# Chapter 20

# Foreland basin systems revisited: variations in response to tectonic settings

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**ABSTRACT**

The four-part districting scheme (wedge-top, foredeep, forebulge, and backbulge depozones) applies to many foreland basin systems worldwide, but significant variations occur in the stratigraphic record. These variations depend on tectonic setting and the nature of the associated fold-thrust belt. Continued growth of the fold-thrust belt by horizontal shortening requires foreland lithosphere to migrate toward the fold-thrust belt. The flexural wave set up by the topographic load may migrate ~1000 km sideways through the foreland lithosphere, a distance that is comparable to the flexural wavelength. This extreme lateral mobility results in the vertical stacking of foreland basin depozones in the stratigraphic record. The standard stratigraphic succession consists of a several km-thick upward coarsening sequence, marked in its lower part by a zone of intense stratigraphic condensation or a major disconformity (owing to passage of the forebulge), and in its upper part by coarse-grained proximal facies with growth structures (the wedge-top depozone). Foredeep deposits always reside between the forebulge disconformity/condensation zone and wedge-top deposits, and backbulge deposits may be present in the lowermost part of the succession. Wedge-top deposits are vulnerable to erosion because of their high structural elevation, and preservation of backbulge and forebulge deposits depends in part on tectonic setting.

Three main types of fold-thrust belt are recognized: retroarc, collisional (or peripheral), and those associated with retreating collisional subduction zones. Retroarc foreland basin systems (such as the modern Andean) are susceptible to far-field dynamic loading transmitted to the foreland lithosphere by viscous coupling between the subducting oceanic slab and the mantle wedge. This long-wavelength subsidence adds to subsidence caused by the topographic flexural wave, allowing for preservation of well-developed forebulge and backbulge depozones. The absence of dynamic subsidence in collisional (peripheral) foreland basin systems (such as the modern Himalayan) renders forebulge and backbulge regions vulnerable to erosion and non-preservation. Retreating collisional foreland basin systems (such as those in the Mediterranean region) are often associated with large subducted slab loads, which produce narrow but very thick accumulations in the foredeep and wedge-top depozones. These foreland basin systems are characterized by very thick foredeep and wedge-top deposits, well beyond what would be expected from topographic loading alone. Changing lithospheric stiffness in collisional settings may affect preservation of the backbulge and forebulge depozones. If these distal foreland basin deposits are not preserved, roughly half the history of the orogenic event (as archived in the stratigraphic record) may be lost.

Many foreland stratigraphic successions provide sufficient information to estimate the velocity of migration of the flexural wave through the foreland, which may in turn be decomposed into propagation and shortening velocities in the thrust belt. Foreland basin subsidence curves may be inverted to produce an idealized flexural profile, from

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which flexural properties of the lithosphere may be derived. However, spatial changes in flexural rigidity, as well as changes in the size of the orogenic load and rates of propagation and shortening in the thrust belt require that the thrust belt-foreland basin system be palinspastically restored in order to understand the long-term geodynamics.

**Keywords:** foreland basins; flexure; orogenic belts; geodynamics; stratigraphy

## INTRODUCTION

Foreland basin systems consist of depositional regions on continental crust in front of major fold-thrust belts. They are among the largest accumulations of sediment on Earth, spanning entire continental landmasses and spilling into adjacent remnant ocean basins and continental shelves on passive margins. Foreland basin systems are of geodynamic interest because of their genetic association with thrust belts and convergent margins (Price, 1973; Dickinson, 1974). An ideal foreland basin system includes four discrete depozones that form under different local kinematic and subsidence conditions (Fig. 20.1). The *wedge-top* depozone includes sediment that buries the active frontal part of the thrust belt. Although textural and compositional immaturity typifies wedge-top deposits, growth structures are their defining feature. The *foredeep* depozone consists of sediment deposited within the flexural "moat" (Price, 1973) formed by the load of the thrust belt. Sediment derived from the thrust belt may prograde beyond the foredeep, into the region of flexural uplift represented by the *forebulge*, or even beyond it into a broad region of shallow secondary flexural subsidence referred to as the *backbulge* depozone (Fig. 20.1). Although the four-part subdivision of foreland basin systems is borne out in many modern examples worldwide (Zagros, Himalayan, Taranaki, Andean, Apennine, Taiwan, north Australian), important distinctions among different foreland tectonic settings have been revealed, and these distinctions leave telling evidence in foreland stratigraphic records (Sinclair, 1997; Catuneanu, 2004). A single general model for all foreland basins is not appropriate. This chapter provides an updated review of the larger scale aspects of foreland basin systems, with emphasis on differences that develop in response to tectonic processes peculiar to specific tectonic settings, and on the geodynamic coupling between thrust belts and adjacent foreland basin systems.

## TECTONIC SETTINGS OF FORELAND BASINS

Foreland basin systems are intimately linked to thrust belts, of which three main types are recognized: retroarc, collisional (or peripheral), and those associated with retreating collisional subduction zones (Fig. 20.2). Local zones of crustal shortening associated with transform fault systems are not considered. Retroarc and collisional thrust belts form in truly convergent plate tectonic settings, where fixed reference points in each plate are converging relatively rapidly (e.g., Heuret and Lallemand, 2005; Schellart, 2008). Retreating collisional thrust belts form in situations where subduction rate exceeds convergence rate and, consequently, the upper plate is thrown into extension or transtension in regions behind the thrust belt (Malinverno and Ryan, 1986; Doglioni, 1991; Royden, 1993; Jolivet and Faccenna, 2000). Although foreland basins formed in each of these tectonic settings ultimately owe their existence to flexural subsidence in response to loading by orogenically thickened crust, significant additional loads exist in each setting (Royden, 1993; DeCelles and Giles, 1996), and the synorogenic sediment derived from each type of orogenic belt is compositionally and thermally distinctive, recording different levels of exhumation and different types of source rocks (Garzanti et al., 2007).

## LITHOSPHERIC FLEXURE IN FORELAND BASINS

All foreland basin systems subside in response to the development of the adjacent thrust belt load (Beaumont, 1981; Jordan, 1981; Karner and Watts, 1983). Numerous studies have documented the genetic linkages between thrust belt kinematics, orogenic loading, and flexural subsidence in foreland basins. Flexural loading of an ideal elastic plate results in a rapidly damped sinusoidal profile

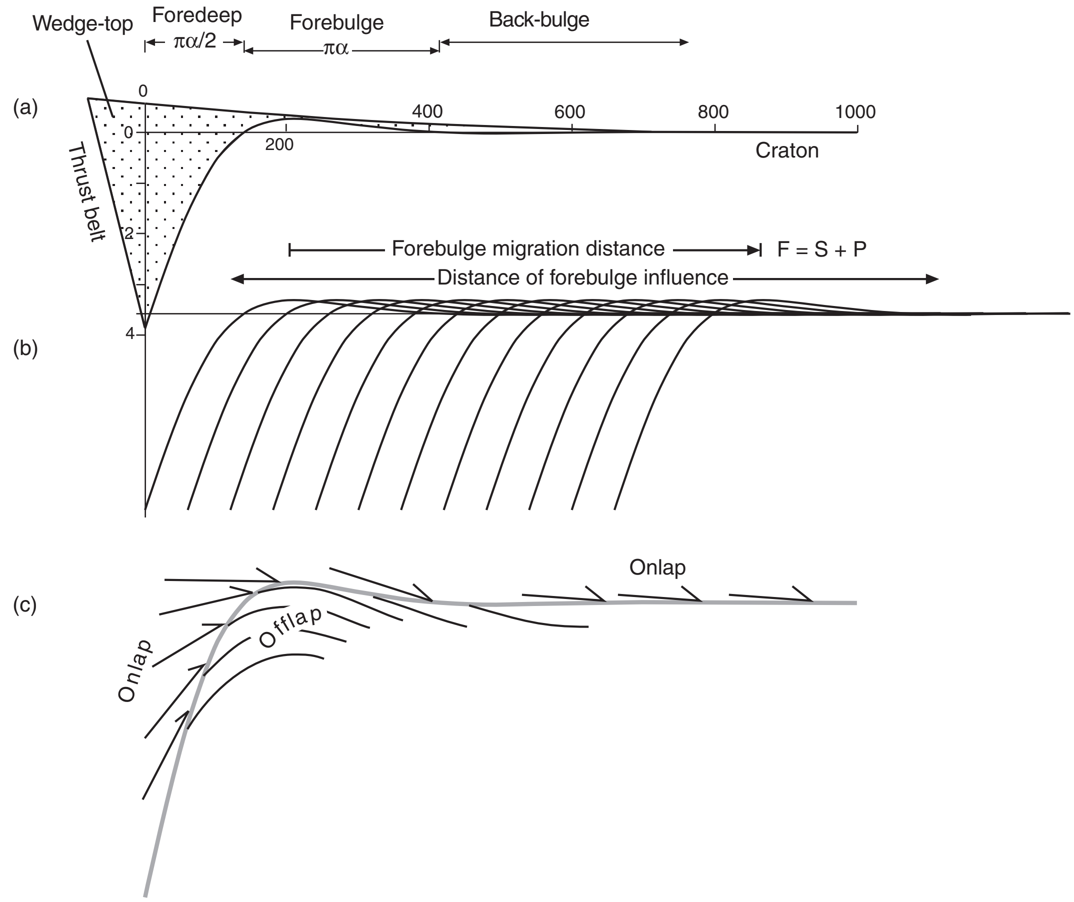



**Fig. 20.1.** (a) Schematic diagram illustrating an ideal foreland basin system in cross section oriented perpendicular to trend of the adjacent thrust belt, inspired by Flemings and Jordan (1989), after DeCelles and Giles (1996). Note the extreme vertical exaggeration. (b) Velocity and distance of influence of the forebulge in a system that migrates 700 km laterally. Flexural wave migration distance ($F$) is the sum of shortening in the thrust belt ($S$) plus the propagation distance of the thrust belt ($P$). (c) Characteristic stratal onlap and offlap patterns that form in association with lateral forebulge migration.

with a large magnitude negative flexure adjacent to the load (the foredeep depozone), a medial positive flexural bulge (the forebulge), and a secondary negative depression in the most distal region (the backbulge depozone) (Fig. 20.1). The amplitude of deflection (negative or positive) decreases by roughly three orders of magnitude from the foredeep to the backbulge depozone. Typical flexural loading of continental lithosphere produces a foredeep depression that scales horizontally with $\pi\alpha$, where $\alpha$ is defined as the flexural parameter (Turcotte and Schubert, 2006):

$$\alpha = \left[\frac{4D}{\Delta\rho g}\right]^{1/4}, \tag{20.1}$$

where $D$ is the flexural rigidity of the lithosphere, $g$ is the acceleration of gravity, and $\Delta\rho$ is the difference in density between the mantle and the basin fill. The wavelength of the flexural profile also depends on whether the load is supported by an effectively continuous plate or a broken plate: continuous plates provide more support for the load than broken plates, spreading the flexural depression over a broader region and resulting in a shallower, longer wavelength depression (Flemings and Jordan, 1989; Turcotte and Schubert, 2006). Debate continues about whether or not the foreland lithosphere behaves purely elastically or experiences visco-elastic relaxation with or without depth-dependent rheological changes (e.g., Quinlan and Beaumont, 1984; Garcia-Castellanos et al., 1997).

Flexural foredeeps produced by line loads on continuous lithosphere are $0.75\pi\alpha$ wide, whereas foredeeps on broken plates are only $0.5\pi\alpha$ wide. Typical values of $D$ for continental lithosphere

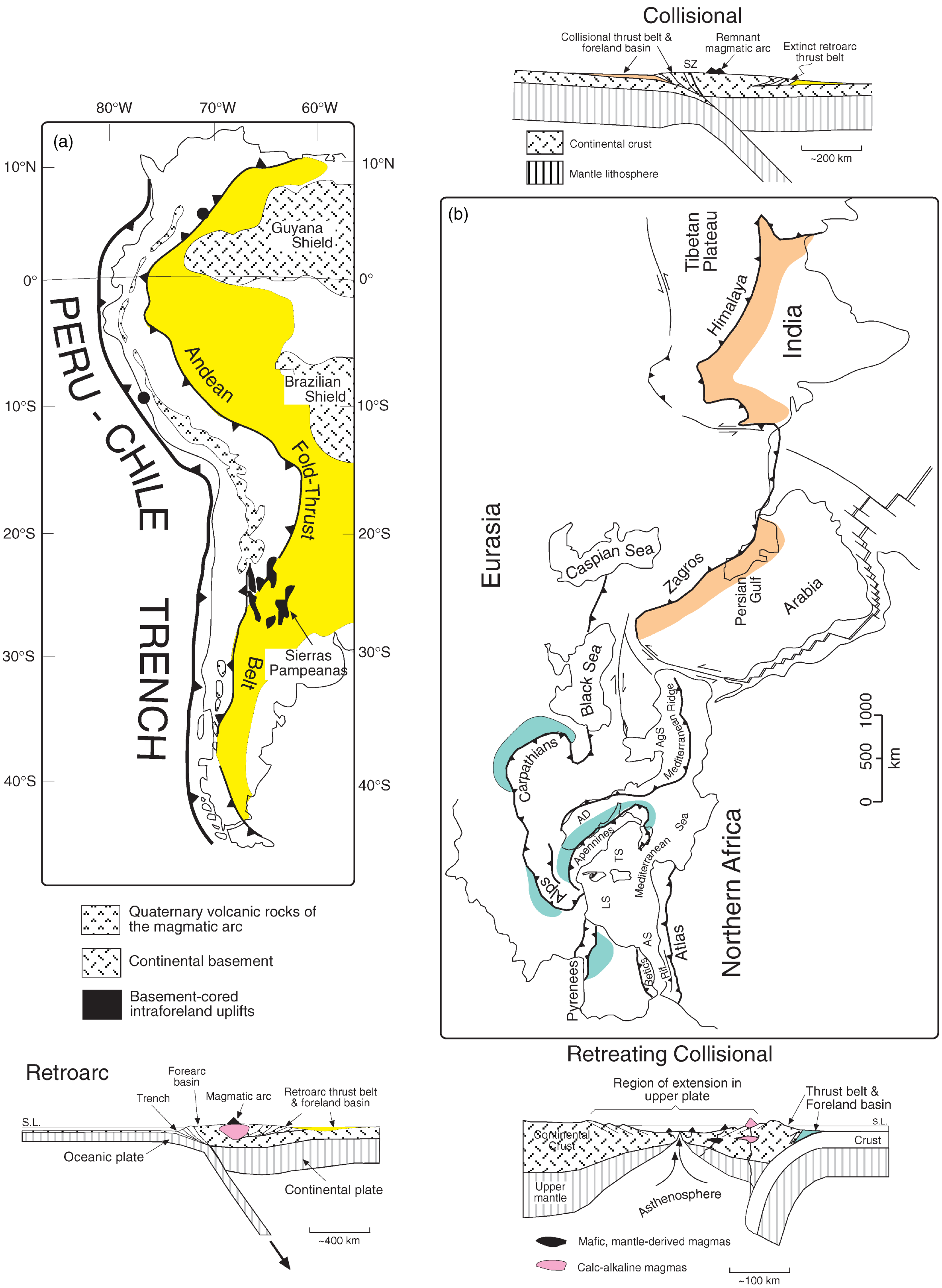



**Fig. 20.2.** Tectonic settings of the three major types of foreland basin systems. (a) The Andean retroarc thrust belt and foreland basin system. (b) Foreland basin systems in collisional (orange) and retreating collisional (blue) settings, with schematic cross-sections above and below the map. Typical collisional settings are the Zagros and Himalaya, and retreating collisional settings are exemplified in the Mediterranean region.

range between $\sim 5 \times 10^{22}$ Nm and $4 \times 10^{24}$ Nm (corresponding to elastic thicknesses of 20–90 km; e.g., Jordan, 1981; Lyon-Caen and Molnar, 1985; Watts, 2001; Roddaz et al., 2005). Thus, for typical basin fill density of 2500 kg/m$^3$, the foredeep is $\sim$110–350 km wide for a broken plate, and $\sim$170–515 km wide for a continuous plate. The forebulge is $\pi\alpha$ wide for both broken and continuous plates, which corresponds to a range of forebulge widths of $\sim$220–690 km for typical continental lithosphere. The amplitude (positive) of the forebulge is $\sim$4–7% of the amplitude of the maximum foredeep (negative) deflection. This translates into heights of $\sim$200–400 m for typical continental lithosphere.

In reality, the size and shape of the load, the amount of sediment in the foreland basin system, the types of sediment transport processes within the basin (Garcia-Castellanos, 2002), normal faulting along the upward flexing portion of the plate (Bradley and Kidd, 1991; Londoño and Lorenzo, 2004), structural inhomogeneities in the continental lithosphere (Flemings and Jordan, 1989; Sinclair et al., 1991; Waschbusch and Royden, 1992; Blisniuk et al., 1998; Cardozo and Jordan, 2001; Cloetingh et al., 2004), and the large-scale three-dimensional shape of the load (Chase et al., 2009) strongly affect the flexural profile. Particularly problematic are ancestral basement structures that are reactivated as the flexural wave migrates through the cratonic lithosphere (Blisniuk et al., 1998; Cardozo and Jordan, 2001), and paleotopographic features that become involved in the flexural profile (Gupta and Allen, 2000; Bilham et al., 2003). Notwithstanding such complications, a rich literature developed over the last 30 years demonstrates that the first-order features of most foreland basin systems are well explained by elastic flexure (for a review, see Allen and Allen, 2005).

## LATERAL MOBILITY OF THE THRUST BELT-FORELAND SYSTEM

Earth's major fold-thrust belts involve hundreds of kilometers of horizontal shortening that accrues over tens of millions of years. For example, the Himalayan thrust belt accommodates at least 500 km and perhaps as much as 900 km of shortening (Coward and Butler, 1984; Srivastava and Mitra, 1994; DeCelles et al., 2001; Robinson et al., 2006; Murphy, 2007); the central Andean thrust belt has been shortened by as much as 400 km (Kley and Monaldi, 1998; McQuarrie, 2002; Arriagada et al., 2008); and the North American Cordilleran thrust belt involves at least 350 km of shortening (DeCelles and Coogan, 2006; Evenchick et al., 2007). Shortening of this magnitude requires, at a minimum, a roughly equal distance of horizontal migration of the flexural wave through the lithosphere. Added to this distance is the lateral displacement of the flexural wave that must take place in order to accommodate the increasing breadth (or propagation distance) of the orogenic wedge. Thus, to first order, the total distance of lateral migration of the flexural wave must approximately equal the sum of the thrust belt propagation distance and the total shortening (Fig. 20.1; DeCelles and DeCelles, 2001). For Earth's continental scale thrust belt-foreland systems, this distance ranges between $\sim$500 km and 1000 km. As discussed in the previous section, the upper limit of this distance scales with typical wavelengths of flexural profiles in foreland basin systems. Therefore, it is likely that over the lifetime of a major thrust belt, the associated foreland basin system will migrate a horizontal distance equal to more than half its own wavelength. In turn, this will result in vertical stacking of depozones in the stratigraphic record (Flemings and Jordan, 1989; Coakley and Watts, 1991; Sinclair et al., 1991; Vergés et al., 1998; Burkhard and Sommaruga, 1998), equivalent to a Walther's law of foreland basin stratigraphy.

"Waltherian" successions of depozones have been documented in a number of foreland basins (e.g., Fig. 20.3), including the North American Cordilleran (Plint et al., 1993; DeCelles and Currie, 1996; Fuentes et al., 2009), Andean (Jordan et al., 1993; DeCelles and Horton, 2003; Uba et al., 2006), North Alpine (Sinclair, 1997; Burkhard and Sommaruga, 1998; Gupta and Allen, 2000), Pyrenean (Vergés et al., 1998), Taiwan (Yu and Chou, 2001; Tensi et al., 2006), Karoo (Catuneanu, 2004), Gangdese (Leier et al., 2007), and Himalayan (DeCelles et al., 1998a, 1998b) systems. The typical stratigraphic pattern is an overall upward coarsening, several km-thick succession punctuated in its lower part by a major erosional disconformity or zone of stratigraphic condensation attributed to passage of the forebulge through the foreland region. Condensation and erosion result from the absence of sediment accommodation in the forebulge region. Nonmarine forebulges are marked by zones of intense pedogenesis, karst weathering (if carbonate substrates

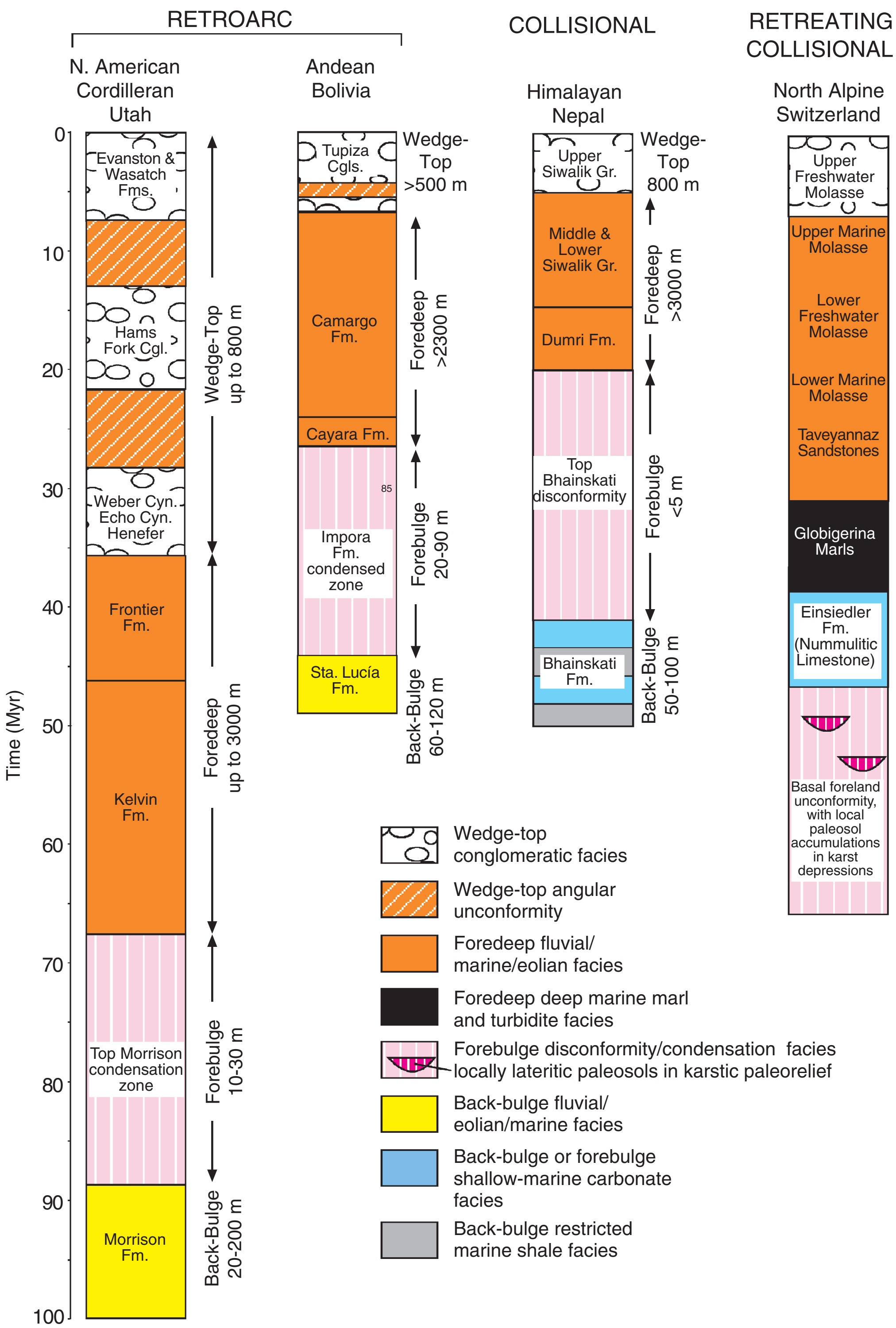



**Fig. 20.3.** Examples of "Waltherian" sequences that result from vertical stacking of foreland basin depozones as the flexural wave migrates through the foreland lithosphere. Note that these are generalized vertical stratigraphic successions for individual localities or local regions in each case; significant time transgressive behavior is present in all examples. Thicknesses represent regional averages for depozones. The vertical axis is time, not age. The Nepal, North Alpine, and Bolivian examples are Cenozoic in age, whereas the Utah example is Late Jurassic-Paleocene in age. *Source*: Nepal, DeCelles et al. (1998a); Bolivia, DeCelles and Horton (2003); Utah, DeCelles (1994) and DeCelles and Currie (1996); and North Alpine, Sinclair (1997), Beck et al. (1998), and Burkhard and Sommaruga (1998).

are present), fluvial erosion, and ultra-stable conglomerate lags (Herb, 1988; Plint et al., 1993; Demko et al., 2004; del Papa et al., 2010). As the flexural forebulge migrates through a region, older faults may be reactivated and local topography may become sufficient to generate coarse-grained sediment in regions far from the orogenic front (Blisniuk et al., 1998; Burkhard and Sommaruga, 1998). Submarine forebulges may be signaled by shallow water carbonate build-ups, ironstones, and evidence for submarine erosion and sediment starvation (Tankard, 1986; Dorobek, 1995; Gupta and Allen, 2000; Allen et al., 2001). Given typical rates of flexural wave migration (10–25 mm/yr) and forebulge widths (~200–700 km), the amount of time represented by the forebulge disconformity/condensation zone (DCZ) is several to several tens of Myr (Fig. 20.3). The age of the forebulge DCZ decreases cratonward, as the thrust belt migrates toward the craton interior (Coakley and Watts, 1991; Crampton and Allen, 1995; Sinclair, 1997; Burkhard and Sommaruga, 1998). In many foreland basin systems, the forebulge DCZ is underlain by distal, fine-grained backbulge deposits, and the DCZ is ubiquitously overlain by the foredeep depozone. The succession is capped by wedge-top sediments containing growth structures and typically coarse-grained, proximal facies (Fig. 20.3; DeCelles, 1994; Ford et al., 1997; Williams et al., 1998; Lawton et al., 1999; Chiang et al., 2004). Wedge-top deposits are characterized by great lateral heterogeneity, numerous local unconformities, and paleogeographic complexity owing to the presence of nearby growing topographic features associated with the encroaching front of the fold-thrust belt.

Although this model adequately predicts many features that are found in the global stratigraphic record of foreland basin systems, several aspects of the general model are expected to be variably expressed depending on tectonic setting. For example, in the absence of extreme sediment flux from the thrust belt and/or regional dynamic subsidence (Gurnis, 1992; Liu et al., 2005), the forebulge and backbulge depozones are not likely to be preserved in the stratigraphic record (Flemings and Jordan, 1989; Sinclair, 1997; Catuneanu, 2004).

## RETROARC FORELAND BASIN SYSTEMS

Retroarc foreland basin systems form along the inboard flanks of Cordilleran (Andean-style) orogenic belts in the upper, continental plate associated with a rapidly converging oceanic-continental subduction zone (Fig. 20.2a; Dickinson, 1974; Jordan, 1995). However, not all convergent oceanic-continental plate boundaries are characterized by Cordilleran orogenic belts and foreland basins; some are dominated by backarc extension in the upper plate. Recent reviews of relationships among the numerous parameters that control shortening in continental plates above subduction zones indicate that rapid trenchward movement of the upper plate and large lateral distance from the edges of the subducting slab are important for developing a strongly contractional retroarc domain (Heuret and Lallemand, 2005; Schellart, 2008). The resulting foreland basins are continental in scale, stretching many thousands of kilometers parallel to and hundreds of kilometers perpendicular to their adjacent orogenic belts. The modern archetype is the Andean foreland basin system, which extends more than 7,000 km along the eastern flank of the Andean thrust belt and separates it from the South American craton to the east (Fig. 20.2a; Jordan, 1995).

The modern Andean foreland basin system has a 50–75 km wide wedge-top depozone, a 250–300 km wide foredeep depozone, a mostly buried forebulge with almost no topographic expression but a 5–10 m positive geoid anomaly indicating upward flexure of the lithosphere (Chase et al., 2009), and a >400 km wide backbulge depozone that is dominated by swampy floodplain and fluvial depositional environments (Fig. 20.4; Horton and DeCelles, 1997). The Andean forebulge is also suggested by arched fluvial terraces in the Beni River basin (Aalto et al., 2003) and detailed sedimentological observations and flexural modeling in the western Amazon drainage basin (Roddaz et al., 2005). A similar foreland basin system developed to the east of the North American Cordilleran thrust belt between Late Jurassic and early Cenozoic time (DeCelles, 2004; Miall et al., 2009). The North American system became dormant as plate kinematics along the western margin of the continent progressively became dominated by transform faults that developed in response to collision with Pacific oceanic spreading centers (Dickinson, 2004). Together, the Andean and Cordilleran foreland basin systems contain $>15 \times 10^6$ $km^3$ of predominantly clastic sediment, as well as significant hydrocarbon accumulations.

In retroarc settings, the foreland basin lithosphere may be affected by an additional load

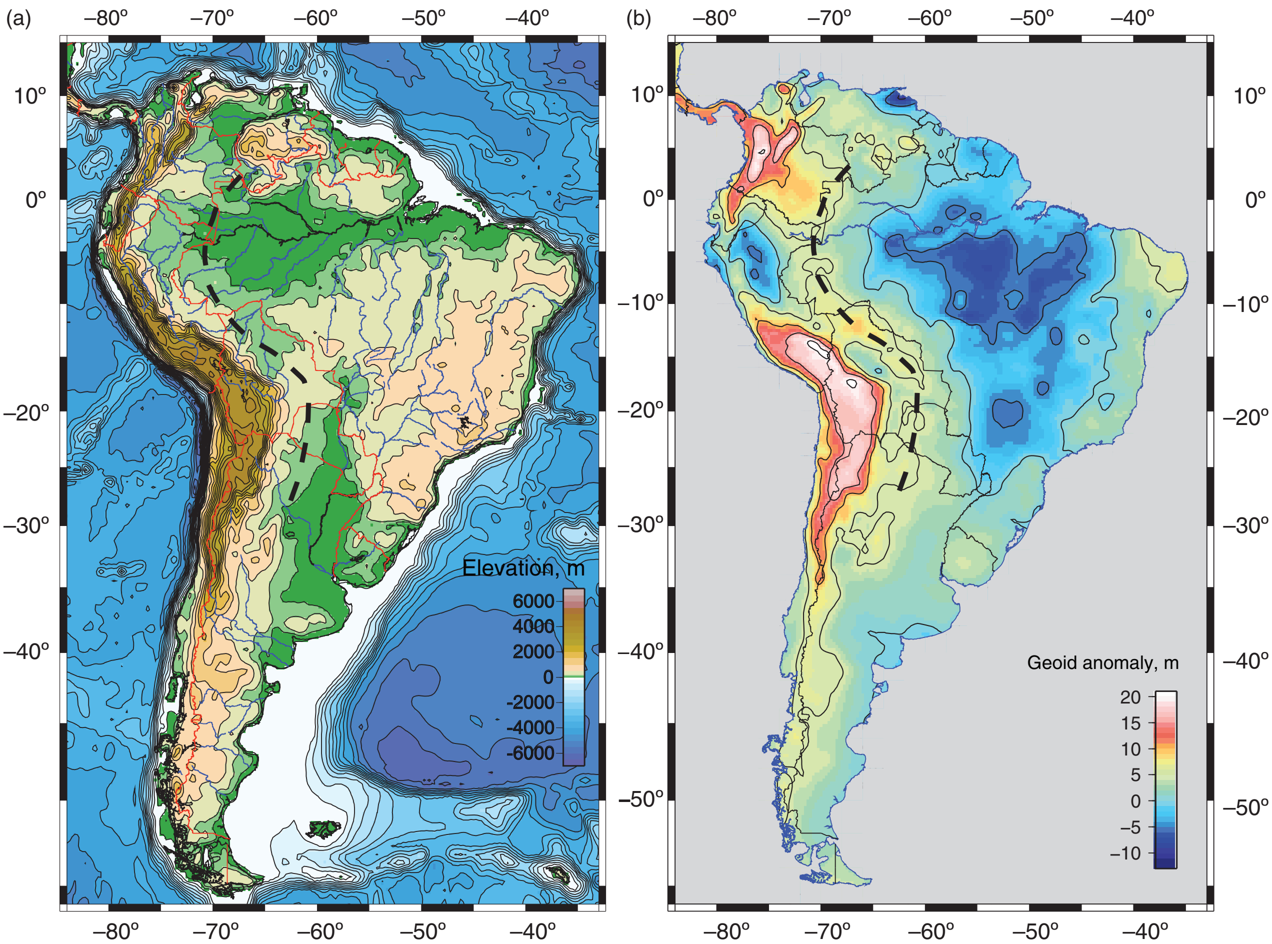



**Fig. 20.4.** The Andean foreland basin system in South America. (a) Topography of South America, with position of geoid anomaly high after Chase et al. (2009) indicated by dashed bold line ~400–500 km east of the Andes. Note that there is virtually no topographic expression of the forebulge, as the basin is overfilled. (b) Geoid anomaly in m, derived from GRACE data (Chase et al., 2009). Position of forebulge crest is indicated by bold dashed line running along the trend of aligned 5–10 m anomalies in the foreland.

related to viscous coupling, via the mantle wedge, between the subducting oceanic plate and the upper plate (Fig. 20.2; Mitrovica et al., 1989; Gurnis, 1992; Lithgow-Bertelloni and Gurnis, 1997). This "dynamic subsidence" may extend more than 1,000 km inboard, tilting the entire flexural profile that develops in response to topographic loading downward toward the trench (DeCelles and Giles, 1996; Pang and Nummedal, 1995; Catuneanu, 2004; Liu et al., 2005). In such cases, the forebulge and backbulge depozones are well developed and highly preservable (Catuneanu, 2004). If the foreland region is above sea level, the forebulge depozone is characterized by a sequence of stacked, hypermature paleosols reflecting the long-term low rates of sediment accumulation, such as those documented in the Paleogene foreland basin succession of the central Andes (DeCelles and Horton, 2003), at the Jurassic-Cretaceous boundary in the western North American Cordilleran foreland basin (Demko et al., 2004; Fuentes et al., 2009), and in the lower Cretaceous Gangdese foreland basin of southern Tibet (Leier et al., 2007). In marine retroarc foreland basin systems, such as the middle Cretaceous of the North American Cordillera, the trace of forebulge migration is marked by shallow marine shoaling (and possible carbonate build-up), erosion, and distinctive offlap and onlap patterns (Fig. 20.1; Tankard, 1986; Plint et al., 1993; Yang and Miall, 2009).

## COLLISIONAL FORELAND BASIN SYSTEMS

Collisional foreland basin systems develop on the subducting plate in intercontinental collisional settings (Fig. 20.2b; Dickinson, 1974; Miall, 1995; Jolivet and Faccenna, 2000). Modern examples are the Himalayan, Zagros, and New Guinean foreland basin systems (Dewey et al., 1989). No presently colliding intercontinental suture zone is longer than ~3000 km; consequently, Earth's modern collisional foreland basin systems are not as large as the retroarc systems in the Americas. Nevertheless, the broad array of colliding continental plates stretching discontinuously from the western Mediterranean to Myanmar forms a string of collisional foreland basins that stretches ~8000 km.

The Himalayan is generally regarded as the type example of an active collisional foreland basin system (Burbank et al., 1996; Najman, 2006). It is actually a composite of at least three foreland basin systems: the central Himalayan foreland basin, and the flanking Bengal and Indus basins (Fig. 20.5). Only the Himalayan sector of this orogenic system is a true collisional foreland basin system; the Bengal and lower Indus basins are founded upon lithosphere that is subducting more rapidly than local convergence rates, resulting in retreating collisional thrust belts and foreland basins (discussed in the following section).

The Himalayan foreland basin system is ~400–450 km wide and ~2000 km long (Fig. 20.5). The modern basin consists of active wedge-top and foredeep depozones. Bilham et al. (2003) argued that the north Indian highlands are the forebulge in front of the Himalayan thrust belt, and the analysis of Duroy et al. (1989) suggests that the Sarghoda Ridge is a prolongation of this forebulge that plunges northwestward beneath the Indus foreland region (Fig. 20.5). The north Indian forebulge consists of a rugged topographic surface with inherited (pre-Cenozoic) geomorphic relief that is flexed upward and incised by young drainages that cut down to the level of active flexure (Fig. 20.5b; Bilham et al., 2003). The magnitude of upward flexure is ~400 m, and the forebulge is >600 km wide. Significant additional topographic relief is inherited from the Paleozoic-Mesozoic uplands of northern India (Bilham et al., 2003). The great size of the north Indian forebulge results from the high flexural rigidity of Indian shield lithosphere and the extreme size of the Himalayan topographic load. Thus, the modern Indo-Gangetic foreland basin is deeply underfilled, with axial fluvial systems that exit the basin at sea level via the flanking lower Indus and Bengal basins, an erosional forebulge, and a non-depositional backbulge region (Fig. 20.5a). The situation has not always been like this however, as data from Eocene strata in the frontal part of the thrust belt indicate deposition of sediment derived from the nascent Himalaya in a restricted shallow-marine setting to the south of a mostly submerged forebulge (DeCelles et al., 1998a, 2004; Najman et al., 2005).

Unlike retroarc foreland basins, collisional foreland basins are not subject to far-field dynamic loading. Consequently, the main control on the shape and magnitude of the flexural profile is the orogenic load. If the foreland lithosphere is extremely rigid, such as that beneath and south of the modern Himalayan foreland basin, then it is unlikely that the forebulge and backbulge depozones will be preserved; instead, passage of the flexural wave should be marked by a strongly erosional unconformity (e.g., Crampton and Allen, 1995). On the other hand, if the foreland lithosphere is only moderately rigid, such as that beneath the Zagros foreland basin (Snyder and Barazangi, 1986; Watts, 2001; Watts and Burov, 2003), then backbulge and even forebulge deposits may be preserved in the stratigraphic record. For example, the marine Eocene backbulge deposits in the Himalayan foreland basin system are capped by an erosional unconformity or extremely mature but very thin Oxisol horizon representing almost all of Oligocene time along the entire west to east extent of the Himalaya (DeCelles et al., 1998a, 2004). Above the Oxisol lie upward coarsening, Miocene-Pliocene foredeep and locally wedge-top deposits (Fig. 20.3; Burbank et al., 1996; DeCelles et al., 1998b). Apparently the Himalayan foreland basin changed through time from a system that enabled preservation of the backbulge and forebulge depozones, to the modern situation in which these depozones are not accumulating significant sediment. The likely culprits for this change in character are the increasing size of the Himalayan thrust belt load and southward increasing flexural rigidity of Indian shield lithosphere. A more appropriate modern analog for the Eocene-Oligocene Himalayan system is the Zagros foreland basin, where the scale of the topographic load and rigidity of the Arabian lithosphere are not as extreme as those in the present Himalayan-Indian setting. Even the

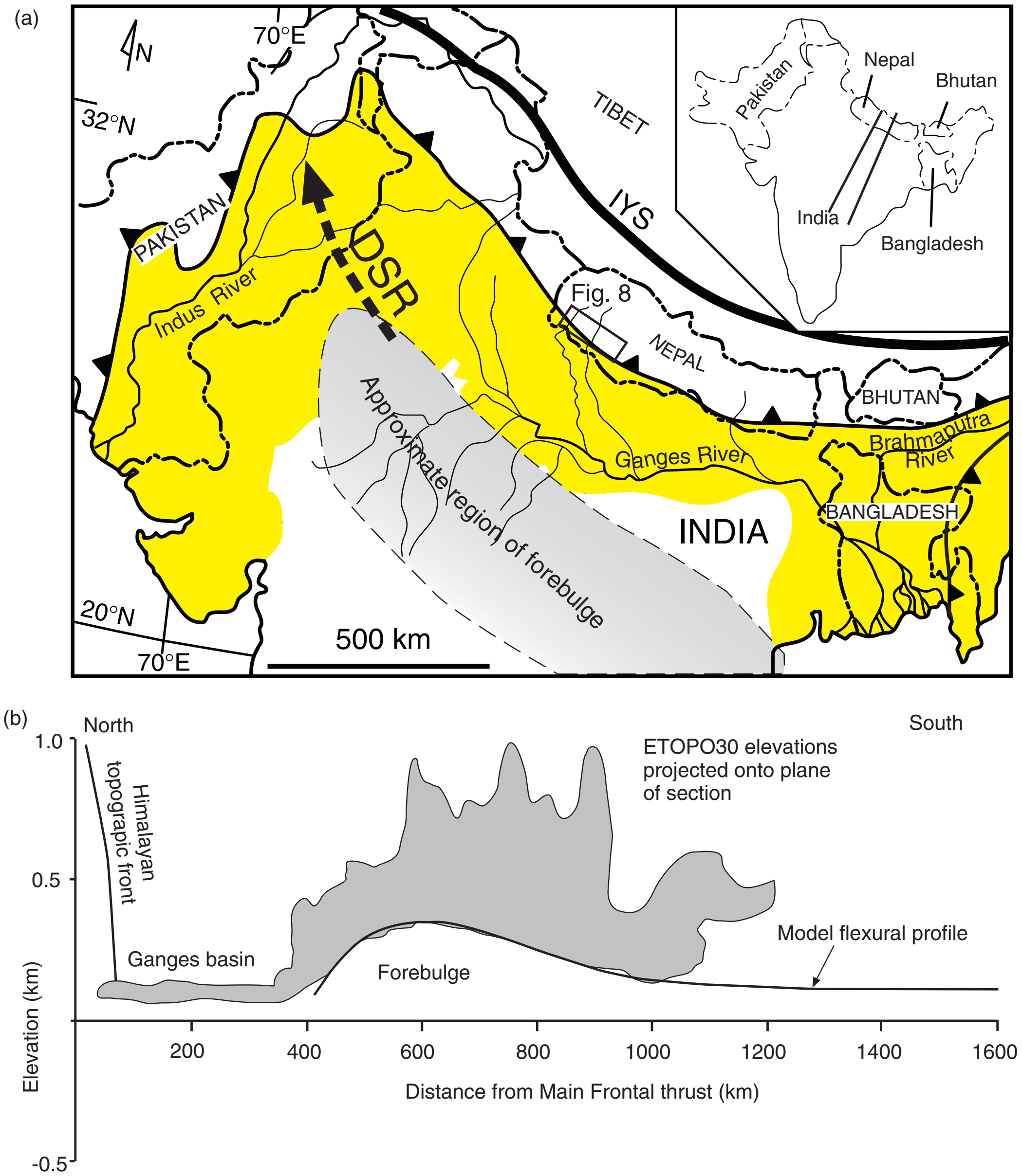



**Fig. 20.5.** (a) Map showing the collisional foreland basins (yellow) associated with the Indo-Asian intercontinental collision in Pakistan, India, and Bangladesh. No distinction is drawn between the central, Himalayan part of the system, and the flanking Indus and Bengal basins, which appear to be associated with retreationg collisional systems. Bold solid line is the Indus-Yarlung suture (IYS) zone; barbed line represents front of the thrust belts. Dashed bold line labeled DSR is the Delhi-Sargoda ridge, interpreted to be a forebulge by Duroy et al. (1989). Rectangular box in western Nepal shows region from which the data shown in Figure 20.8 were collected. Gray shaded area in northern India represents approximate area of the Himalayan flexural forebulge, based on GRACE geoid data (see www.csr.utexas.edu/grace/gravity/). (b) Mean topography versus distance from the frontal Himalaya in Nepal along longitudinal profiles at 25° and 30° (see profiles in part a inset), after Bilham et al. (2003). Solid line represents modeled time-dependent erosion surface (for effective elastic thickness of 105 km) that fits the bottom of envelope of elevation data (gray area) projected onto a single plane. This upward convex erosional base was interpreted by Bilham et al. (2003) as the geomorphic response to recent incision of older topography in northern Indian highlands as the Indian plate rides up over the flexural forebulge in front of the Himalayan thrust belt.

Zagros, however, is not a perfect analog for the early Himalayan system because its foreland region is affected by distal subsidence associated with regional tilting of Arabia during Neogene opening of the Red Sea and proximity to the Afar hotspot (Ali and Watts, 2009).

## RETREATING COLLISIONAL FORELAND BASIN SYSTEMS

Where local subduction rate exceeds regional convergence rate in a collisional tectonic setting, continued subduction requires that the hinge line of the subducting plate roll through the plate in a retrograde direction (i.e., opposite to the direction of subduction). In order to fill the gap that would be created between the two plates, the upper plate experiences extension and thinning (Fig. 20.2b; Malinverno and Ryan, 1986; Doglioni, 1991; Royden, 1993). The best known examples of this type of thrust belt-foreland basin system are in the Mediterranean region, where promontories along the northern margin of the African plate (e.g., Adria, Arabia) have been colliding with the Eurasian plate since latest Cretaceous time (Cavazza et al., 2004; Schmid et al., 2008), producing collisional orogenic belts and stranding aerially restricted slabs of Neotethyan oceanic lithosphere in the areas between the promontories (Fig. 20.6; Jolivet and Faccenna, 2000; Faccenna et al., 2004; Spakman and Wortel, 2004). Density-driven subduction of these oceanic slabs continues in spite of the reduced rate of plate convergence caused by collision along the promontories, and the subduction zone hinges are forced to roll back in a direction more or less opposite to that of plate subduction in order to accommodate ongoing subduction. The fold-thrust belt in the upper plate continues to migrate toward the foreland, but experiences regional extension and crustal thinning in its trailing part (Doglioni, 1991; Cavinato and DeCelles, 1999). Small oceanic basins may open in the wake of these migrating subduction zones, and in some cases, microplates have detached from the Eurasian plate and migrated rapidly across the central Mediterranean (e.g., the Corsica-Sardinia and Calabrian microplates and the Ballearic Islands; Malinverno and Ryan, 1986; Dewey et al., 1989; Bonardi et al., 2001; Gutscher et al., 2002; Booth-Rea et al., 2007; Fig. 20.6). The process is analogous to backarc spreading that characterizes upper plates in western Pacific oceanic subduction systems.

In contrast to collisional and retroarc systems, retreating collisional systems are characterized by relatively short (arc lengths of ca. 1000 km), highly arcuate (with up to 180° of curvature), low-elevation (<2 km) fold-thrust belts that

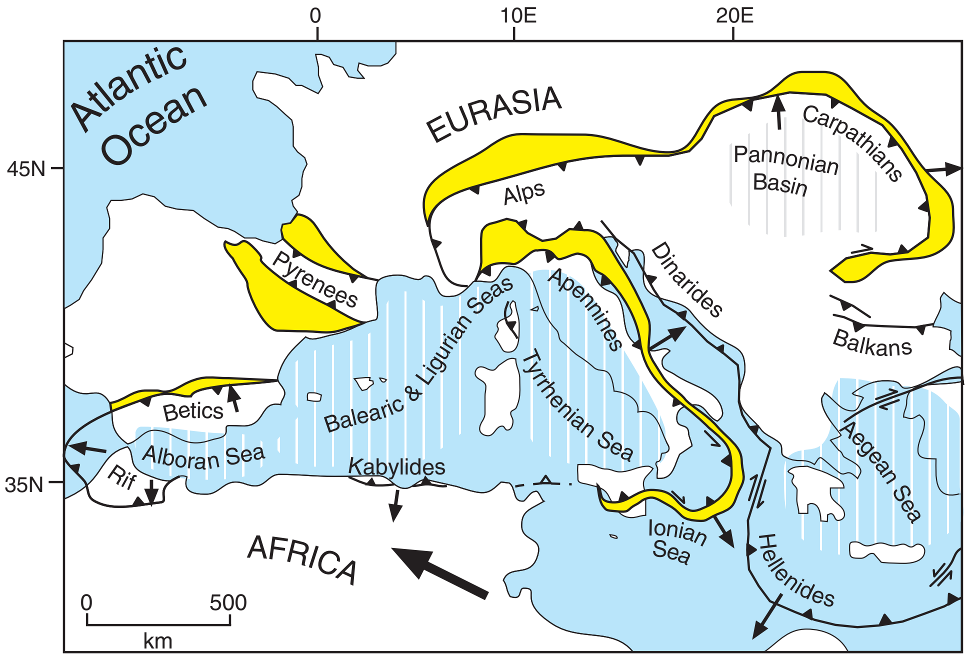



**Fig. 20.6.** Tectonic map of the central and western Mediterranean region, showing retreating subduction systems in front of highly arcuate Betic-Rif, Apennine-Ionian-Kabylidean, Carpathian, and Hellenide fold-thrust belts (barbed lines, with directions of subduction zone retreat/thrust belt advance shown by small arrows). Foreland basins are highlighted in yellow. Regions of crustal extension in upper plate are shown by vertical ruled pattern. Present motion of Africa relative to Eurasia is shown by large arrow. *Source*: After Jolivet and Faccenna (2000) and Schmid et al. (2008).

involve mainly unmetamorphosed sedimentary rocks (Royden, 1993). Youthful examples include the Apennine, Carpathian, Betic-Rif, and possibly North Alpine fold-thrust belts and their associated foreland basins (Fig. 20.6); older examples include the Antler (Devonian-Mississippian; Giles and Dickinson, 1995), Taconic (Ordovician; Jacobi, 1981; Hiscott et al., 1986), and Ouachita (Carboniferous; Houseknecht, 1986) foreland basins. Foreland basin systems associated with retreating subduction zones typically contain foredeep deposits that are more than twice as thick as would be expected from the sizes of the topographic loads, and probably require subducted slab loads (Karner and Watts, 1983; Royden, 1993). For example, the foreland basin in front of the Apennines thrust belt has accumulated locally more than 7 km of wedge-top and proximal foredeep sediment since Messinian time (Ori et al., 1986; Bigi et al., 1992). Royden and Karner (1984) showed that this extreme subsidence is not explained solely by flexural subsidence under the rather modest Apennine load; an additional subducted Adriatic slab load exerts a bending moment on the foreland lithosphere and drives the majority of the subsidence. A similar situation prevails in the Carpathian foreland basin (Royden and Karner, 1984).

Foreland basin systems in retreating collisional settings are characterized by relatively short wavelength, narrow foredeeps (ca. 100 km), narrow forebulges, and poorly developed backbulge depozones. Wedge-top deposits may be extremely thick (e.g., Bigi et al., 1992). Carbonate build-ups commonly are present on the forebulge, and turbidites and other deep marine deposits are prevalent in the foredeep depozone (Hiscott et al., 1986; Houseknecht, 1986; Wuellner et al., 1986; Giles and Dickinson, 1995; Sinclair, 1997). Along-strike variations in flexural rigidity and lateral tears in the subducting (retreating) slab are also common in retreating collisional systems, and these complexities strongly partition the foreland basin system (Matenco and Bertotti, 2000; Spakman and Wortel, 2004; Ustaszewski et al., 2008)

## STRATIGRAPHIC RECORDS OF FORELAND BASINS: DECONVOLVING TECTONIC PROCESSES

In principle, the flexural profile for a given orogenic load-foreland basin system may be inverted to produce the subsidence history for any point along the profile, if the rate of migration of the system can be estimated. The first derivative of the flexural profile, with time substituted for horizontal distance (Fig. 20.7), would represent the subsidence history. Comparison of various modeled subsidence histories with the actual subsidence history could allow estimation of flexural rigidity. Figure 20.8 illustrates an application of this technique to the Himalayan foreland basin record of Nepal. In this case, a rate of southward relative migration of 20 mm/yr is assumed on the basis of GPS data (Bettinelli et al., 2006) coupled with estimates for the long-term Neogene migration history (DeCelles et al., 1998b; Lavé and Avouac, 2000). The best-fit flexural rigidity is $1.0 \times 10^{24}$ Nm, which is consistent with gravity data and previous flexural analyses (Lyon-Caen and Molnar, 1985).

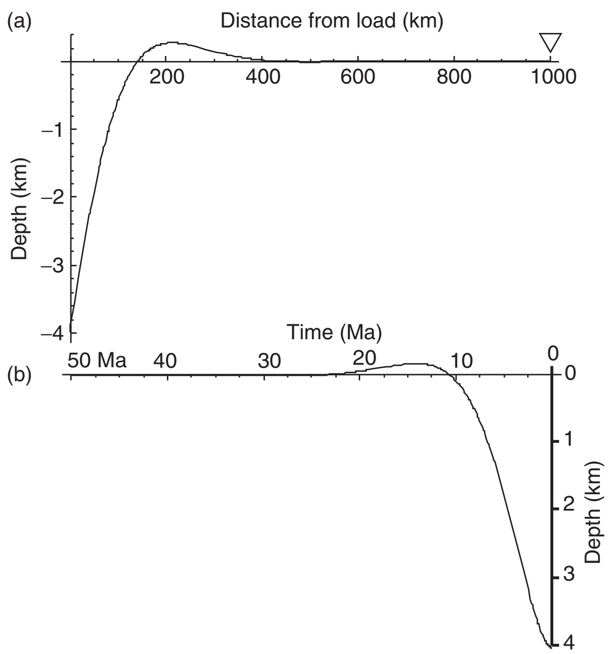



**Fig. 20.7.** (a) Flexural profile developed in response to a line load on a broken plate, positioned at 0 on the x-axis, with flexural rigidity = $10^{23}$ Nm, a contrast of 600 kg/m$^3$ between the density of the basin fill and that of the mantle, and a maximum basin depth of 4 km. The triangle at 1000 km on x-axis represents a location in the foreland, past which the flexural profile will migrate and produce the subsidence history shown in part (b). (b) The first derivative of the flexural profile shown in (a), with time substituted for space, in order to simulate the migration of the flexural profile past the upside down triangle at 1000 km at a constant rate of 20 mm/yr. This curve is the inverted subsidence history at the triangle.

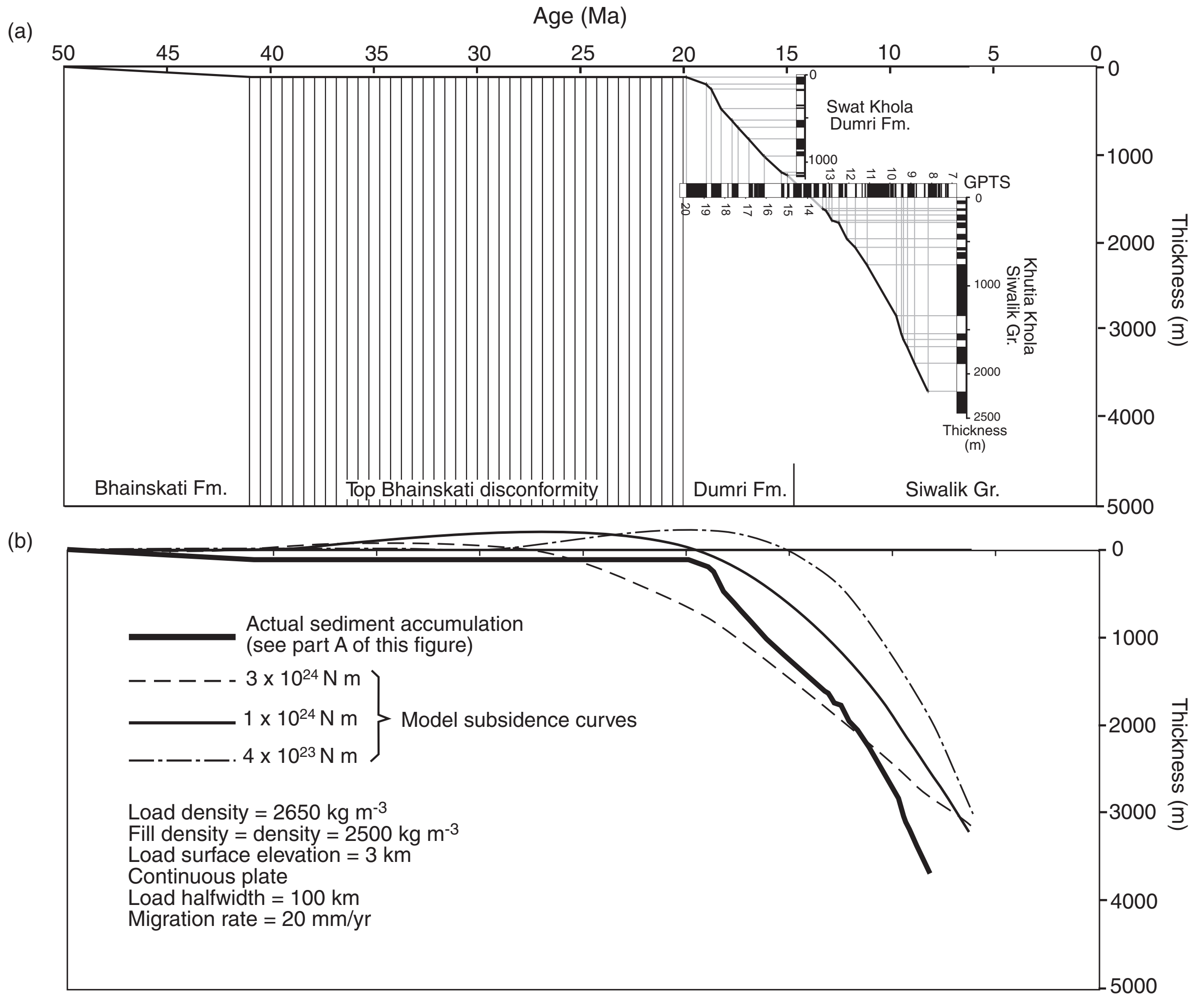



**Fig. 20.8.** (a) Time vs. thickness (not decompacted) curve for Himalayan Cenozoic foreland basin system in western Nepal. The ~20–8 Ma portion of the record is documented by magnetostratigraphic correlation of sections at Swat Khola and Khutia Khola in western Nepal (Ojha et al., 2008) to the Global Paleomagnetic timescale of Lourens et al. (2004). The 50–20 Ma part of the record is based on DeCelles et al. (1998a). (b) Bold line reproduces curve from part (a). Thinner lines represent theoretical subsidence curves inverted from flexural profiles using the flexural rigidities indicated in the legend, and assuming a 20 mm/yr migration rate of the flexural wave. All parameters of the flexural model (except flexural rigidity) were held constant, and are listed in the legend. Note that the vertical maximum predicted subsidence is easily adjusted to better match the actual curve by slightly changing the size of the load. The key aspect of these modeled curves is whether or not they exhibit the proper wavelength. Only the curve based on rigidity of $1 \times 10^{24}$ Nm comes close to matching the actual accumulation history, with a rapid transition from forebulge to foredeep depozones at ca. 20 Ma.

The problem with this simple approach is that the size of the orogenic load and the flexural rigidity of the foreland lithosphere, both of which influence the flexural wavelength, are not likely to remain constant over the time scales involved (≥50 Myr). In many cases, the orogenic load and the flexural rigidity are likely to increase over time, as the growing orogenic belt migrates farther onto older, stiffer cratonic lithosphere. Thus, it is necessary to first palinspastically restore the orogenic system, including the foreland basin system, and to make estimates of the size of the orogenic load through time. For example, Jordan (1981), Beaumont (1981), Quinlan and Beaumont (1984), and Liu et al. (2005) used temporally evolving orogenic loads in the Appalachian and Cordilleran orogenic belts to model the migration history of foredeep depozones. An additional potential problem with this approach is that rates of shortening may vary during the lifetime of a thrust belt, and consequently the rate of flexural wave migration may be highly unsteady.

Incorporation of the four-part foreland basin system into palinspastic restorations allows for more constrained estimates of flexural wave migration because both the foredeep and forebulge may be used to position the flexural wave through time within the context of palinspastically restored foreland basin stratigraphy. Of particular value is the temporal duration of the forebulge depozone/disconformity. When coupled with the rate of migration of the system, the width of the forebulge depozone can constrain the flexural rigidity; conversely, if the flexural rigidity can be estimated, then the temporal duration of the forebulge depozone in any given vertical stratigraphic section may be used to calculate the rate of migration of the flexural wave. In orogenic systems that are still active or have not been completely dismembered by later extension, the distance of flexural wave migration ($F$) is approximately equal to the sum of the width of the thrust belt ($P$) plus the shortening ($S$) (Fig. 20.1b). Thus, with reasonable estimates of any two of these variables ($F, P, S$) the third can be calculated (DeCelles and DeCelles, 2001).

Figure 20.9 illustrates a palinspastic flexural model of the Himalayan thrust belt and foreland basin system in Nepal, based on incremental restorations of balanced regional cross-sections (DeCelles et al., 2001; Robinson et al., 2006). Thrust loads are simplified as rectangular blocks, mainly because the paleotopography is unconstrained. The different tectonostratigraphic subzones of the Himalayan thrust belt (Tethyan, Greater, Lesser, and Sub-Himalayan zones) are shown by different colored blocks. The model depicts palinspastically restored reference points within the context of the evolving foreland basin flexural wave and the growing Himalayan thrust belt load. Although this model is consistent with available stratigraphic records from the foreland basin preserved in the Lesser Himalayan and Subhimalayan zones, and with shortening estimates in the thrust belt, large parts of the record are poorly documented. In particular, the shortening history of the thrust belt during pre-Miocene time is not well understood. What is known is that (a) the foreland region was the locus of shallow, restricted marine deposition derived in part from the nascent thrust belt during Eocene time (Fig. 20.9c; DeCelles et al., 2004; Najman et al., 2005), (b) that the region was beveled by a major disconformity during much of Oligocene time (Fig. 20.9d–e), and that the region was in the foredeep by early Miocene time at the latest (Fig. 20.9f). Moreover, frames g and h are well constrained by cross-cutting relationships and the modern to Miocene stratigraphic record of the foreland basin (Lyon-Caen and Molnar, 1985; DeCelles et al., 1998b; Szulc et al., 2006; Ojha et al., 2008). Flexural rigidity through time is unconstrained, but can be inverted from flexural profiles that fit best with the palinspastically restored positions of foreland depozones.

## INFLUENCE OF TECTONIC SETTING

Tectonic setting of the foreland basin system should be reflected in its stratigraphic record. Collisional systems lack a mechanism for dynamic subsidence, so forebulge erosion should be the norm, particularly in situations with high flexural rigidity of the foreland lithosphere such as the modern Himalayan foreland basin system. Retroarc systems, which may be strongly influenced by dynamic subsidence (Catuneanu, 2004; Liu et al., 2005), have well-preserved forebulge depozones and relatively thick backbulge deposits (Currie, 2002; Leier et al., 2007). Retreating collisional settings should be characterized by narrow, overthick, deep-marine foredeep depozones, and thick, well-developed wedge-top depozones (Ori et al., 1986). Forebulges in these settings are marked mainly by unconformities, and backbulge deposits are rare to absent (Crampton and Allen, 1995; Sinclair, 1997; Burkhard and Sommaruga, 1998). To these general stratigraphic characteristics may be added distinctions based on the composition of detritus in the foreland basin fill. Garzanti et al. (2007) reviewed this topic in detail, showing that significant differences exist among what are referred to in this chapter as retroarc, collisional, and retreating collisional foreland basin systems (referred to by Garzanti et al. (2007) as Apennine-, Andean-, and Alpine-type, respectively).

## REMAINING QUESTIONS

Some key remaining questions about the stratigraphic records of foreland basin systems include

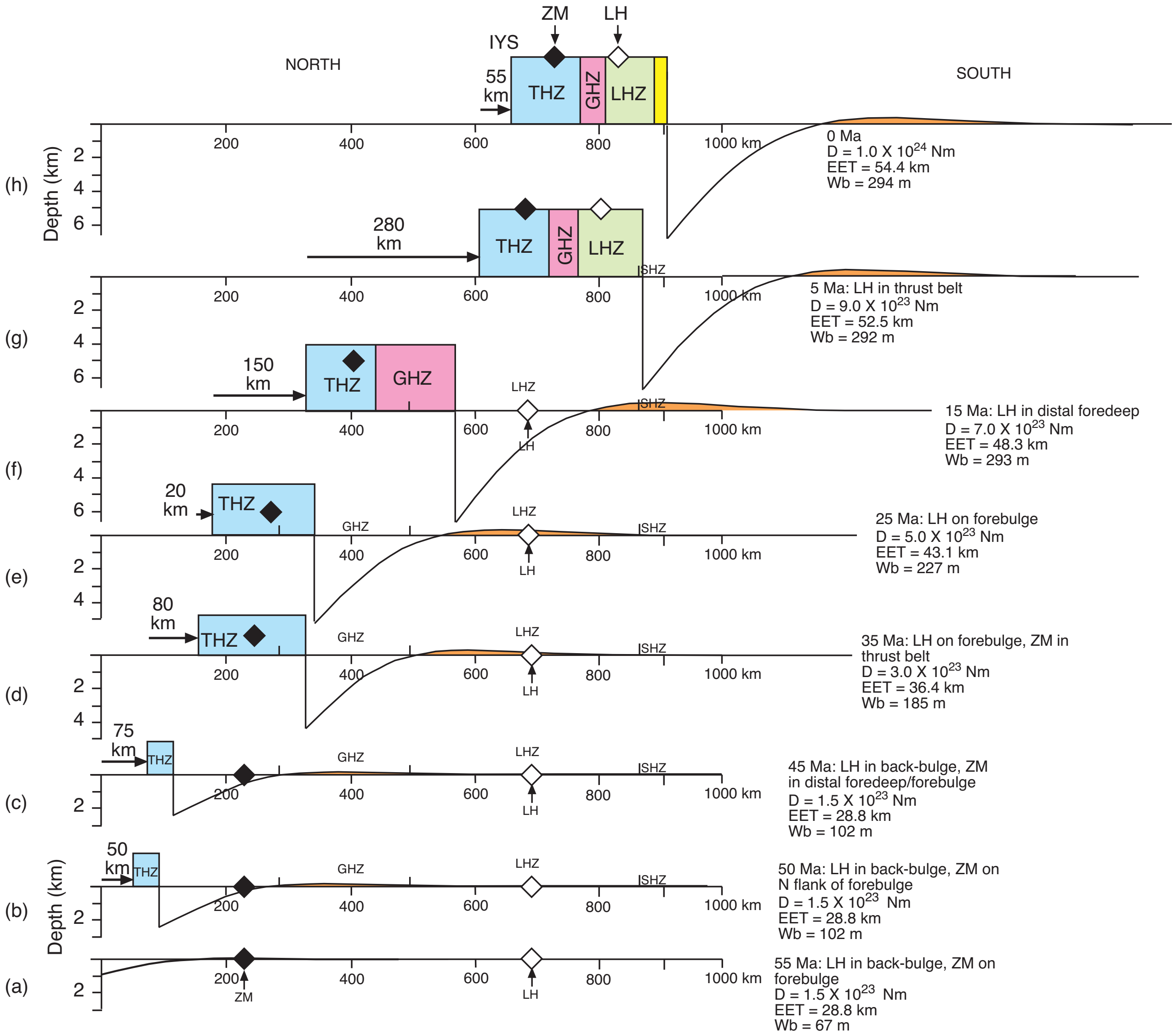



**Fig. 20.9.** Flexural model for the Himalayan thrust belt and foreland basin system, using rectangular loads and based on palinspastic restoration of the thrust belt since 55 Ma (DeCelles et al., 2001). Initial condition uses a line load on a broken plate; all subsequent models use a continuous plate and rectangular loads as illustrated. Density of the loads is 2650 kg/m$^3$; density of basin fill is 2500 kg/m$^3$. Amounts of shortening are indicated at each stage by arrows on left sides of northernmost block. Diamonds labeled ZM and LH represent palinspastic locations of the Zhepure Mountain and Lesser Himalayan sections of Tertiary strata. THZ = Tethyan Himalayan zone; GHZ = Greater Himalayan zone; LHZ = Lesser Himalayan zone. Values of changing flexural rigidity (*D*), effective elastic thickness (*EET*), and forebulge height (*Wb*) are indicated at each stage of the model.

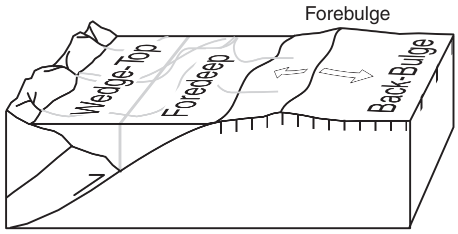
the issue of forebulge and backbulge preservation, the implications of long-distance flexural wave migration, the effect on foreland basin systems of structural complexity in the foreland lithosphere, and the largely untapped archive of information in foreland basins concerning relationships between tectonics and climate.

### Preservation of forebulge and backbulge depozones

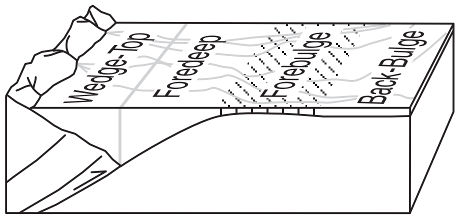
The issue of stratigraphic preservation is an important topic because the backbulge and forebulge depozones may represent half of the temporal evolution of the adjacent orogenic belt.

In fact, these deposits are commonly overlooked in stratigraphic analyses of foreland basins because of their thinness, textural and compositional maturity, and “pre-orogenic” appearance.

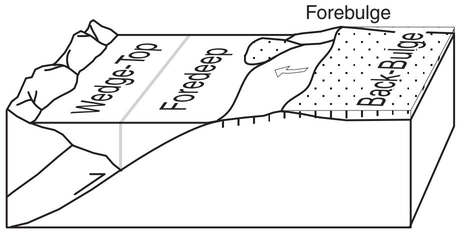
Two key controls on the preservation potential of backbulge sediments in a foreland basin system are the level of the sedimentary fill relative to the crest of the forebulge and the polarity of sediment supply (Fig. 20.10; e.g., Giles and Dickinson, 1995). Sediments may be derived from either or both sides of the foreland basin system, and if the forebulge is exposed and subject to erosion it too may supply sediment to both the backbulge and foredeep depozones (Blisniuk et al., 1998). If the foredeep and backbulge depozones are both underfilled with respect to the crest of the forebulge, then the forebulge will be exposed and eroded, particularly in nonmarine settings (Fig. 20.10a; Jacobi, 1981; Coakley and Watts, 1991; Crampton and Allen, 1995). In foreland basin systems in which the forebulge is not buried, the rate of sediment accumulation in the most distal part of the foredeep depozone may be low enough that paleosols overprint much of the stratigraphic record (Tandon et al., 2008). In such cases it is unlikely that much backbulge sediment will be preserved in the stratigraphic sequence of the foreland basin system. Even if the forebulge is below sea level, submarine erosion may bevel strata at its crest (Crampton and Allen, 1995). If the foredeep and backbulge depozones are both filled to levels above the crest of the forebulge, then the forebulge will be a zone of deposition of severely condensed strata and backbulge strata may be preserved (Fig. 20.10b; Flemings and Jordan, 1989; DeCelles and Horton, 2003). Similarly, if the foredeep is underfilled but the backbulge depozone is filled, then some sediment deposited in the backbulge depozone may be preserved from erosion over the crest of the forebulge (Fig. 20.10c). Data from foreland basins worldwide suggest that forebulge and backbulge depozones are best preserved in retroarc settings, perhaps because long-wavelength dynamic subsidence may overwhelm the entire flexural profile. On the other hand, preservation of thick wedge-top deposits, which are normally vulnerable to erosion in the frontal thrust belt, may be enhanced by slab load-driven subsidence in retreating collisional foreland basin systems.




(a) Foredeep and forebulge underfilled: Erosion of forebulge, development of unconformity, axial foredeep drainage




(b) Foredeep and forebulge overfilled: Forebulge is buried and subdued, stratigraphic condensation along crest of forebulge, transversal foredeep drainage




(c) Foredeep is underfilled and back-bulge is filled: Cratonic side of forebulge is buried and stratigraphic condensation results; foredeep side of forebulge experiences erosion

**Fig. 20.10.** Schematic block diagrams illustrating various aggradation/degradation situations in a nonmarine foreland-basin system, modified after Crampton and Allen (1995). (a) Foredeep is not filled to the crest of the forebulge (“underfilled” state); forebulge is exposed and forms a topographic drainage divide between foredeep and backbulge depozones. Deep weathering (shown by vertical-hatching) of pre-orogenic rocks and unconformity development take place in the forebulge district. Depending on base level, the backbulge region may also experience erosion and unconformity development. (b) Forebulge is buried by sediment (“overfilled” state), derived from the fold-thrust belt and/or the craton. Although aggradation occurs on forebulge crest, the rates of accumulation are very low and sediments are intensely weathered and condensed (vertical hatching). (c) The foredeep is underfilled but the backbulge depozone is filled to crest of forebulge with sediment derived from either the craton or out-of-plane pathways from the fold-thrust belt. Locally, sediments may spill into the foredeep from the backbulge depozone.

## Significance of lateral mobility

As pointed out by Flemings and Jordan (1989), understanding the lateral mobility of the coupled thrust belt-foreland basin system is crucial for proper interpretation of the foreland basin stratigraphic record. Onset of rapid subsidence is commonly viewed as the timing of onset of thrusting and flexural loading in the orogenic belt. However, the onset of rapid flexural subsidence across a foreland region is highly time-transgressive (Vergés et al., 1998), such that regions that lie in the backbulge or forebulge depozones during early thrusting will not begin to subside until the foredeep migrates into the region. At typical rates of flexural wave migration, this may take several tens of millions of years. If backbulge deposits are not preserved, then the early stratigraphic record of orogeny may be lost as erosion claims the early wedge-top and foredeep records.

The issue of lateral mobility is also important for stratigraphic modeling of grain size trends and unconformities in medial to distal foreland basin deposits. The popular two-phase model of foreland basin stratigraphy, in which fine-grained deposits are regarded as syntectonic whereas coarse-grained facies are regarded as "anti-tectonic" (Heller et al., 1988), is founded upon the notion that the foreland basin flexes downward and upward in response to changes in load size (e.g., Catuneanu et al., 1997; Willis, 2000). For a given flexural rigidity, growing loads generate greater flexural subsidence in the foredeep and trap coarse-grained sediment close to the thrust belt; erosionally decaying loads promote isostatic uplift of the thrust belt and proximal foredeep, thereby reducing accommodation in the proximal region and allowing coarse sediment to prograde into distal parts of the system. Conceptually attractive as this model is, it relies upon a static load location and episodes of inactivity in the thrust belt. Whereas some thrust belts may experience periods of tectonic inactivity, continuing plate convergence is expected to maintain an actively shortening thrust belt, which in turn drives the foreland flexural wave laterally across the lithosphere throughout the history of the thrust belt. Even during periods of out-of-sequence thrusting, the foreland lithosphere must continue to feed rocks into the shortening thrust belt. Palinspastic restoration of the foreland basin system (e.g., Homewood et al., 1986; Pfiffner, 1986; Sinclair, 1997; Vergés et al., 1998; Currie, 2002), as well as the thrust belt itself, is necessary for proper interpretation of long-term textural trends in the basin fill.

## Inherited and reactivated structures

Located on flexed continental lithosphere, foreland basin systems are sensitive to reactivation of inherited basement structures and subtle partitioning by minor faults associated with the upflexed forebulge arch. Structural fabrics associated with ancient cratonic structures and pre-existing rifted margins are especially susceptible to reactivation (Schwartz and DeCelles, 1988; Bradley and Kidd, 1991; Meyers et al., 1992; Crampton and Allen, 1995; Blisniuk et al., 1998; Gupta and Allen, 2000; Zaleha, 2001; Londoño and Lorenzo, 2004). The effects of subtle intraforeland structures are strongest in the forebulge region because the onset of flexural uplift produces tensional fiber stresses that break the upper crust along minor but potentially numerous normal faults. Although these structures are minor in amplitude (typically only tens of meters), they may control drainage patterns and local sediment provenance in the distal foredeep and forebulge depozones. These types of structures are also of potential importance for hydrocarbon exploration because they may influence reservoir architecture and distribution.

## The underutilized archive

As much as foreland basin systems have been studied, they remain an underutilized archive of information about the interactions between orogenic and climatic processes. Major thrust belts, such as the Andes and the Himalaya, control the greatest orographic climate gradients on Earth (Grujic et al., 2006; Strecker et al., 2007; Bookhagen and Strecker, 2008) and their flanking foreland basins contain the records of changing climate and exhumation patterns in the forms of autochthonous and detrital organic matter, in situ chemically precipitated constituents (e.g., cements and paleosol components), and mineral detritus. Recent breakthroughs in detrital geochronology and thermochronology, including multi-dating methods, have potential to reveal complex cooling and provenance histories (e.g., Bernet et al., 2006; Carrapa et al., 2009; Najman et al., 2009). Synthesis of these new data sets into a geodynamic framework established by stratigraphic analysis in terms of foreland basin depozones will provide holistic insights into the evolution of orogenic and climatic systems.

## SUMMARY

- The well-documented, four-part districting of foreland basin systems in both modern and ancient (stratigraphic) contexts provides a powerful constraint for unraveling the stratigraphic history of orogeny. Long-distance (up to $\sim$1000 km) lateral migration of foreland flexural waves stacks depozones vertically in a "Waltherian" succession which may be interpreted in terms of thrust belt propagation and shortening velocities, given sufficient chronostratigraphic information. Of particular utility is the forebulge disconformity/condensation zone, because its duration may be used to estimate the speed of flexural wave migration. Lack of recognition of forebulge and backbulge deposits in foreland basin successions precludes stratigraphic identification of the early history of an orogeny.
- Forebulge and backbulge depozone preservation is enhanced by dynamic subsidence in retroarc settings, whereas collisional and retreating collisional systems commonly do not preserve forebulge and backbulge deposits because they lack regional far-field subsidence mechanisms and because they develop high and/or narrow forebulge arches. Nevertheless, the early (Eocene-Oligocene) Himalayan foreland basin system contains thin backbulge deposits and a major disconformity/paleosol condensation zone that marks forebulge passage. The transition to the modern situation in which neither forebulge nor backbulge is depositional suggests that increasing stiffness of Indian lithosphere involved in the Indo-Asian collision zone is to blame for loss of the Miocene-Recent forebulge and backbulge stratigraphic records.
- Inversion of foreland basin subsidence curves as a proxy for the time-transgressive flexural profile provides a first order estimate of the flexural properties of the foreland lithosphere. However, likely significant changes (usually increases) in flexural rigidity of underthrusting foreland lithosphere, its velocity of underthrusting, and the size of the orogenic load require palinspastic restoration of the thrust belt-foreland basin system to fully characterize the long-term geodynamics.
- Activation and reactivation of subtle intraforeland structures are common in foreland basin systems, particularly in the forebulge depozone. Although they are typically minor in scale, these structures are capable of controlling local drainage and provenance patterns, as well as the distribution of lithofacies.

## ACKNOWLEDGMENTS

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