins (e.g. Jordan 1995; Ramos 2009; Lawton 2019). Intraforeland uplifts also would be of insufficient scale to account for the spatial extent of the Paleogene condensed section for up to 1000 km along strike.

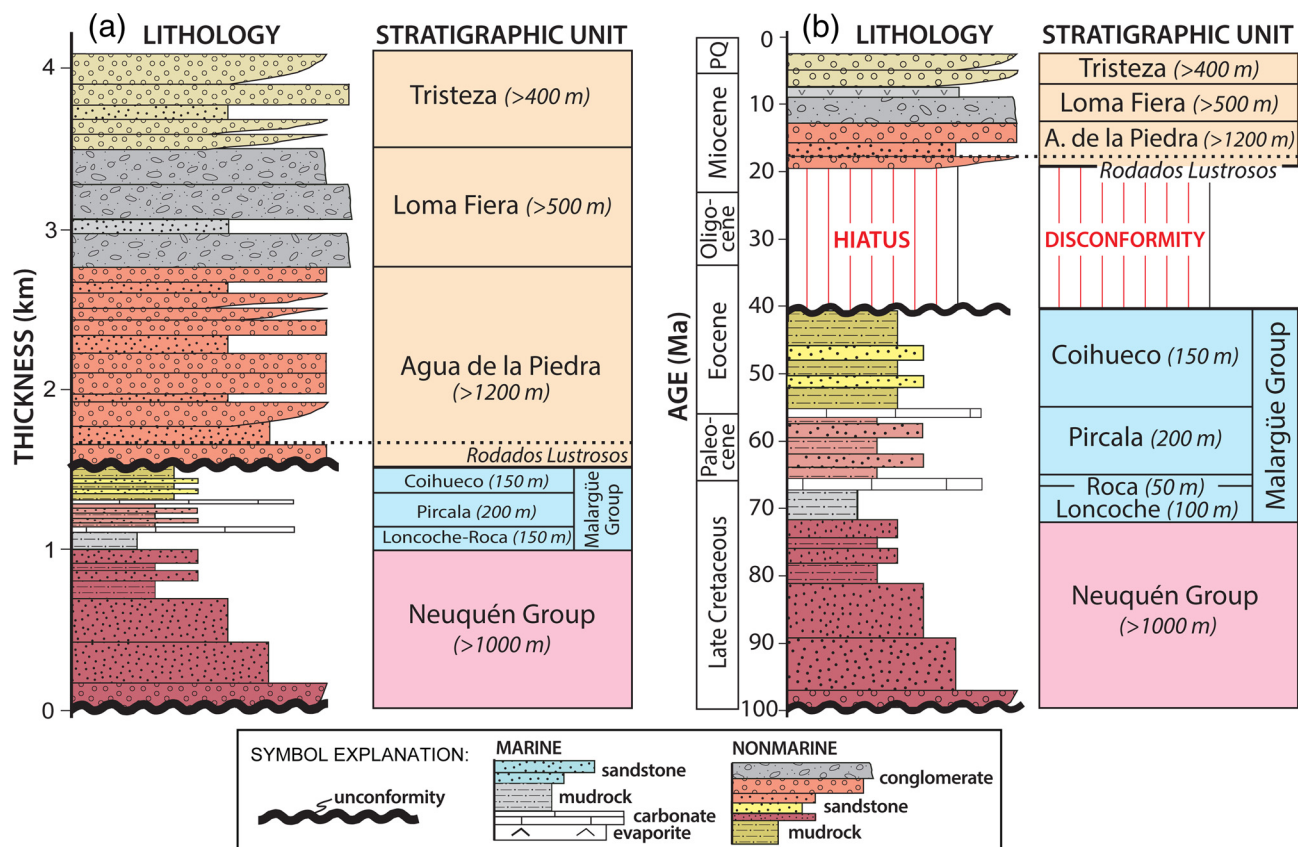
### ***Forebulge growth and advance (Fig. 3b)***

The favoured explanation is that diminished Paleogene sediment accumulation is the result of the growth and cratonward migration of a flexural forebulge. The accumulation of a palaeosol-rich condensed section can be ascribed to low-accommodation and low-erosion conditions across a broad-wavelength forebulge with limited relief. The large magnitude of trench-perpendicular shortening (>200–300 km), commensurate with the advance of fold–thrust deformation, and attendant crustal thickening (DeCelles and Horton 2003; McQuarrie *et al.* 2005; Uba *et al.* 2009) suggest a high lateral mobility of the foreland basin (as outlined in Fig. 2a).

The interpretation of an eastward advancing forebulge is consistent with topographic loading driven by shortening in a fold–thrust belt above a mid-crustal décollement (McQuarrie 2002). Further, the estimated 15–20 myr duration of the condensed section (Fig. 5a) is compatible with the timescales of forebulge migration expected for the reported values of Andean shortening, flexural rigidity and regional sedimentary thicknesses (e.g. Horton and DeCelles 1997; DeCelles and Horton 2003; Echavarría *et al.* 2003; McQuarrie *et al.* 2005; Anderson *et al.* 2018; Calle *et al.* 2018; Rahl *et al.* 2018).

### ***Mid-Cenozoic hiatus, abandonment surface and gravel lag, Neuquén Basin, central Argentina***

A disconformity marked by an abandonment surface and capping gravel lag demarcates a mid-Cenozoic (*c.* 40–20 Ma) hiatus in the



**Fig. 6.** Two stratigraphic columns for the northern Neuquén Basin of west-central Argentina showing (a) thickness and (b) time (age) (from Horton *et al.* 2016; Horton and Fuentes 2016). The columns show the same Upper Cretaceous–Cenozoic stratigraphic framework, including rock units, ages, thicknesses and simplified lithologies (compare with Fig. 5b). The two columns show the value of comparing stratigraphic (thickness) to time (age) depictions in assessing stratigraphic discontinuities (unconformities, condensed sections or abandonment surfaces with gravel lags) and temporal hiatuses. Site locality is shown in Figure 4.

Neuquén Basin of west-central Argentina (Fig. 4). The base of the Agua de la Piedra Formation is composed of a 2–20 m thick ultrastable conglomerate referred to as the ‘Rodados Lustrosos’ (Fig. 5b), which contains distinctive shiny (‘lustrous’) pebble and cobble clasts that are extremely smooth, well-rounded and have been polished by large-distance transport and high degrees of weathering, oxidation, recrystallization and rock varnish development (Horton *et al.* 2016). This diagnostic unit occurs in the middle of an Upper Cretaceous–Cenozoic foreland basin succession up to 4–5 km thick (Fig. 6). In most localities, the conglomeratic facies define a disconformable, non-angular contact with limited incision into underlying the Upper Cretaceous to middle Eocene clastic facies of the Neuquén Group and finer grained Malargüe Group (Fig. 6).

In contrast with the preceding example in the central Andes, there is no evidence of lengthy pedogenesis in the Neuquén Basin; rather, a geomorphic abandonment surface (or multiple closely spaced surfaces) is capped by a thin (<20 m) veneer of gravel. A reduction in long-term sediment accumulation and the paucity of pedogenic facies suggest that intrabasinal sediment transport from the middle Eocene to earliest Miocene resulted in bypass to distal regions to the east. A further indication of sediment bypass – as opposed to sediment starvation – includes local erosion represented by palaeovalleys in which the ‘Rodados Lustrosos’ unit disconformably overlies uppermost Cretaceous strata (Garrido *et al.* 2012). This pervasive conglomeratic unit has been identified over a >400 km distance along strike, from 33 to 37° S (Gorroño *et al.* 1979; Yrigoyen 1993; Sempere *et al.* 1994). An absolute age of *c.* 40–20 Ma is assigned to the hiatus (Figs 5b and 6) on the basis of maximum depositional age constraints provided by detrital zircon U–Pb geochronological results for sandstones with considerable

volcanogenic material (Horton *et al.* 2016; Horton and Fuentes 2016; Horton 2018a; Fuentes and Horton 2020).

Several possible options could explain the *c.* 20 myr mid-Cenozoic depositional hiatus in the Neuquén Basin at 33–37° S, including: (1) intraforeland basement uplifts; (2) forebulge growth and advance; (3) a change in climate or sea-level; and, the favoured interpretation, (4) reduced accommodation during tectonic quiescence.

#### *Intraforeland basement block uplift (Fig. 3c)*

The present foreland region is broken by the San Rafael uplift, a basement-involved contractional structure *c.* 100 km east of the modern Andean topographic front. The timing of this structure is much younger than the 40–20 Ma hiatus, with most rock uplift accomplished in the late Miocene (Ramos and Kay 2006; Ramos and Folguera 2009, 2011). No comparable mid-Cenozoic basement-involved structures of sufficient scale have been recognized at depth within the foreland basin (Boll *et al.* 2014).

#### *Forebulge growth and advance (Fig. 3b)*

Diminished accumulation may be the product of forebulge advance (Giambiagi *et al.* 2001). In this instance, however, the narrow width of the fold–thrust belt (<100 km), the low magnitude of shortening (15–45 km) and the strong influence of pre-existing Mesozoic normal faults (e.g. the Malargüe fold–thrust system; Mancada and Figueroa 1995; Giambiagi *et al.* 2008, 2012; Turienzo 2010; Fuentes *et al.* 2016) are insufficient to drive the required >100 km lateral migration of the foreland subsidence profile (flexural wave).

*Climate change or sea-level fall (Fig. 3f)*

A shift to a humid climate may have promoted heightened erosion and highly efficient sediment transport across the foreland, to such an extreme as to leave no deposits, but only an abandonment surface and gravel lag. This mechanism would involve a major mismatch between the sediment transport capacity and accommodation, and would require large-magnitude exhumation in the thrust belt. To date, the available fission track and U–Th/He thermochronological results show no enhanced exhumation during the 40–20 Ma window (Folguera *et al.* 2015b; Bande *et al.* 2020).

Alternatively, an Oligocene drop in base level may partially account for the negligible accumulation in the Argentina foreland (Fuentes and Horton 2020). Although this lowstand would briefly suppress accommodation at a regional scale, this mechanism alone would be insufficient to account for the long duration and virtually complete elimination of accommodation across the foreland from the middle Eocene to earliest Miocene.

*Reduced accommodation during tectonic quiescence or oblique deformation (Fig. 3d, e)*

The preferred interpretation for the ‘Rodados Lustrosos’ gravel lag and associated 20 myr hiatus is a prolonged cessation of shortening and flexural loading. The elimination of accommodation would have led to abandonment of the foreland basin, with long-distance sediment transport and/or long residence times for recycled coarse material from the inactive thrust belt. Isolated palaeovalley formation suggests local erosion, potential reflecting a drop in base level or some degree of isostatic rebound. Of particular importance is the absence of a structural or thermochronological record of contemporaneous shortening. Instead, the 40–20 Ma time frame coincided with a period of modest hinterland extension of the Andean orogenic belt at these latitudes (Jordan *et al.* 2001b; Burns *et al.* 2006; Charrier *et al.* 2007; Folguera *et al.* 2010; Rojas Vera *et al.* 2014).

In summary, the mid-Cenozoic disconformity in the Neuquén Basin of west-central Argentina (Fig. 5b) is thought to reflect a sustained pause in flexural subsidence that resulted in no net accumulation across an inactive foreland basin, which may be attributable to the cessation of sediment transport or, more likely, to basinwide sediment bypass (Legarreta and Uliana 1991; Horton and Fuentes 2016). In this example, any viable geodynamic mechanisms for foreland quiescence must account for a coeval shift to a neutral or extensional tectonic regime in the corresponding hinterland regions. In the absence of high topography and a thickened crust suitable for gravitational collapse (e.g. Giovanni *et al.* 2010; Horton 2012), a favoured option is a reduction in mechanical coupling stemming from slab steepening, trench rollback or a decrease in the overriding plate velocity (Maloney *et al.* 2013; Horton 2018b).

***Paleogene erosional unconformity, Magallanes–Austral Basin, southernmost Chile and Argentina***

A Paleogene unconformity spanning 15–20 myr is marked by an erosional surface within the Magallanes–Austral foreland basin (Fig. 4). The unconformity (Fig. 5c, d) is situated within the intermediate to upper levels of the 5–7 km thick Upper Cretaceous through Cenozoic foreland succession. The erosional surface incises upper Maastrichtian to lower Paleocene shelf-edge deltaic deposits of the Dorotea and Cerro Dorotea formations and is capped by middle Eocene shallow marine and estuarine deposits of the Man Aike and Rio Turbio formations and the equivalent local units (Fig. 5c, d).

The stratigraphic discontinuity can be traced in surface and subsurface datasets over *c.* 700 km along strike, from 47 to 54° S. Over

much of its extent, the hiatus spans *c.* 15–20 myr, from the early Paleocene through middle Eocene (Fig. 5c). However, the duration varies significantly from north to south, with regional diachroneity defined by a local maximum hiatus of *c.* 40–60 myr in the north (from Lago Pueyrredón to Lago Viedma at 47–50° S; Fig. 5d) and a continuous Paleocene–Eocene section in the south (within Tierra del Fuego at 53–55° S) (Biddle *et al.* 1986; Ramos 1989; Wilson 1991; Malumián 2002; Ghiglione *et al.* 2014; Sickmann *et al.* 2018; Ronda *et al.* 2019; Fosdick *et al.* 2020; George *et al.* 2020).

Although the Paleogene unconformity lacks an extensive angular discordance, Fosdick *et al.* (2011) interpreted ‘a subtle angular unconformity’ in selected seismic data. The unconformity cuts across different horizons in the underlying stratigraphic units, suggesting erosional relief exceeding several tens of metres. Although the eroded thickness of the former overburden has been estimated at *c.* 5 km (Fosdick *et al.* 2015), vitrinite reflectance data indicate limited erosion, with the removal of no more than *c.* 500 m during the formation of the unconformity (George *et al.* 2020).

A range of possible geodynamic mechanisms may explain the diachronous Paleogene unconformity in the Magallanes–Austral Basin at 47–54° S. With no clear preference, the choices include: (1) shortening in the proximal foreland; (2) forebulge growth and advance; (3) isostatic rebound during tectonic quiescence; (4) regional uplift during flat-slab subduction; and (5) uplift associated with slab window formation or slab break-off.

*Shortening in the proximal foreland (Fig. 3a)*

The erosional unconformity may be part of a regional growth structure linked to syndepositional shortening in the frontal zone of the east-directed Patagonian fold–thrust belt (Fosdick *et al.* 2014). Although younger Neogene growth strata are preserved in units above the unconformity (Malumián *et al.* 2000; Ghiglione *et al.* 2016a), the absence of thickness variations or pronounced angular discordance within the Paleogene succession, and the lack of a candidate contractional structure, suggests that localized thrust belt shortening was not responsible for the regional hiatus. The spatial continuity of the preserved erosion surface over hundreds of kilometres would also require a structure of exceptional strike length, which is incompatible with mapped features.

*Forebulge growth and advance (Fig. 3b)*

A cratonward-advancing forebulge linked to the Patagonian fold–thrust belt could produce an erosional unconformity of long duration and large spatial extent (Wilson 1991). In this explanation, the forebulge unconformity would overlie a thin accumulation of distal backbulge deposits (Figs 1a and 2a), which conflicts with the large (>4 km) thickness of the underlying Upper Cretaceous foreland basin succession. In addition, the required magnitude of lateral migration (>100 km) is inconsistent with the reported values of crustal shortening, which are generally <30–40 km (Alvarez-Marrón *et al.* 1993; Fosdick *et al.* 2011).

*Isostatic rebound during tectonic quiescence (Fig. 3d, e)*

A termination of shortening and crustal loading would be sufficient to induce erosional unloading of the foreland basin and generate a regional unconformity (George *et al.* 2020). Testing this hypothesis is inhibited by the absence of direct age control on any candidate structures that may have been active during Late Cretaceous to early Eocene time. Reported shifts in Paleogene exhumation and provenance (e.g. Fosdick *et al.* 2020) permit, but do not require, synchronous deformation. Direct structural timing constraints are necessary to better separate phases of non-deposition from the regional effects of sea-level lowstands (Malumián 2002; Olivero and Malumián 2008).

### Regional uplift during flat-slab subduction (Fig. 3g)

A potential decline or termination of arc magmatism at roughly 65–45 Ma may reflect a phase of Paleogene flat-slab subduction, as proposed farther north for northern Patagonia at 40–46° S (Gianni *et al.* 2018; Horton 2018b; Butler *et al.* 2020). George *et al.* (2020) point to an apparent lull in arc magmatism at 60–45 Ma on the basis of detrital zircon age distributions at 50–51° S, which would be consistent with an extinguished arc during shallow subduction. If sufficient geochronological and geochemical datasets for Patagonia were available, then the history of subduction-related arc magmatism could be discriminated from intraplate magmatism to identify any inboard progression of arc magmatism.

### Uplift associated with slab window formation or slab break-off (Fig. 3h)

Paleogene slab window development has been discussed for a northern region (44–51° S), where diminished arc magmatism coincided with a phase of intraplate magmatism fed by primitive mantle sources (as represented by extrusive units such as the middle Eocene Posadas basalt; Fig. 5d). In this region, collision of the Aluk-Farallon spreading ridge led to progressive Eocene (roughly 57–40 Ma) opening of a slab window (Ramos 2005; Aragón *et al.* 2013; Gianni *et al.* 2018). Although the Eocene time frame partially overlaps with the hiatus in the Magallanes–Austral Basin at 47–54° S, the predicted timing of slab window formation at more southern latitudes would be substantially younger, by up to 10 myr. Therefore, although the hypothesis would be consistent with the possible 60–45 Ma lull in arc magmatism (George *et al.* 2020), this diachroneity of slab window opening and incomplete records of intraplate magmatism present challenges to this interpretation.

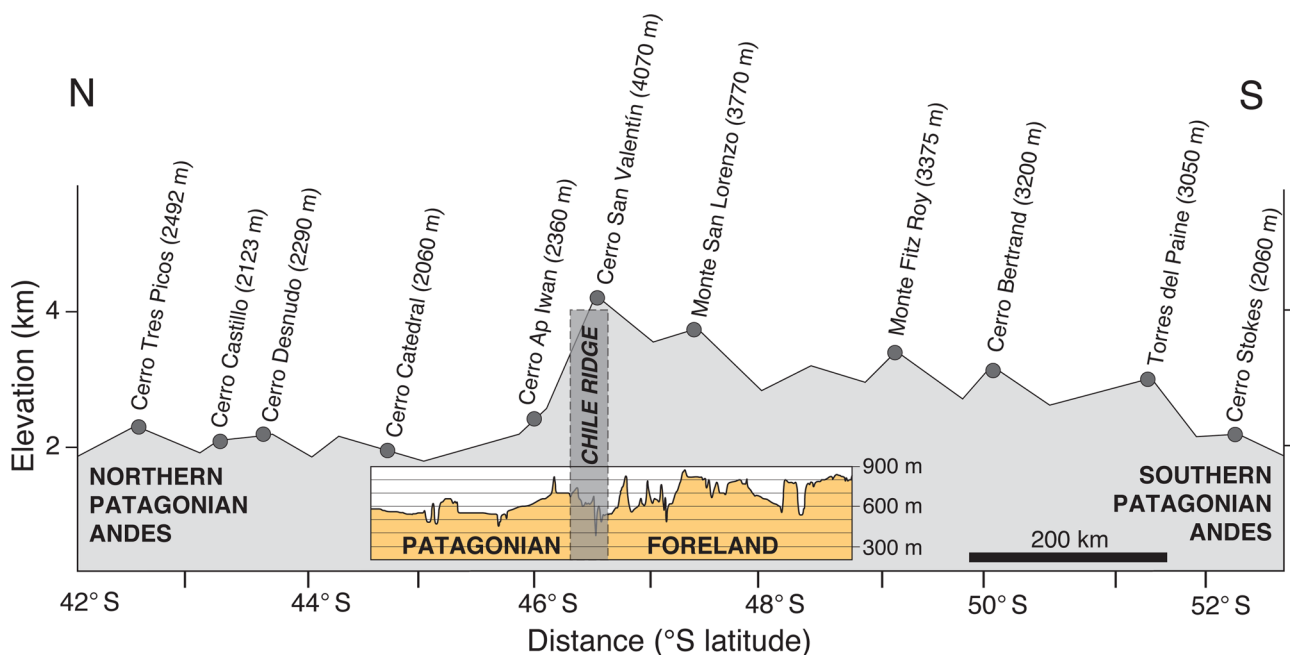
### Late Cenozoic sediment bypass in the Patagonian foreland, southern Argentina

The Patagonian fold–thrust belt at 43–53° S (Fig. 4) is flanked to the east by a low-relief erosional plain that is no longer an actively subsiding foreland basin (Fig. 5d). This retroarc landscape of Argentina includes flat-lying Cretaceous–Cenozoic foreland basin

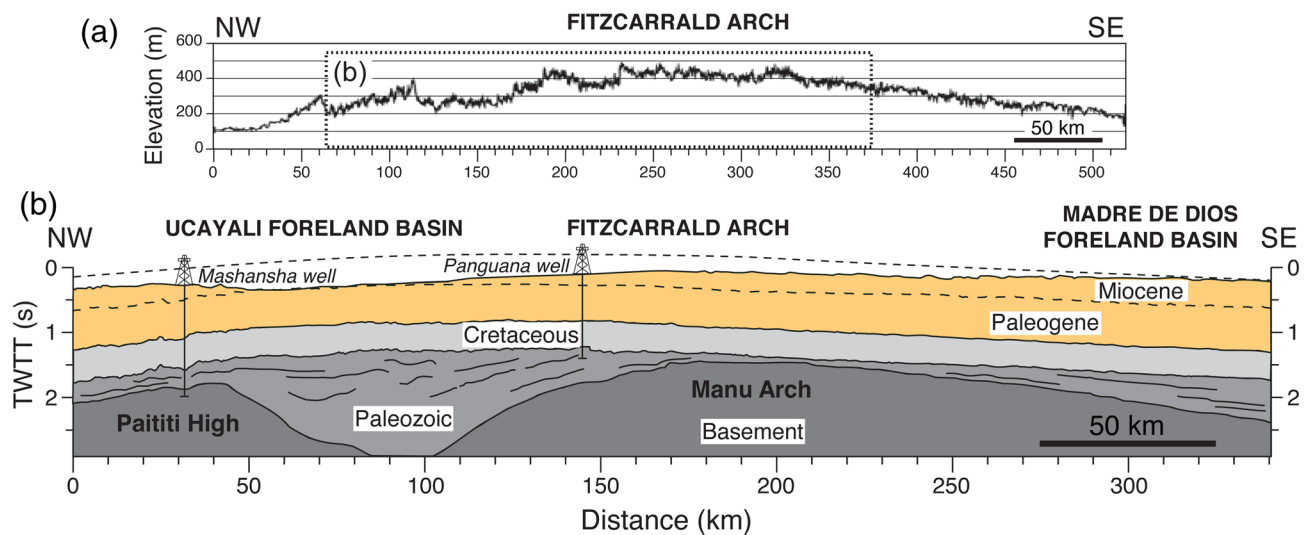
strata that are now undergoing widespread erosion (Ramos 1989, 2005; Bouza and Bilmes 2020). The relict foreland contains an anomalously high topography (roughly 500 m above sea-level), the highest Andean retroarc region (Fig. 4) that is not directly linked to upper crustal faulting. Modest regional uplift across this narrow continental strip is registered by late Miocene–Quaternary fluvial and marine terraces up to several hundred metres above sea-level (Feruglio 1950; Guillaume *et al.* 2009; Padoja *et al.* 2011).

The switch from foreland subsidence to regional erosion occurred in the middle Miocene at *c.* 15 Ma (Fig. 5c, d; Ghiglione *et al.* 2016b; Dávila *et al.* 2019; Fosdick *et al.* 2020). Although the retroarc region ceased to accommodate and preserve sedimentary deposits, this did not prohibit the production and transport of clastic sediments across the abandoned foreland basin. Large-scale exhumation of the fold–thrust belt through fluvial and aeolian processes since *c.* 15 Ma and active glaciation since *c.* 6 Ma have supplied considerable sediment to the eastern plains (Thomson *et al.* 2001; Guillaume *et al.* 2013; Ghiglione *et al.* 2019; Willett *et al.* 2020). This sediment has bypassed the relict foreland to be deposited farther east in the offshore Argentine and Malvinas basins of the Atlantic passive margin (Fig. 4; Ghiglione *et al.* 2016b). The onshore record of sediment bypass is represented by an abandonment surface capped by a thin (<10–20 m thick), but extensive, gravel lag, the ‘Rodados Patagónicos’ (Feruglio 1950; Parras *et al.* 2008; Ghiglione *et al.* 2016b; Barberón *et al.* 2019; Bouza and Bilmes 2020), which Darwin (1842) referred to as the ‘Patagonian Shingle Formation’ (Martínez *et al.* 2009).

Folguera *et al.* (2015a) outlined possible geodynamic controls on the modern foreland configuration; most are encapsulated in the unconformity generation mechanisms summarized previously (Fig. 3). In the Patagonian foreland at 43–53° S, basin abandonment corresponds with the opening of an active slab window (Fig. 3h). Both the Patagonian Andes and adjacent foreland show an abrupt topographic step that overlaps with the position of the Chile triple junction at 46° S (Fig. 7). Ramos (2005) attributed the elevated topography to enhanced dynamic uplift linked to northward progression of the Nazca–Antarctic spreading ridge (the Chile Ridge) to where it now intersects the trench at 46° S (Figs 4 and 7). In this context, a slab window has opened over the past *c.* 16 myr



**Fig. 7.** Topographic profiles of the Patagonian Andes (after Ramos 2005) and corresponding retroarc foreland basin showing the relatively higher mean topography and topographic relief south of the Chile Ridge (the Nazca–Antarctic spreading ridge), which has advanced northward through time. Map-view profile traces are shown in Figure 4; stratigraphic framework is shown in Figure 5c, d.



**Fig. 8.** Profiles of the Fitzcarrald Arch (Figs 4 and 5e) along the boundary between the Ucayali Basin and Madre de Dios Basin, Peru (after [Espurt et al. 2007, 2010](#)). (a) Regional topographic profile showing the c. 500 km wavelength of the uplifted Fitzcarrald Arch. (b) Seismic reflection profile showing the broad warping of the basement and the overlying Phanerozoic strata, as well as the erosional removal of the Neogene strata.

and now underlies the Patagonian foreland from 46 to 55° S ([Lagabrielle et al. 2004](#); [Breitsprecher and Thorkelson 2008](#)). This geodynamic history contrasts with the rest of the Andes. Therefore, rather than focused crustal thickening (e.g. [Stevens Goddard and Fosdick 2019](#)), regional uplift and accommodation reduction along this segment of the Andean margin can be attributed to dynamic mantle processes (Fig. 3h) ([Guillaume et al. 2009](#); [Folguera et al. 2015a](#); [Dávila et al. 2019](#)).

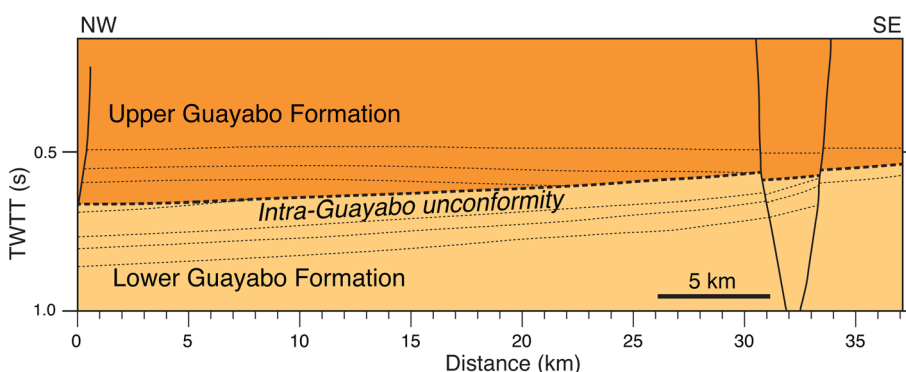
#### *Late Cenozoic flat-slab subduction beneath the Ucayali Basin, central Peru*

The flat-slab province of Peru provides an opportunity to judge the retroarc response to a rapid shift from steep to shallow subduction, a potential mechanism for regional unconformity development in foreland basins (Fig. 3g). The modern zone of flat-slab subduction at 4–15° S is linked to subduction of the buoyant Nazca Ridge (Fig. 4), an aseismic ridge that intersected the margin at c. 10 Ma in northern Peru and swept progressively southward to its current location in southern Peru ([Hampel 2002](#); [Rosenbaum et al. 2005](#); [Ramos and Folguera 2009](#); [Antonić et al. 2015](#)).

The modern foreland basin of Peru consists of actively subsiding, low-elevation (<100–150 m) plains flanked by the Andean fold-thrust belt ([Dumont 1996](#); [Mora et al. 2010](#)). The different geographical designations within the Peruvian (Amazonian) foreland – including the Marañón, Ucayali and Madre de Dios basins – contain comparable stratigraphic records showing large-volume sediment accumulation during the Miocene–Pliocene (Fig. 5e) ([Espurt et al. 2010, 2011](#); [Roddaz et al. 2010](#); [Antoine](#)

[et al. 2016](#); [Iribarne et al. 2018](#); [Zamora and Gil 2018](#)). A critical departure from this trend exists in the Ucayali Basin at roughly 8–12° S, where large-wavelength uplift of Miocene–Pliocene foreland basin fill is expressed in topographic, stratigraphic and structural data ([Espurt et al. 2007, 2010](#)). This anomalous region constitutes the Fitzcarrald Arch, a 400 000 km<sup>2</sup> foreland zone situated at c. 600 m above sea-level, considerably higher than the subsiding plains of the flanking northern (Ucayali) and southern (Madre de Dios) sectors of the modern basin (Fig. 4). Although this elevated region in the proximal foreland could be argued to represent an aggradational fluvial megafan similar to other Andean settings (e.g. [Räsänen et al. 1992](#); [Horton and DeCelles 2001](#)), surface and subsurface geometries showing broad-wavelength (300–600 km) warping of Miocene–Pliocene stratigraphic units since c. 4 Ma (Fig. 8) demonstrate a structural origin for the Fitzcarrald Arch ([Espurt et al. 2007, 2010](#); [Regard et al. 2009](#)). Moreover, the present drainage catchment geometries reveal an increase in relief from fluvial incision ([Regard et al. 2009](#)) rather than the topographic levelling that would be expected for megafan construction during aggradation in a subsiding foreland basin.

The temporal and spatial correspondence of the subducted Nazca Ridge and the Fitzcarrald Arch (Fig. 4) indicate the vital role of flat-slab subduction in driving a reversal from foreland aggradation to degradation. Although uncertainty remains in gauging the exact processes responsible, the increased buoyancy of the aseismic ridge likely caused slab shallowing, which, in turn, led to increased mechanical coupling along the contact between the subducting and overriding plates ([Gutscher et al. 2000](#); [Martinod et al. 2010](#); [Bishop et al. 2018](#)). The foreland response to Nazca Ridge



**Fig. 9.** Line tracing of a seismic reflection profile in the Llanos Basin, Colombia showing an unconformity within the upper Miocene–Pliocene Guayabo Formation (after [Delgado et al. 2012](#)). The intra-Guayabo unconformity indicates erosional truncation of the underlying NW-dipping reflectors (dipping toward the orogen) and progressive onlap by overlying reflectors toward the SE (toward the craton), consistent with rapid regional tilting.

subduction has been regional dynamic uplift (i.e. wholesale uplift in the absence of crustal thickening) that exceeds the coeval flexural subsidence due to shortening in the Andean fold–thrust belt (Fig. 3g).

### ***Late Cenozoic regional tilting in the Llanos Basin, Colombia***

The Pliocene–Quaternary record of the Llanos foreland basin in Colombia (Fig. 4) is linked to the geodynamic evolution of the northern Andes. Using an extensive subsurface dataset, Delgado *et al.* (2012) defined an intrabasinal unconformity that suggests a major shift in foreland accommodation. The unconformity occurs in proximal to medial segments of the Llanos Basin within intermediate levels of the c. 3–4 km thick upper middle Miocene–Pliocene Guayabo Formation (Fig. 5f). This intra-Guayabo unconformity exhibits low-angle discordance over tens of kilometres. Seismic images reveal initial erosional bevelling of older strata that dip gently toward the orogen (northwestward) followed by progressive cratonal (southeastward) onlap of the unconformity surface by younger strata (Fig. 9). This geometry is consistent with sharply asymmetrical subsidence, with rapid regional tilting toward the Andean thrust front accompanied by erosion in more distal parts of the basin, potentially including a broad forebulge or cratonic margin. Multiple small-offset normal faults may have helped facilitate regional tilting (Delgado *et al.* 2012).

Limited age control prevents a clear understanding of the duration of the intraformational hiatus. Regional correlations and past estimates suggest a late Miocene to mid-Pliocene age for the Guayabo Formation (Cooper *et al.* 1995; Mora *et al.* 2008; Parra *et al.* 2009, 2010; Bande *et al.* 2012; Reyes-Harker *et al.* 2015). The precise age of the discordance, although speculative, is inferred to be within the 7.1–4.8 Ma range (Fig. 5f), the reported age of the T-17 palynological biozone (Duarte *et al.* 2017; Jaramillo *et al.* 2017). The actual hiatus may be markedly shorter than this 2.3 myr time frame.

Several possible interpretations warrant consideration. A latest Miocene–early Pliocene age for the intra-Guayabo discordance matches the proposed timing of initial flat-slab subduction beneath the Colombian Andes (Wagner *et al.* 2017). Establishment of this flat-slab configuration at 2–7° N (Fig. 4) may have prompted a pulse of basement-involved shortening within the foreland. However, unlike other proposals of wholesale regional uplift during subduction shallowing, the unconformity geometry would require spatially variable uplift that imparted a regional tilt to the basin prior to erosional bevelling (Fig. 9).

Separately, the apparent time frame for unconformity generation overlapped with reported increases in thrust belt shortening, mean elevation, orographic rainfall and erosional exhumation (Mora *et al.* 2008, 2013). These factors may have led to heightened topographic loading, greater subsidence and, with an increased sediment supply, prospective overfilling of the Llanos foreland basin. Depending on the distribution of tectonic and sedimentary loads, these processes may also have induced a modest retreat of the forebulge or distal basin margin toward the thrust belt, accentuating regional tilting and further promoting unconformity growth in the proximal to medial foreland.

### **Discussion**

The examples presented here provide insights into the geodynamic mechanisms of unconformity development in foreland basins. Consideration of modern and ancient examples helps promote a workflow that evaluates multiple possible explanations (Fig. 3), recognizing the potential for separate mechanisms to operate simultaneously. This paper emphasizes lengthy stratigraphic

discontinuities, which require a severe reduction or elimination of accommodation over geological time frames. This involves (1) erosion, (2) sediment bypass with the transport of materials to more distal regions or (3) the extremely slow accumulation of a condensed stratigraphic interval (Figs 1 and 2). A single locality may fluctuate among these three modes.

This discussion emphasizes criteria pertaining to the interpretation of unconformity generation mechanisms in Andean-type retroarc settings. These mechanisms can be distinguished on the basis of the spatial extent, structural situation, stratigraphic position, palaeoenvironmental conditions and duration of the unconformities and condensed sections within the broader tectonic context of the convergent margin.

### ***Spatial extent and relationship with identified structures***

The geographical distribution, along-strike continuity and geometric relationships with surface or subsurface structures are essential observables. Is the spatial extent of the unconformity compatible with known features, such as a frontal fold–thrust structure (Fig. 3a), a broad flexural forebulge (Fig. 3b) or an intraforeland basement block uplift (Fig. 3c)? The magnitude of any angular discordance, along with correlations of age-equivalent strata and assessments of lateral thickness variations, will aid in the recognition of synorogenic growth strata linked to individual structures. Using this rationale, we can readily dismiss proposed structural controls in circumstances where discernible structures are absent or identified structures are of insufficient strike length, displacement magnitude or time span to account for the location and extent of the unconformity in question.

### ***Age, duration and potential diachroneity of an unconformity***

Information on the age and duration of a stratigraphic discontinuity will help to distinguish short-term eustatic, climatic or autogenic fluctuations (Fig. 3f) from longer term forcings. Several mechanisms are time-transgressive, such as the cratonward migration of a flexural forebulge during propagation of an orogenic wedge (Fig. 3b), or the progressive inboard advance of a flat slab during subduction shallowing (Fig. 3g). The unconformity or condensed section created by such diachronous processes must be compatible with the reconstructed topographic loads, mechanical properties of the foreland lithosphere, and records of deformation and magmatism. Further, the duration of the hiatus may be highly variable, as in episodes of isostatic rebound where greater erosional removal in the proximal foreland yields a stratigraphic gap of diminishing age toward the distal foreland (Fig. 3d, e).

### ***Stratigraphic and palaeoenvironmental history before and after unconformity development***

Determining whether an unconformity or condensed interval is positioned at the base or internally within the foreland basin succession is necessary for proper interpretation (Fig. 2). This distinction resolves whether a stratigraphic discontinuity may mark the inception of orogenesis or some later synorogenic process. In assessing unconformity genesis, the preserved rock records directly above and below the unconformity (or the facies within a condensed section) provide an unambiguous record of accommodation variations immediately before, after and potentially during the hiatus. The relative magnitude of accommodation interruption will scale differently for the potential drivers. Changes in sea-level, climate or erosion (Fig. 3f) will also produce recognizable shifts in palaeoenvironments and sediment transport parameters reflected in the preserved facies.

### Tectonic regimes within the flanking orogenic belt

Plate reconstructions and deformational histories inform geological analyses of foreland basin evolution. One challenge for contractional systems involves periods in which shortening may cease or be greatly lessened. During a shutdown in shortening, the tectonic regime of the orogen may shift to neutral conditions, a tensile mode with hinterland extension (Fig. 3d) or margin-wide strike-slip deformation (Fig. 3e), any of which would induce basin abandonment. The time–space relationships between unconformity development and magmatism are further valued because a pause in subduction-related arc magmatism can reflect (1) a tectonic shift to an oblique, translational or transpressional orogenic system (Fig. 3e), (2) flat-slab subduction (Fig. 3g) or (3) slab window formation (Fig. 3h).

### Time–stratigraphic relationships

In evaluating different options, chronostratigraphic cross-sections (Wheeler diagrams) prove effective in documenting lateral and vertical relationships within the stratigraphic record (Fig. 2). These constraints stimulate estimates of the potential relationships to crustal structures or broader geodynamic elements of the orogenic system. Even where the underlying mechanisms of unconformity generation are not apparent, such time–space assessments are useful in discriminating among viable explanations. In practice, this regularly helps catalyse a discussion of alternative interpretations. For example, the identification of contrasting stratigraphic intervals of comparable age permits the isolation of key variables to better understand why an unconformity or condensed section formed in one region, but not in another.

### Modern analogues

Interpretations of foreland basin unconformities and condensed sections can be challenging, as demonstrated by ancient examples from the Andes (Figs 4 and 5). For the cases described, there are multiple possibilities (Fig. 3) and debate commonly focuses on specific field relationships, stratigraphic correlations or age control within the basin (Fig. 5). Some explanations are mutually exclusive, but others are not.

Case studies from the modern Andean foreland provide greater clarity regarding what may otherwise involve complex or non-unique explanations. These observations address the current plate configuration (Fig. 4) and active foreland basin processes in terms of erosion, sediment bypass or subsidence, enabling interpretations with higher confidence. For the late Cenozoic situations considered, the effects of subduction-related parameters are clear, such that the foreland responses can be observed directly in instances of flat-slab subduction, ocean ridge subduction and slab window generation (Figs 3g, h, 7 and 8). Enhanced exploration of the late Cenozoic to present day foreland conditions will help to improve the ability to accurately reconstruct the geodynamic evolution of ancient convergent margins from the foreland basin stratigraphic record.

**Acknowledgements** Reviewers German Bayona and Matías Ghiglione provided constructive comments that improved this paper. Discussions with Ryan Anderson, Amanda Calle, Tomas Capaldi, Julia Clarke, Kurt Constenius, Andrés Folguera, Facundo Fuentes, Sarah George, Lily Jackson, Chelsea Mackaman-Lofland, Andrés Mora, Mauricio Parra, Nicholas Perez, Sebastian Ramirez, Zachary Sickmann and Daniel Starck helped clarify the presentation.

**Author contributions** BKH: conceptualization (lead), funding acquisition (lead), investigation (lead), writing – original draft (lead), writing – review & editing (lead)

**Funding** This work was funded by the National Science Foundation (EAR-1918541, EAR-1925898 and EAR-1946700).

**Data availability** Data sharing is not applicable to this article as no dataset was generated or analysed during the current study.

*Scientific editing by Feng Cheng*

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