

# Foreland basin systems

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## ABSTRACT

A foreland basin system is defined as: (a) an elongate region of potential sediment accommodation that forms on continental crust between a contractional orogenic belt and the adjacent craton, mainly in response to geodynamic processes related to subduction and the resulting peripheral or retroarc fold–thrust belt; (b) it consists of four discrete depozones, referred to as the *wedge-top*, *foredeep*, *forebulge* and *back-bulge* depozones – which of these depozones a sediment particle occupies depends on its location at the time of deposition, rather than its ultimate geometric relationship with the thrust belt; (c) the longitudinal dimension of the foreland basin system is roughly equal to the length of the fold–thrust belt, and does not include sediment that spills into remnant ocean basins or continental rifts (impactogens).

The wedge-top depozone is the mass of sediment that accumulates on top of the frontal part of the orogenic wedge, including ‘piggyback’ and ‘thrust top’ basins. Wedge-top sediment tapers toward the hinterland and is characterized by extreme coarseness, numerous tectonic unconformities and progressive deformation. The foredeep depozone consists of the sediment deposited between the structural front of the thrust belt and the proximal flank of the forebulge. This sediment typically thickens rapidly toward the front of the thrust belt, where it joins the distal end of the wedge-top depozone. The forebulge depozone is the broad region of potential flexural uplift between the foredeep and the back-bulge depozones. The back-bulge depozone is the mass of sediment that accumulates in the shallow but broad zone of potential flexural subsidence cratonward of the forebulge. This more inclusive definition of a foreland basin system is more realistic than the popular conception of a foreland basin, which generally ignores large masses of sediment derived from the thrust belt that accumulate on top of the orogenic wedge and cratonward of the forebulge.

The generally accepted definition of a foreland basin attributes sediment accommodation solely to flexural subsidence driven by the topographic load of the thrust belt and sediment loads in the foreland basin. Equally or more important in some foreland basin systems are the effects of subduction loads (in peripheral systems) and far-field subsidence in response to viscous coupling between subducted slabs and mantle–wedge material beneath the outboard part of the overlying continent (in retroarc systems). Wedge-top depozones accumulate under the competing influences of uplift due to forward propagation of the orogenic wedge and regional flexural subsidence under the load of the orogenic wedge and/or subsurface loads. Whereas most of the sediment accommodation in the foredeep depozone is a result of flexural subsidence due to topographic, sediment and subduction loads, many back-bulge depozones contain an order of magnitude thicker sediment fill than is predicted from flexure of reasonably rigid continental lithosphere. Sediment accommodation in back-bulge depozones may result mainly from aggradation up to an equilibrium drainage profile (in subaerial systems) or base level (in flooded systems). Forebulge depozones are commonly sites of unconformity development, condensation and stratal thinning, local fault-controlled depocentres, and, in marine systems, carbonate platform growth.

Inclusion of the wedge-top depozone in the definition of a foreland basin system requires that stratigraphic models be geometrically parameterized as doubly tapered prisms in transverse cross-sections, rather than the typical ‘doorstop’ wedge shape that is used in most

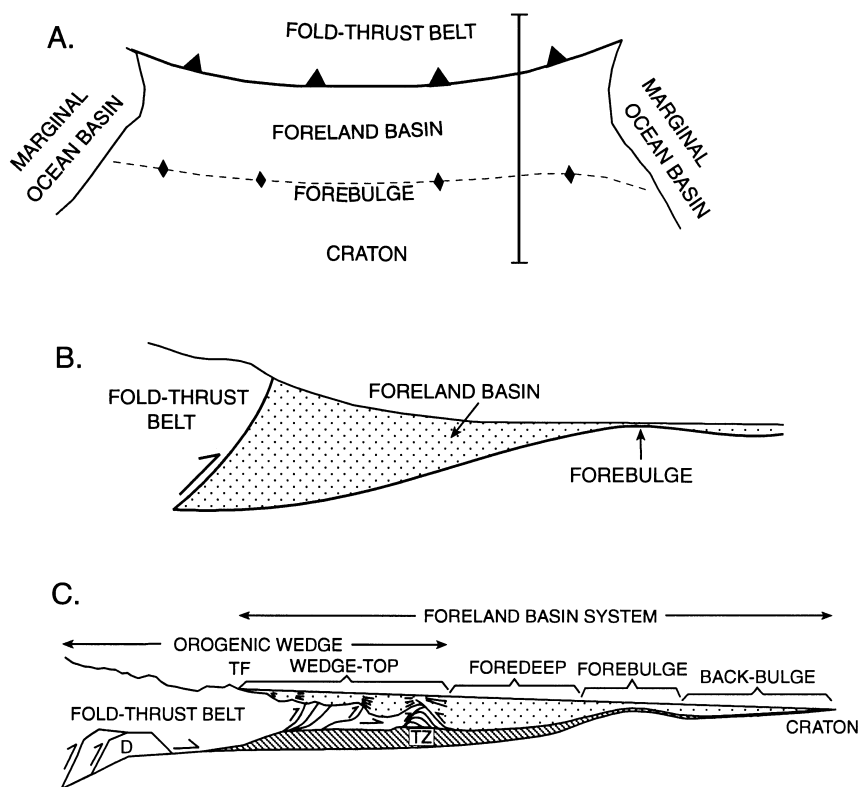
models. For the same reason, sequence stratigraphic models of foreland basin systems need to admit the possible development of type I unconformities on the proximal side of the system. The oft-ignored forebulge and back-bulge depozones contain abundant information about tectonic processes that occur on the scales of orogenic belt and subduction system.

**INTRODUCTION**

This paper addresses the existing concept of a foreland basin (Fig. 1A,B): its definition, areal extent, pattern of sedimentary filling, structure, mechanisms of subsidence and the tectonic implications of stratigraphic features in the basin fill. Our aim is to point out some inadequacies in the current conception of a foreland basin, propose a more comprehensive definition and to elaborate upon some of the features of this expanded definition.

A foreland basin generally is defined as an elongate trough that forms between a linear contractional orogenic belt and the stable craton, mainly in response to flexural subsidence that is driven by thrust-sheet loading in the

orogen (Fig. 1A,B; e.g. Price, 1973; Dickinson, 1974; Beaumont, 1981; Jordan, 1981, 1995; Lyon-Caen & Molnar, 1985). The term ‘foredeep’ (Aubouin, 1965) is used interchangeably with foreland basin. Important ancillary concepts that are equally entrenched in the literature are: (i) foreland basin sediment fill is wedge-shaped in transverse cross-section, with the thickest part located directly adjacent to, or even partially beneath, the associated thrust belt (Fig. 1B; Jordan, 1995); (ii) foreland basin sediment is derived principally from the adjacent thrust belt, with minor contributions from the cratonward side of the basin (Dickinson & Suczek, 1979; Schwab, 1986; DeCelles & Hertel, 1989); and (iii) a flexural bulge, or forebulge, may separate the main part



**Fig. 1.** (A) Schematic map view of a ‘typical’ foreland basin, bounded longitudinally by a pair of marginal ocean basins. The scale is not specified, but would be of the order of 102–103 km. Vertical line at right indicates the orientation of a cross-section that would resemble what is shown in part B. (B) The generally accepted notion of foreland-basin geometry in transverse cross-section. Note the unrealistic geometry of the boundary between the basin and the thrust belt. Vertical exaggeration is of the order of 10 times. (C) Schematic cross-section depicting a revised concept of a foreland basin system, with the wedge-top, foredeep, forebulge and back-bulge depozones shown at approximately true scale. Topographic front of the thrust belt is labelled TF. The foreland basin system is shown in coarse stipple; the diagonally ruled area indicates pre-existing miogeoclinal strata, which are incorporated into (but not shown within) the fold-thrust belt toward the left of diagram. A schematic duplex (D) is depicted in the hinterland part of the orogenic wedge, and a frontal triangle zone (TZ) and progressive deformation (short fanning lines associated with thrust tips) in the wedge-top depozone also are shown. Note the substantial overlap between the front of the orogenic wedge and the foreland basin system.

of the foreland basin from the craton (e.g. Jacobi, 1981; Karner & Watts, 1983; Quinlan & Beaumont, 1984; Crampton & Allen, 1995). Most workers in practice consider the basin to be delimited by the thrust belt on one side and by the undeformed craton on the other side, although some well-known foreland basins 'interfere' with extensional basins orientated at a high angle to the trend of the orogenic belt (e.g. the Amazon and Alpine forelands; the 'impactogens' of Şengor, 1995). Longitudinally, foreland basins commonly empty into marginal or remnant oceanic basins (Fig. 1A; Miall, 1981; Covey, 1986; Ingersoll *et al.*, 1995) or zones of backarc spreading (Hamilton, 1979). Dickinson (1974) distinguished between 'peripheral' foreland basins, which form on subducting plates in front of thrust belts that are synthetic to the subduction direction, and 'retroarc' foreland basins, which develop on the overriding plates inboard of continental-margin magmatic arcs and associated thrust belts that are antithetic to the subduction direction. Although this distinction has stood the test of two decades of research, only recently have the geodynamic differences between these two, fundamentally different, types of foreland basins been recognized (e.g. Gurnis, 1992; Royden, 1993).

The concept of a foreland basin as outlined above is incomplete in two important respects. First, it is clear from many modern and ancient foreland settings that sediment derived from the thrust belt, as well as sediment derived from the forebulge region and craton and intrabasin carbonate sediment, may be deposited over areas extending far beyond the zone of major flexural subsidence (i.e. cratonward from the forebulge). Examples include the sediment accumulations associated with the Cordilleran, Amazonian and Indonesian orogenic belts, which extend hundreds of kilometres beyond their respective limits of major flexural subsidence (Fig. 2; Ben Avraham & Emery, 1973; Jordan, 1981; Karner & Watts, 1983; Quinlan & Beaumont, 1984). On the other hand, sediment accumulations in some foreland settings, such as the Swiss molasse basin (Sinclair & Allen, 1992), the Taiwan foreland basin (Covey, 1986) and the Po-Adriatic foreland basin (Royden & Karner, 1984; Ricci Lucchi, 1986; Ori *et al.*, 1986), are much narrower and confined to the zone of major flexural subsidence. This gives rise to the concept that foreland basins can exist in underfilled, filled or overfilled states (Covey, 1986; Flemings & Jordan, 1989). In practice, however, the part of the basin fill that extends toward the craton beyond the flexural bulge is given only passing attention in the literature (e.g. Quinlan & Beaumont, 1984; Flemings & Jordan, 1989; DeCelles & Burden, 1992). A key question is what causes the widespread accommodation and sediment accumulation cratonward of the crest of the forebulge.

Also ignored by the popular conception of foreland basins is a substantial amount of sediment derived from the orogenic wedge that accumulates on top of the wedge.

These accumulations generally are considered as 'piggyback' or 'thrust top' basins (Ori & Friend, 1984) that may or may not be isolated from the main part of the foreland basin fill. Although some piggyback basins are, indeed, geomorphically isolated for significant time periods (e.g. Beer *et al.*, 1990; Talling *et al.*, 1995), most active wedge-top accumulations are completely contiguous with the foreland basin. In the Upper Cretaceous foreland basin fill of the western interior USA, isopach contours are continuous on both sides of the basin, demonstrating that the basin fill tapers toward the thrust belt and is not wedge-shaped in transverse cross-section (Fig. 3). The type examples of piggyback basins cited by Ori & Friend (1984) in the Po-Adriatic foreland completely bury the underlying thrust-related basement topography beneath >3 km of sediment (Fig. 4); in the nonmarine part of the basin, tributaries from the northern Apennines thrust belt are graded across a smooth alluvial plain to the modern Po River. The limit of topographic expression of the thrust belt is along the Apennine front, but seismicity associated with blind thrusting and related folding extends at least 50 km to the north and north-east (Ori *et al.*, 1986). Thick and areally widespread synorogenic sediments on top of ancient thrust belts have been documented as well (Burbank *et al.*, 1992; DeCelles, 1994; Pivnik & Johnson, 1995; among others). If these accumulations are included in the foreland basin fill, as we believe they should be, then the geometry of the fill no longer has the wedge shape that is routinely used when modelling foreland basin subsidence and sedimentation patterns (e.g. Heller & Paola, 1992). Instead, in transverse cross-section, the basin fill tapers toward both craton and orogenic belt (Fig. 1C), and the asymmetric wedge shape, where it exists, is a result of *post-depositional* structural processes (mainly truncation by thrust faults), rather than a direct result of interacting depositional processes and subsidence patterns.

Several problems in current foreland basin modelling and field studies, examples of which will be discussed in this paper, can be traced to the inadequate concept of a foreland basin as outlined above. The remainder of this paper is devoted to proposing a more comprehensive definition for foreland basins, and discussing some of the implications of this definition for our understanding of foreland basin strata in terms of tectonic processes.

## FORELAND BASIN SYSTEMS DEFINED

The discussion above highlights the fact that 'foreland basins' are geometrically complex entities, comprising discrete parts that are integrated to varying degrees. Thus, we introduce the concept of a *foreland basin system*. (i) Foreland basin systems are elongate regions of potential sediment accommodation that form on continental crust between contractional orogenic belts and cratons in response to geodynamic processes related to the orogenic belt and its associated subduction system. (ii) Foreland

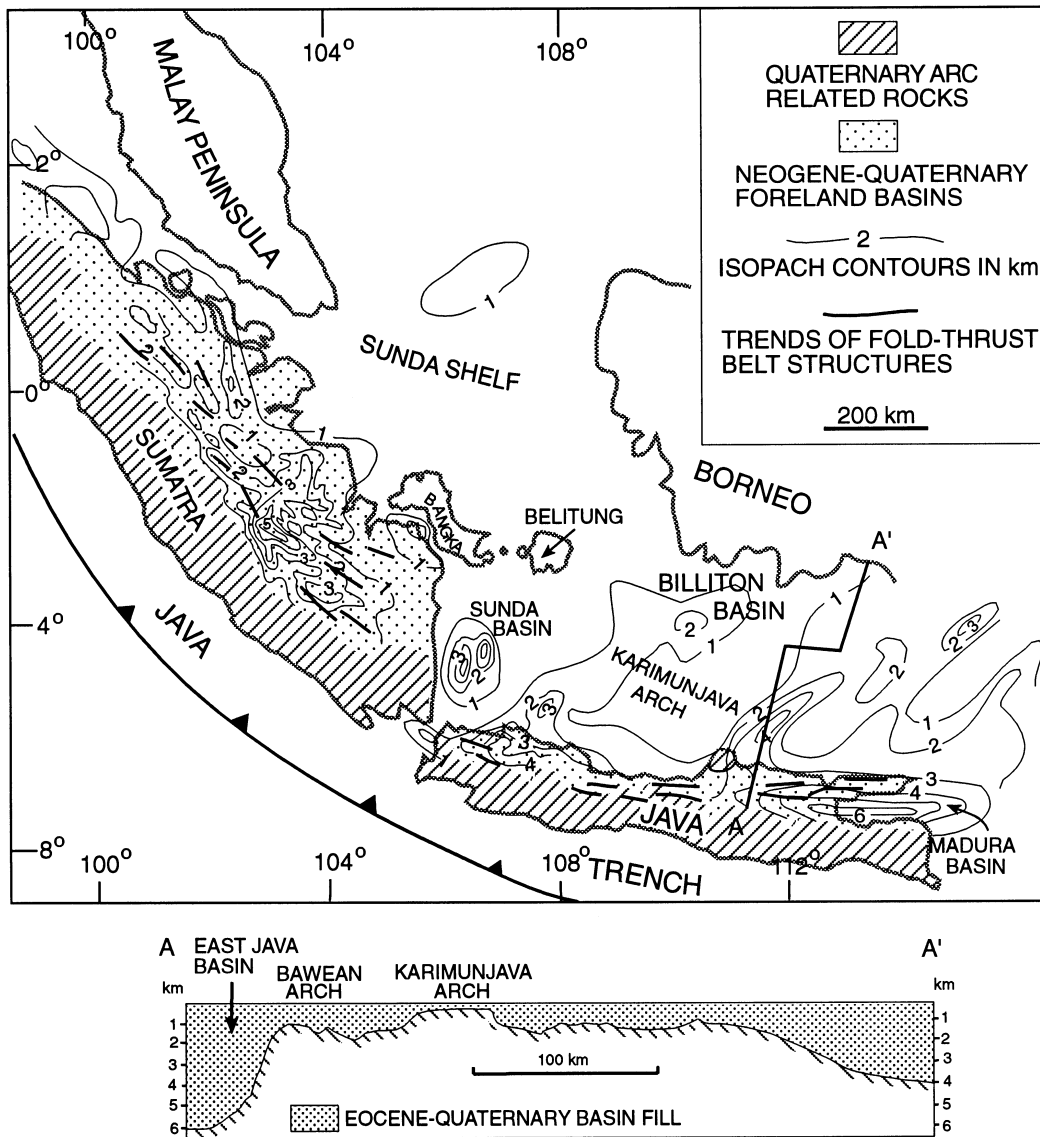


Fig. 2. Generalized map of the Sunda shelf and Indonesian orogenic system, after Ben Avraham & Emery (1973) and Hamilton (1979). A complex, retroarc foreland basin system is present along the north-eastern and northern side of the Sumatra–Java magmatic arc and associated fold-thrust belts. Isopach contours indicate the thickness (in km) of Neogene sediment. In cross-section A–A' note the broad uplifted region of the Bawean and Karimunjawa arches, which separates an obvious foredeep depozone from regions of lesser but broader scale subsidence.

basin systems may be divided into four depozones, which we refer to as wedge-top, foredeep, forebulge and back-bulge depozones (Fig. 1C). Which of these depozones a sediment particle occupies depends on its location *at the time of deposition* (Fig. 5). Boundaries between depozones may shift laterally through time. In some foreland basin systems, the forebulge and back-bulge depozones may be poorly developed or absent. (iii) The longitudinal dimension of the foreland basin system is roughly equal to the length of the adjacent fold-thrust belt. We exclude masses of sediment that spill longitudinally into remnant oceanic basins (e.g. the Bengal and Indus submarine fans) or rifts, because they may not be controlled directly by geodynamic processes related to the orogenic belt. Missing from this definition is any mention of specific

subsidence mechanisms because sediment accommodation in each of the four depozones is controlled by a different set of variables, which we will discuss below. The principal mechanisms of lithospheric perturbation in foreland basin systems are flexure in response to orogenic loading and subsurface loads, but this flexure may be manifested differently in each depozone.

### Wedge-top depozone

In many continental thrust belts, the limit of significant topography is far to the rear of the frontal thrust, and large amounts of synorogenic sediment cover the frontal part of the fold-thrust belt (Fig. 4). This is because frontal thrusts commonly are blind, tipping out in the

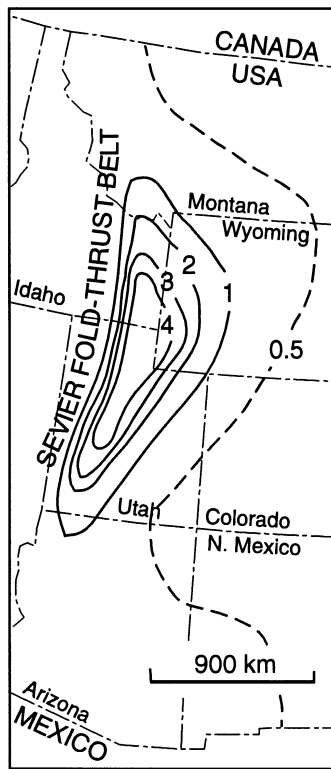


Fig. 3. Isopach map of the upper Albian to Santonian fill of the western interior USA foreland basin system (after Cross, 1986). Note that the basin fill tapers toward both the Sevier thrust belt and the craton.

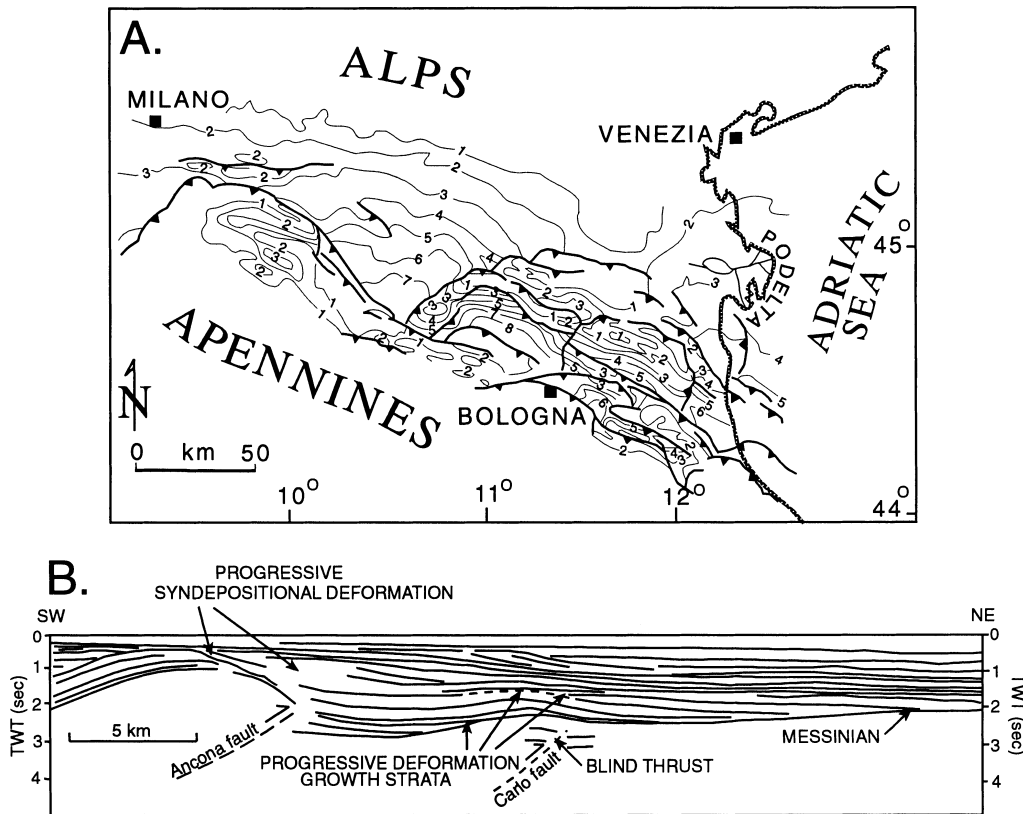
cores of fault propagation anticlines (e.g. Vann *et al.*, 1986; Mitra, 1990; Yeats & Lillie, 1991), triangle zones (e.g. Jones, 1982; Lawton & Trexler, 1991; Sanderson & Spratt, 1992) or passive roof duplexes (Banks & Warburton, 1986; Skuce *et al.*, 1992) in the subsurface, whereas much larger, trailing fault-bend and fault-propagation folds develop above major structural ramps and duplexes further toward the hinterland (Fig. 1C, e.g. Boyer & Elliott, 1982; Pfiffner, 1986; Rankin *et al.*, 1991; Srivastava & Mitra, 1994). In addition, the rocks that are involved in deformation along the fronts of thrust belts are usually relatively young, soft sediments, whereas older, typically more durable rocks are exposed in the hinterland (DeCelles, 1994). The sediment that accumulates on top of the frontal part of the orogenic wedge constitutes the wedge-top depozone (Fig. 1C). Its extent toward the foreland is defined as the limit of deformation associated with the frontal tip of the underlying orogenic wedge. This includes piggyback or thrust-sheet-top (Ori & Friend, 1984) and 'satellite' (Ricci Lucchi, 1986) basins, large feeder canyon fills in the interiors of thrust belts (e.g. Vincent & Elliott, 1995; Coney *et al.*, 1995), deposits associated with local backthrusts and out-of-sequence or synchronous thrusts (Burbank *et al.*, 1992; DeCelles, 1994), and deposits of regionally extensive drainage systems that are antecedent to younger structures and topography toward the foreland (Schmitt & Steidtmann, 1990). In subaerial settings,

these deposits consist of the coarsest material in the basin fill, usually alluvial and fluvial sediments that accumulate proximal to high topographic relief; in subaqueous settings, wedge-top deposits typically consist of mass-flows and fine-grained shelf sediments (e.g. Ori *et al.*, 1986; Baltzer & Purser, 1990).

The wedge-top depozone tapers onto the orogenic wedge, and may be many tens of kilometres in length parallel to the regional tectonic transport direction. Examples abound: Upper Cretaceous to Palaeocene wedge-top sediments are widespread on top of the frontal 75 km of the Sevier thrust belt in Utah and Wyoming (Coogan, 1992; DeCelles, 1994); Eocene–Oligocene wedge-top sediments cover the frontal 30–40 km of the south Pyrenean thrust belt (Puigdefabregas *et al.*, 1986); Pliocene–Quaternary wedge-top sediments bury the frontal 50 km of the active northern Apennines thrust belt (Ricci Lucchi, 1986); the frontal 50 km of the Zagros thrust belt are mantled by Pliocene–Quaternary sediments (British Petroleum, 1956); and  $\approx 100$ –150 km of the active frontal thrust belt in northern Pakistan are covered by young syntectonic sediments (Burbank *et al.*, 1986; Yeats & Lillie, 1991; Pivnik & Johnson, 1995).

The main distinguishing characteristics of wedge-top deposits are the abundance of progressive unconformities (Riba, 1976) and various types of growth structures (Fig. 4B), including folds, faults and progressively rotated cleavages (Anadon *et al.*, 1986; Ori *et al.*, 1986; DeCelles *et al.*, 1987, 1991; Lawton & Trexler, 1991; Suppe *et al.*, 1992; Jordan *et al.*, 1993; Lawton *et al.*, 1993). These features indicate that wedge-top sediment accumulates and is then deformed while at or very near the synorogenic erosional/depositional surface (as opposed to deeply buried and isolated from the surface). The wedge-top depozone actually is part of the orogenic wedge while it is deforming, and hence it is useful for delimiting the kinematic history of the wedge. Aerially extensive aprons of alluvial sediment or shallow shelf deposits commonly drape the upper surface of the orogenic wedge during periods when the wedge is not deforming in its frontal part (Ori *et al.*, 1986; DeCelles & Mitra, 1995), and large, long-lived feeder canyons may develop and fill in the interior parts of orogenic wedges (Vincent & Elliott, 1995; Coney *et al.*, 1995).

The frontal edge of a wedge-top depozone may shift laterally in response to behaviour of the underlying orogenic wedge; thus it may be difficult to distinguish from the proximal foredeep depozone in an ancient foreland basin system. Key distinguishing features of the wedge-top depozone include progressive deformation, numerous local and regional unconformities, regional thinning toward the orogenic wedge and extreme textural and compositional immaturity of the sediment. Sediment derived from the hinterland flanks of frontal anticlinal ridges may be shed back toward the hinterland (e.g. Schmitt & Steidtmann, 1990), and local lacustrine deposits may develop in geomorphically isolated piggyback basins (Lawton *et al.*, 1993). Theoretically, the



**Fig. 4.** (A) Isopach map of post-Messinian sediments and blind thrust faults beneath the Po alluvial plain in northern Italy, an active peripheral foreland basin system (after Pieri, in Bally, 1983). The topographic front of the northern Apennines trends WNW just south of Bologna. Note that the frontal 50+ km of the thrust belt are buried beneath as much as 8 km of wedge-top sediment. The Po delta is prograding eastward into the northern Adriatic Sea, which is the marine part of the system. (B) Interpreted seismic line across a part of the northern Adriatic Sea, off the coast of Conero, Italy, in the Po-Adriatic foreland basin system (after Ori *et al.*, 1986). The topographic front of the Apennines thrust belt is to the left (south-west) of the section. Note the progressive deformation (growth fault-propagation folds) in Pliocene-Quaternary sediments on top of the frontal part of the orogenic wedge. Also note that this part of the basin is submarine, currently receiving shallow-marine sediments.

wedge-top depozone should thicken toward the boundary it shares with the foredeep depozone (Fig. 1C), but post-depositional deformation and cannibalization of wedge-top sediment may obscure this simple concept. In orogenic belts that are not deeply eroded, however, the distinction between wedge-top and foredeep sediment is fairly straightforward (Burbank *et al.*, 1992).

**Foredeep depozone**

The foredeep depozone is the mass of sediment that accumulates between the frontal tip of the orogenic wedge and the forebulge. It consists of the cratonward tapering wedge of sediment that generally has been the focus of most foreland basin studies (cf. Jordan, 1995). Foredeep depozones are typically 100–300 km wide and 2–8 km thick. Voluminous literature documents that subaerial foredeep depozones receive sediment from both longitudinally and transversely flowing fluvial and alluvial deposystems, and subaqueous foredeeps are occupied by shallow lacustrine or marine deposystems that range from deltaic to shallow shelf to turbidite fans. Numerous

studies of peripheral foredeep depozones (e.g. Covey, 1986; Sinclair & Allen, 1992) have documented a transition from early deep-marine sedimentation ('flysch') to later coarse-grained, nonmarine and shallow-marine sedimentation ('molasse'). This transition most likely reflects the fact that peripheral foreland basin systems originate as oceanic trenches and later become shallow marine or nonmarine as continental crust enters the subduction zone. Modern submarine foredeeps on continental crust are characterized by shallow shelf deposits that are accumulating in water generally less than 200 m deep (Figs 2, 4B and 6). Modern foredeep depozones in subaerial foreland basin systems are occupied by fluvial megafans and axial trunk rivers that are fed by tributaries from both the thrust belt and the craton (Räsänen *et al.*, 1992; Sinha & Friend, 1994). Where subaerial foredeeps become submarine along tectonic strike, large deltas often occupy the transition zone (Fig. 4A; Miall, 1981; Baltzer & Purser, 1990).

Foredeep sediment is derived predominantly from the fold-thrust belt, with minor contributions from the forebulge and craton (Schwab, 1986; DeCelles & Hertel,

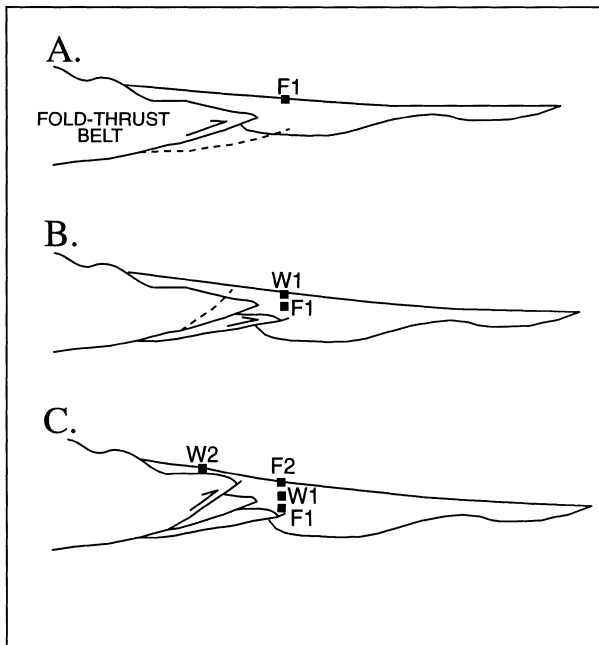


Fig. 5. Schematic diagram showing the deposition and burial of sediment particles (solid squares) in a foreland basin system. Barbed line indicates the zone of active frontal thrust displacement. (A) Particle F1 is deposited and buried in the foredeep depozone. (B) The zone of thrusting steps toward the foreland, incorporating F1 into the hangingwall of the orogenic wedge; particle W1 is deposited and buried in what is now the wedge-top depozone, above F1. (C) The zone of thrusting steps back toward the hinterland (out of sequence); particle F2 is deposited in the foredeep depozone, while particle W2 is deposited in the wedge-top depozone. Particles retain their original depozone signatures, unless they are eroded from the orogenic wedge and reincorporated into the active depositional regime. In this fashion, the boundary between the wedge-top and foredeep depozones shifts progressively toward the foreland through time. Similarly, particles deposited in the forebulge and back-bulge depozones could eventually become buried by foredeep particles.

1989; Critelli & Ingersoll, 1994). Rates of sediment accumulation within the foredeep depozone increase rapidly toward the orogenic wedge (Flemings & Jordan, 1989; Sinclair *et al.*, 1991). Most model studies predict that unconformities should be scarce in the axial part of the foredeep because of the generally high rates of subsidence and sediment supply associated with crustal thickening and orogenic loading (e.g. Flemings & Jordan, 1989; Coakley & Watts, 1991; Sinclair *et al.*, 1991). It must be remembered, however, that the proximal foredeep merges with the distal wedge-top depozone, where sediment bypassing and widespread development of unconformities are common features.

### Forebulge depozone

The forebulge depozone consists of the region of potential flexural uplift along the cratonic side of the foredeep (Fig. 1C). When a foreland basin is modelled as the

flexure of a thin elastic plate floating above a fluid mantle substrate (e.g. Walcott, 1970; Turcotte & Schubert, 1982), the forebulge has a horizontal width of  $\pi\alpha$  (for an infinite plate) or  $3\pi\alpha/4$  (for a broken plate) measured perpendicular to the axis of loading, where  $\alpha$  is defined as the flexural parameter and depends mainly on the flexural rigidity of the lithosphere and the contrast in density between the mantle and the basin fill. For flexural rigidities of 1021 N m to 1024 N m,  $\alpha$  ranges from 26 km to 150 km for density contrasts of 800 kg/m<sup>3</sup>. Thus, the basic flexural equation predicts that the forebulge in a typical flexural basin filled to the crest of the forebulge should be of the order of 60–470 km wide, and a few tens to a few hundred metres high (Fig. 7). Broken plates and plates with lower flexural rigidity should have higher, narrower forebulges than infinite plates and more rigid plates (Turcotte & Schubert, 1982).

In reality, geological forebulges in foreland basin systems have proven difficult to identify unequivocally, particularly in ancient systems (e.g. Crampton & Allen, 1995). One reason for this is that the forebulge is a positive, and potentially migratory, feature, which may be eroded and leave only an unconformity as it passes through a region (Jacobi, 1981). Modelling studies (e.g. Flemings & Jordan, 1989; Coakley & Watts, 1991) have shown that, in basins where sediments derived from the thrust belt prograde cratonward of the region of expected forebulge uplift, the additional load of the sediment interferes with the flexural response to the orogenic load and the forebulge may be buried and morphologically suppressed. In addition, some forebulges may not migrate steadily, instead remaining stationary for long periods and then 'jumping' toward or away from the orogenic belt (e.g. Patton & O'Connor, 1988). If the continental crust in this broad region of potential upward flexure contains pre-existing weaknesses, then local fault-controlled uplifts and depocentres, rather than a smooth flexural profile, may develop (Waschbusch & Royden, 1992). For example, the south-western part of the modern Sunda shelf, which is adjacent to the retroarc foreland basin system of Sumatra and Java, is partitioned by a complex pattern of arches and local depocentres in the region of expected forebulge uplift (Fig. 2). Extensional fault systems have been documented in regions of putative forebulge uplift, both as new crustal features related to tensional stresses and as reactivated older structures (e.g. Hanks, 1979; Quinlan & Beaumont, 1984; Houseknecht, 1986; Tankard, 1986; Wuellner *et al.*, 1986; Bradley & Kidd, 1991).

Because it is an elevated feature, the forebulge generally is considered to be a zone of nondeposition or erosion, and the resulting unconformity has been used to track its position through time (Jacobi, 1981; Mussman & Read, 1986; Stockmal *et al.*, 1986; Tankard, 1986; Patton & O'Connor, 1988; Bosellini, 1989; Flemings & Jordan, 1990; Coakley & Watts, 1991; Sinclair *et al.*, 1991; McCormick & Grotzinger, 1992; Plint *et al.*, 1993; Currie, 1994; Crampton & Allen, 1995). Key features of uncon-

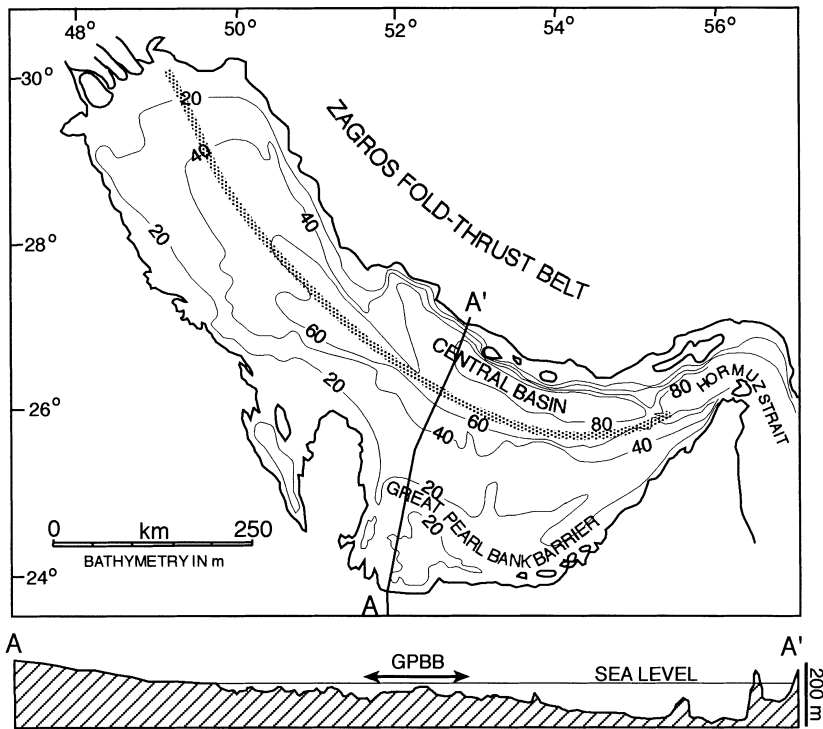


Fig. 6. Bathymetric map and cross-section of the Persian Gulf, which is the shallow-marine part of the peripheral foreland basin system along the south-west side of the Zagros collisional orogenic belt. The shaded line on the map shows the approximate front of the Zagros fold-thrust belt and the modern wedge-top depozone. The Great Pearl Bank Barrier (GPBB in cross-section) is a broad carbonate shoal that may be related to forebulge uplift. After Kassler (1973).

formities produced by migration of forebulges include progressive onlap in a cratonward direction by foredeep strata onto the unconformity, a cratonward increasing stratigraphic gap on the foredeep side of the forebulge, and regional, low-angle ( $\ll 1^\circ$ ) bevelling of a maximum few hundred metres of the pre-existing stratigraphic section (Plint *et al.*, 1993; Crampton & Allen, 1995).

The existence of foreland basin systems in which sediment progrades into the forebulge region indicates that this zone also may be the site of substantial sediment accumulation. For example, Holt & Stern (1994) suggested that  $\approx 400$  m of Oligocene–Miocene, shallow-marine sediment accumulated in the forebulge area of the Taranaki foreland basin in New Zealand (Fig. 8). Eocene–Quaternary sediments derived from the Sumatra–Java retroarc fold-thrust belts and magmatic arc extend several hundred kilometres north-eastward of the limit of major foredeep subsidence (Fig. 2; Ben Avraham & Emery, 1973; Hamilton, 1979). Lower Cretaceous nonmarine sedimentary rocks in the western interior USA foreland basin in Montana (Fig. 9A) and Utah (Fig. 9B) show pronounced thinning (but not complete disappearance) along linear zones,  $\approx 50$ – $100$  km wide, parallel to the front of the Sevier thrust belt, suggesting the presence of forebulges that were overlapped by synorogenic fluvial sediment. Sediment derived from the modern Subandean fold-thrust belt in Peru and Bolivia is being deposited into regions far beyond the zone of major foredeep subsidence, suggesting the presence of an overtopped forebulge (Jordan, 1995).

In submarine foreland basin systems in which the foredeep depozone is not filled up to the crest of the

forebulge, local carbonate platforms may develop in the forebulge depozone (Wuellner *et al.*, 1986; Patton & O'Connor, 1988; Allen *et al.*, 1991; Dorobek, 1995). Extensive forebulge carbonate platforms and ramps can connect the foredeep depozone with the back-bulge depozone (see below) and the craton (Bradley & Kusky, 1986; Pigram *et al.*, 1989; Reid & Dorobek, 1993; Giles & Dickinson, 1995). The Great Pearl Bank Barrier along the south-west side of the Persian Gulf, for example, is a region of widespread, almost pure carbonate accumulation (Wagner & van der Togt, 1973). Water depth above the barrier is  $\approx 10$  m, and increases toward both the Zagros foredeep (maximum depth of  $\approx 80$  m) and the Arabian landmass (Fig. 6), suggesting the morphology of a subdued forebulge. In ancient foreland basin systems, these large carbonate deposits typically are not considered part of the foreland basin system, but their stratal geometries clearly are influenced by the flexural processes associated with the fold-thrust belt and may provide sensitive indicators of regional subsidence history (Grotzinger & McCormick, 1988; Dorobek, 1995).

In subaerial foreland basin systems in which the foredeep is not filled to the crest of the forebulge, the forebulge region should be a zone of erosion, with streams draining both toward and away from the orogenic belt (Flemings & Jordan, 1989; Crampton & Allen, 1995). If thrust-belt-derived sediment progrades into the forebulge depozone, a relatively thin flap of highly condensed (relative to the foredeep depozone) fluvial and aeolian sediment is deposited over a broad region (Fig. 9). These deposits usually are portrayed as distal foredeep sediments, but they are markedly different from typical foredeep deposits in terms of their regionally consistent



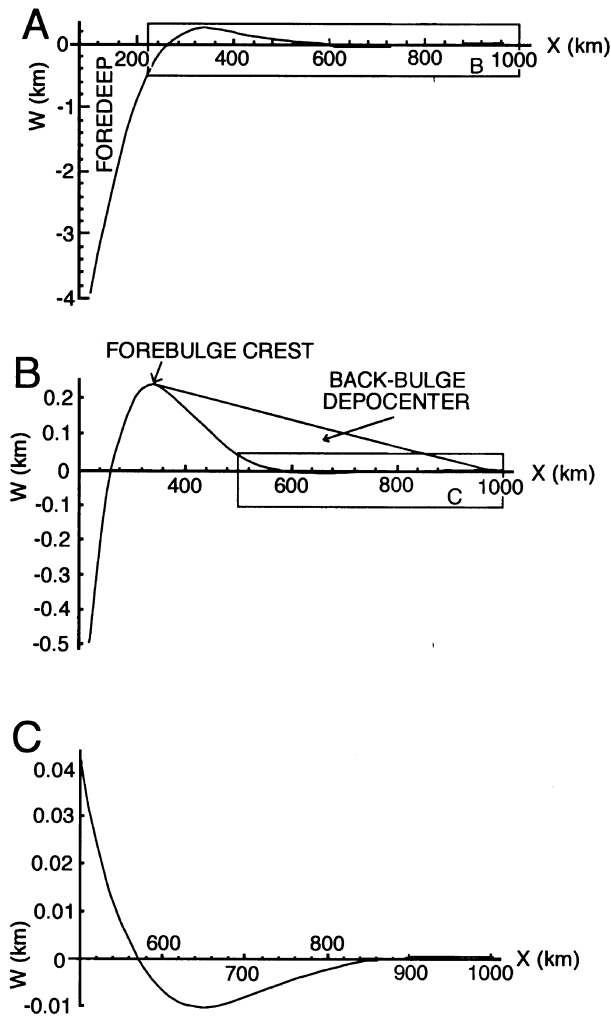


Fig. 7. (A) Flexural profile for a thin, infinite, elastic plate floating on a fluid substrate, subjected to a rectangular topographic load 100 km wide and 2.5 km high located on its left side, but not shown, and a load of sediment that fills the foredeep to the level of the forebulge. The density of the load is  $2650 \text{ kg m}^{-3}$  and density of the basin fill is  $2400 \text{ kg m}^{-3}$ . The resulting flexural parameter ( $\alpha$ ) is 97.6 km. The area outlined by the box is magnified in (B). (B) Magnified plot of the forebulge and back-bulge areas shown in (A). Note the change in vertical scale. Sloping line from forebulge crest represents a linear approximation of an exponential, aggradational profile connecting the thrust belt and the undeformed foreland. Aggradation up to this profile would produce more than 100 m of accumulation. Area outlined in box is shown in further detail in (C).

and minor thicknesses, lithofacies (distal fluvial and aeolian), abundance of well-developed palaeosols, relatively low subsidence rates and parallel internal time planes (DeCelles & Burden, 1992).

### Back-bulge depozone

The back-bulge depozone constitutes the sediment that accumulates between the forebulge depozone and the craton (Fig. 1C). Although the bulk of this sediment is

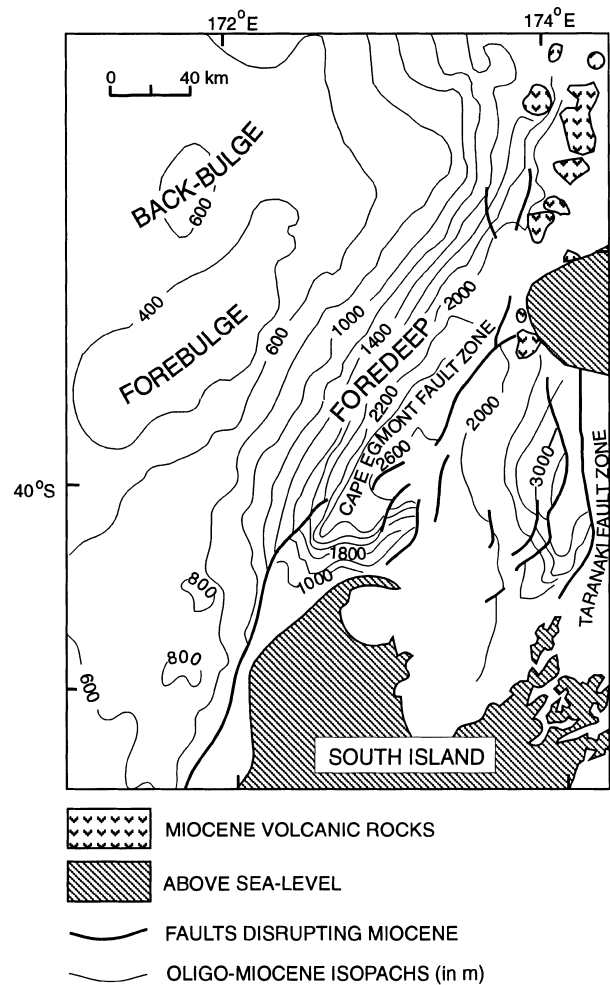


Fig. 8. Isopach map of the Taranaki retroarc foreland basin in New Zealand, after Holt & Stern (1994). Note the well-developed, Oligo-Miocene foredeep, forebulge and back-bulge depozones.

derived from the orogenic belt, contributions from the craton and development of carbonate platforms may be significant in submarine systems. Flemings & Jordan (1989) referred to this depozone as an 'outer secondary basin', and examples have been documented in the Taranaki foreland basin (Holt & Stern, 1994; Fig. 8), the Sunda shelf region (Ben Avraham & Emery, 1973; Fig. 2) and the Cordilleran foreland basin (DeCelles & Burden, 1992; Plint *et al.*, 1993; Fig. 9). A key aspect of these back-bulge accumulations is that isopach patterns show regional closure around a central thick zone, which suggests that sediment accommodation may involve some component of flexural subsidence cratonward of the forebulge (Fig. 7). The back-bulge depozone also has been referred to as the overfilled part of the basin fill (e.g. Flemings & Jordan, 1989; DeCelles & Burden, 1992), but this usage is undesirable because it presents a spatial contradiction: how can the overfilled part of a basin be part of the basin if it, by definition, extends beyond the basin?

Because of the relatively low rates of subsidence in the

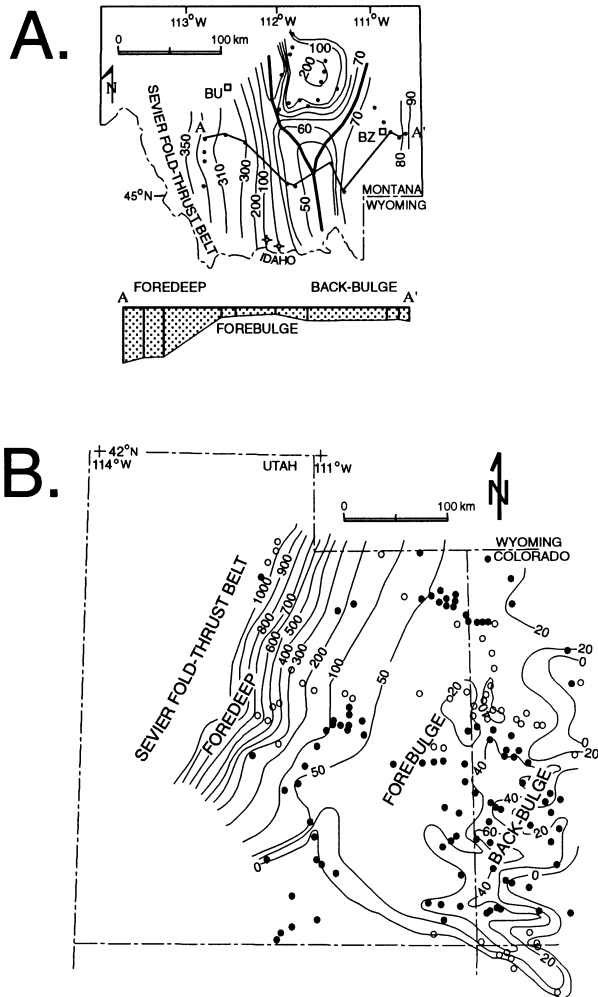


Fig. 9. Isopach maps (contours in metres) showing proposed foredeep, forebulge and back-bulge depozones in Lower Cretaceous fluvial and lacustrine deposits associated with the eastward-vergent, retroarc Sevier fold-thrust belt in the western interior foreland basin of the United States. (A) The Kootenai Formation in south-western Montana, USA, which exhibits evidence for a bifurcated forebulge axis (bold Y-shaped lines) and a broad back-bulge depozone (data from DeCelles, 1984). (B) The Cedar Mountain Formation in east-central Utah, USA (after Currie, 1994); zero-thickness contour is a result of erosion, not depositional pinch-out. Note the region of irregular thickening (before erosional truncation) in the back-bulge region. Solid circles represent surface sections from Currie (1994) and Craig (1955); open circles represent well data from Currie (1994), Craig (1955) and Sprinkel (1994).

back-bulge depozone, stratigraphic units are much thinner than those in the foredeep depozone and time planes are subparallel over lateral distances of several hundred kilometres perpendicular to the orogenic belt (Flemings & Jordan, 1989; DeCelles & Burden, 1992). Depositional systems in the back-bulge depozone are dominantly shallow marine (<200 m bathymetry; Fig. 2; Ben Avraham & Emery, 1973; Holt & Stern, 1994) and nonmarine, and the calibre of sediment is generally fine because of the large distance from its principal source in the orogenic belt. Local accumulations of coarse-grained

sediment may be present on the flank of the uplifted forebulge area (Giles & Dickinson, 1995).

## SUBSIDENCE AND SEDIMENT ACCOMMODATION

### Flexure due to topographic loads

Because they are by definition associated with fold-thrust belts, all foreland basins are affected by the 'topographic loads' of their adjacent thrust belts, which typically produce flexural responses over lateral distances of several hundred kilometres in the foreland plate (Fig. 10; e.g.

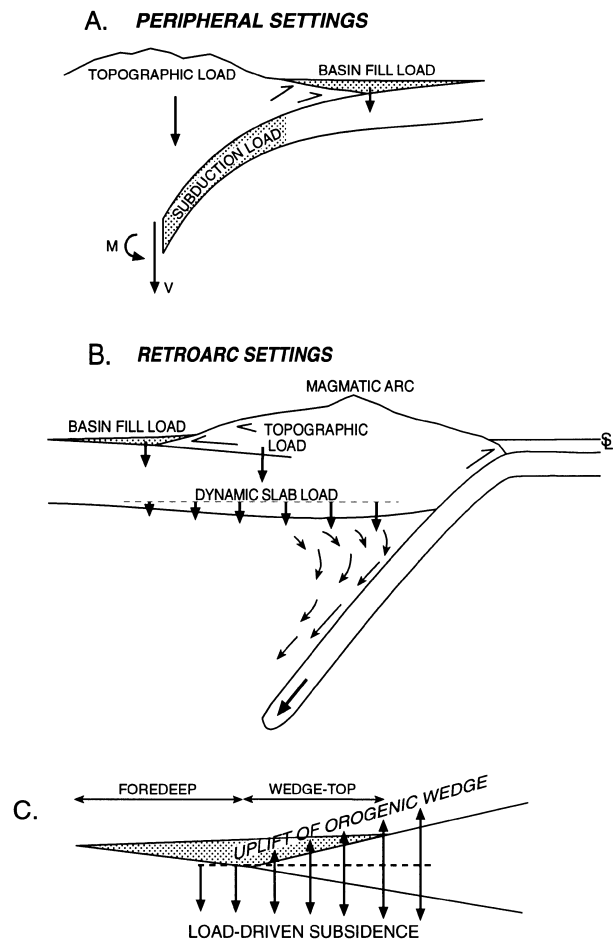


Fig. 10. (A) Schematic diagram showing the principal loads in peripheral foreland basin systems. In addition to the topographic and sediment loads, a subduction load, due to a vertical shear force ( $V$ ) and bending moment ( $M$ ) on the end of the subducted slab, may exist at depths of 50–200 km (Royden, 1993). (B) Retroarc foreland basin systems involve topographic and sediment loads as well as a dynamic slab load caused by viscous coupling between the subducting slab, overlying mantle-wedge material and the base of the overriding continental plate (Mitrovica *et al.*, 1989; Gurnis, 1992). (C) Accumulation in the wedge-top depozone takes place under competing influences of regional, load-driven subsidence (downward pointing arrows), and local uplift of the orogenic wedge in response to shortening and thickening (upward pointing arrows).

Price, 1973; Beaumont, 1981; Jordan, 1981). The primary flexural responses to the topographic load are a deep flexural trough (the foredeep depozone), the forebulge and a zone of extremely minor ( $\approx 10$  m for typical flexural rigidities) flexural subsidence in the backbulge region (Fig. 7). Modelling studies suggest that sediment and water loads should alter this primary flexural response by suppressing the forebulge and spreading the flexure further onto the craton (e.g. Flemings & Jordan, 1989). Rates of flexural subsidence may also overwhelm rates of uplift in the frontal part of the orogenic wedge, producing accommodation for wedge-top sediment (Fig. 10C).

### Flexure due to subduction loads

Many foredeep depozones exhibit subsidence that is greater and/or more widespread than expected from the observable topographic load and the sediments and water that occupy the basin (Karner & Watts, 1983; Royden & Karner, 1984; Cross, 1986; Royden, 1993). The Po–Adriatic foredeep depozone, for example, is three to four times deeper than expected from the mass of the Apennine fold-thrust belt; the likely cause of most of the flexural subsidence in the foreland is the downward pull of a dense subducted oceanic slab 50–150 km beneath the Apennines (Royden, 1993). Because of its position on the subducting plate, any peripheral foreland basin system may be subject to a ‘subduction load’ of oceanic lithosphere (Fig. 10A); however, with continental collision and continued partial subduction of transitional or continental lithosphere, the effect of the subduction load will be lessened and topographic loading will dominate the net subsidence profile (Karner & Watts, 1983; Royden, 1993).

### Flexure due to dynamic subducted slabs

In contrast to peripheral foreland basins, retroarc forelands are situated on continental plates above subducting slabs, and may be influenced by anomalous, far-field subsidence related to the presence of the slabs (e.g. Cross, 1986). Mitrovica *et al.* (1989) and Gurnis (1992) have shown that subducted oceanic slabs can cause rapid, long-wavelength (more than 1000 km from the trench) subsidence and uplift of the order of a kilometre or so on the overlying continental plate. This ‘dynamic slab-driven’ effect results from viscous coupling between the base of the continental plate and downward circulating mantle-wedge material that is entrained by the subducting slab (Fig. 10B). Because this effect operates over long wavelengths, it can explain anomalously widespread shallow-marine and nonmarine sediment accumulations in cratonal areas far inboard of the main flexural depression due to the orogenic thrust wedge (e.g. Mitrovica *et al.*, 1989; Lawton, 1994). A modern example of a retroarc foreland region that may be experiencing both short-wavelength, thrust-driven subsidence and longer-wavelength, dynamic slab-driven subsidence is the Sunda

shelf region north-east of Java and Sumatra, where shallow seas cover a vast area of continental crust lying inboard of active retroarc fold-thrust belts and thick (3–6 km) Neogene foredeep depozones on both islands (Fig. 2; Hamilton, 1979). Gurnis (1992, 1993) suggested that the widespread shallow-marine, flooded continental crust in the northern and western Pacific basin is a result of dynamic slab-driven subsidence, and several authors (Holt & Stern, 1994; Coakley & Gurnis, 1995; Lawton, 1994; Pang & Nummedal, 1995) have recently explained anomalous, relatively uniform subsidence in distal parts of foreland basin systems as the result of a similar process.

### Interference of flexural responses

The flexural responses to topographic, subduction and dynamic slab loads operate over different wavelengths and can therefore interfere either constructively or destructively. For example, Royden (1993) demonstrated that subsidence in the foreland basin systems associated with the western European Alps is the net result of interference between topographic- and subduction-load-driven profiles. The depths of the Alpine foredeep depozones are shallower than predicted from the observable topographic load, probably because of interference between forebulge uplift associated with a subduction load and foredeep subsidence associated with the topographic load of the Alps. Similarly, the long-wavelength flexural response to a dynamic-slab load may interfere with the shorter-wavelength flexure caused by a topographic load. Lawton (1994) suggested that the Tin Islands (Bangka and Belitung, Fig. 2), offshore Sumatra, are a forebulge due to the dynamic-slab load of the subducting Indian plate. The Sumatran back-arc region is characterized by low-amplitude open folds with minor crustal shortening and thickening (Hamilton, 1979), and the regional topographic load, which is generally less than  $\approx 2$  km elevation, is insufficient to explain the thick Neogene retroarc foredeep (locally 3–6 km). In addition, the Sunda shelf is a broad, shallow-marine, continental shelf that probably owes its submergence to the presence of subducting slabs (Gurnis, 1992). Thus, it is plausible that flexural subsidence in the Sumatran retroarc region is due to combined topographic and dynamic-slab loads.

### Sediment accommodation

Sediment accommodation in foreland basin systems is controlled primarily by flexural subsidence, as discussed above. Local structural damming and base-level or eustatic sea-level variations also contribute to the accommodation signal recorded by strata in foreland basin systems. Each depozone should be characterized by its own peculiar subsidence and accommodation patterns, because each responds differently to a given process in the orogenic belt and subduction system.

Sediment accumulation in the wedge-top depozone is the net result of competition between regional, load-

driven subsidence, and regional and local uplift of the orogenic wedge owing to crustal thickening or isostatic rebound (Fig. 10C). In addition, local accumulations of wedge-top sediment may result from structural damming by uplift of anticlinal ridges in the frontal foothills of the fold-thrust belt (Lawton & Trexler, 1991; Talling *et al.*, 1995). In wedge-top depozones that are marginal to marine seaways, changes in eustatic sea-level also may play an important role in development and destruction of sediment accommodation. In general, periods of widespread shortening and uplift in the orogenic wedge are marked in the wedge-top depozone by syndepositional, thrust-related deformation and development of unconformities owing to erosion and sediment bypassing to the foredeep depozone. Periods during which the frontal part of the orogenic wedge is not shortening are marked in the wedge-top depozone by continued unconformity development and subsequent regional onlap of sediment that is not syndepositionally deformed (DeCelles, 1994).

As demonstrated by numerous previous studies, sediment accommodation in foredeep depozones is primarily a response to loading by the adjacent orogenic wedge and sediment eroded from the wedge (e.g. Price, 1973; Beaumont, 1981; Jordan, 1981; and many others more recently), as well as subsurface loading (Mitrovica *et al.*, 1989; Royden, 1993). Foredeep depozones also may be affected by regional isostatic uplift during erosion of the orogenic load and by uplift associated with advance of the orogenic thrust wedge or retrograde migration of the forebulge (Quinlan & Beaumont, 1984; Heller *et al.*, 1988; Flemings & Jordan, 1990; Sinclair *et al.*, 1991). Changes in relative sea level can cause increased or decreased sediment accommodation in foredeep depozones. For example, the entire Persian Gulf foredeep depozone, which currently is the site of active deposition (Fig. 6), was subaerially exposed and incised by fluvial channels during the Pleistocene lowstand (Kassler, 1973).

Forebulge depozones are commonly areas of subaerial exposure and erosion (Crampton & Allen, 1995), but a number of modern and ancient foreland basin systems contain forebulges that are buried by synorogenic sediment (Figs 2, 8 and 9). It is important to note that these are situations in which isopach patterns indicate that, contrary to results of recent modelling studies (e.g. Flemings & Jordan, 1989), the structural forebulge exists in spite of being buried by sediment derived from the thrust belt. Thus, the geomorphological manifestation of a forebulge may be absent or subdued, even if it is structurally and/or stratigraphically well defined. The possible causes of sediment accumulation in the forebulge depozone include regional, long-wavelength subsidence due to a dynamic slab (in retroarc systems), and aggradation up to base level or an equilibrium drainage profile that crosses the crest of the forebulge. In the modern Persian Gulf, sediment is accumulating in the region of predicted forebulge uplift along the Great Pearl Bank Barrier (Fig. 6; Purser, 1973), mainly because sea level is relatively high at present.

The mechanisms of sediment accumulation and preservation in back-bulge depozones are poorly understood. Simple flexural theory predicts the presence of a broad region, approximately the same width as the forebulge depozone, of extremely minor (of the order of 10 m for typical flexural rigidities) subsidence in the back-bulge depozone (Fig. 7B,C). Modelling by Flemings & Jordan (1989), however, suggests that when sediment progrades cratonward of the foredeep, the resulting flexural profile in regions beyond the foredeep should be nearly planar. The presence of thrust-belt-derived sediment in regions cratonward of the foredeep ('overfilled' basins) does not necessarily indicate the existence of a region of secondary flexural subsidence, because sediments may simply spill cratonward of the foredeep depozone. Nevertheless, discrete back-bulge depozones that are bounded by prominent forebulges have been amply documented in both marine and nonmarine foreland basin systems, and contain accumulations of sediment that range in thickness from a few tens of metres to more than 600 m (e.g. Figures 8 and 9; Quinlan & Beaumont, 1984; DeCelles & Burden, 1992; Plint *et al.*, 1993; Currie, 1994; Holt & Stern, 1994; Giles & Dickinson, 1995). These thicknesses are an order of magnitude greater than what would be expected if accommodation resulted from flexural subsidence alone (Fig. 7).

Plausible mechanisms for such thick back-bulge accumulations include regional, long-wavelength subsidence due to dynamic slabs (Mitrovica *et al.*, 1989; Lawton, 1994) and aggradation up to an equilibrium drainage profile (Leopold & Bull, 1979) or to base level (in subaqueous settings). As an example of the former, the forebulge in the flexural profile shown in Fig. 7 is  $\approx 230$  m high. An equilibrium depositional surface extending from the front of a 2.5-km-high thrust belt to the craton and tangent to the forebulge would accommodate well over 100 m of sediment in the axis of the back-bulge depozone, equal to  $\approx 10\%$  of the volume of the foredeep depozone (Fig. 7B). If the equilibrium profile at the crest of the forebulge rested on forebulge sediment, instead of basement, or if both the back-bulge and forebulge were submarine, the potential accommodation in the back-bulge depozone would be much greater. Addition of this sediment to the back-bulge depozone would drive further, albeit minor, subsidence. Thus flexural subsidence (due to topographic and subduction loads) is the main control on accommodation in foredeep depozones, but the elevation of the forebulge, relative sea level and availability of sediment may be more important in the back-bulge depozone. In retroarc foreland basin systems, regional platform subsidence due to dynamic subducted slabs may add significantly to the available accommodation in back-bulge depozones (Fig. 2; Holt & Stern, 1994; Lawton, 1994).

## DISCUSSION

Foreland basin systems are complex, large-scale features that respond to interacting complexes of variables

(Flemings & Jordan, 1989; Jordan & Flemings, 1991). Changes in variables may affect different depozones in different ways, so the concept of a foreland basin as a single depozone can lead to erroneous interpretations of the stratigraphic record. Our hope is that the expanded concept of a foreland basin will encourage workers to characterize explicitly the various depozones in terms of geometry, depositional systems and palaeogeography, sediment composition, structure, sequence stratigraphy, and subsidence patterns. Integration of these characteristics throughout the entire foreland basin system with available information about the overall tectonic setting (retroarc vs. peripheral) of the system should help to clear up some apparent ambiguities that have resulted from conceptualizing foreland basins according to the previous definition.

A key point in the expanded definition for foreland basin systems is that a depozone is defined in terms of its position during deposition, rather than its eventual position with respect to the thrust belt (Fig. 5). Once a particle of sediment is buried and incorporated into the long-term sedimentary record, its depozone cannot change. This is an important distinction to make because using depositional facies to understand the history of a thrust belt depends on the interaction of tectonics and syndepositional stratigraphic architecture, not post-depositional architecture that has been modified by thrust-related deformation. For example, a sediment particle that is deposited in a foredeep depozone arrives at its site of deposition under the influence of processes peculiar to the foredeep. Although this particle subsequently may be incorporated into the orogenic wedge by frontal imbrication, this does not transform the particle into a part of the wedge-top depozone (Figs 5A,B). Rather, it becomes part of the active orogenic wedge with respect to all contemporaneously mobile sediment particles, and no longer is part of an active depozone. It remains part of a now ancient foredeep depozone, unless it is cannibalized and reincorporated into the active depositional regime. A wedge-top particle that is deposited above the original foredeep particle remains a wedge-top particle, even if out-of-sequence thrusting causes the front of the thrust belt to migrate back toward the hinterland (Fig. 5C). In three dimensions, boundaries between the various depozones should be broadly irregular, intertonguing interfaces that may be difficult to locate precisely because of later erosion and burial. Because of these complexities, it is best to identify a given mass of sediment in terms of depozone type by placing it in the context of an incrementally restorable geological cross-section that incorporates all available constraints on timing and spatial distribution of deformation with respect to foreland sedimentation.

Distinguishing the true depozone(s) of sediments in a given stratigraphic section is critical for interpreting subsidence histories of foreland basin systems. For example, sediment accumulation (and/or subsidence) rate curves from many foreland basin systems exhibit a

characteristic, long-term, sigmoidal shape, with initially slow accumulation followed by a period of rapid accumulation, in turn followed by a return to slower rates (Cross, 1986; Johnson *et al.*, 1986). This pattern could be interpreted as a response to changes in foredeep subsidence rates related to thrust-displacement events, but in some cases (DeCelles, 1994) it can be directly correlated with a change from distal foredeep, to proximal foredeep, to wedgetop depozones at a given locale, a process that takes place as the orogenic wedge propagates forward. A similar problem results if back-bulge deposits are not distinguished from foredeep deposits. The Upper Jurassic Morrison Formation in the western interior USA may be an example of this problem. Although the Morrison thickens three-fold from central Wyoming to north-eastern Utah, it does not exhibit the rapid thickening of a typical foredeep deposit (Currie, 1994). Heller *et al.* (1986) and Yingling & Heller (1992) interpreted the lack of abrupt westward thickening in the Morrison as evidence that thrusting in the Sevier orogenic belt did not commence until after Late Jurassic time. On the other hand, Currie (1994) has shown that Morrison thickness patterns can be reconciled with deposition in a back-bulge depozone. Goebel (1991) showed that the failure to recognize back-bulge deposits in the Frasnian–Famnenian Pilot Shale as part of the Antler foreland basin system in Nevada has led previous workers to suggest that the Antler orogeny did not begin until Early Middle Mississippian time. Many thrust belts (particularly peripheral thrust belts) exhibit horizontal displacements that are comparable to typical flexural wavelengths in their adjacent foreland basin systems (> 150 km), which suggests that long-term migration of depozones in the direction of thrust-belt propagation should stack depozones vertically in the stratigraphic record. Changes in the slopes of long-term subsidence curves from foreland basin systems thus probably represent temporal changes in depozone type at a given locale (e.g. Johnson *et al.*, 1986).

Explicit inclusion of the wedge-top depozone in a foreland basin system helps to explain apparent contradictions that have arisen from modelling and field-based studies of foreland basin stratigraphy. For example, when foreland basins are modelled as wedge-shaped (in transverse cross-section) prisms of accommodation created by loading along one side of the basin, a two-phase pattern of basin filling results (e.g. Beck *et al.*, 1988; Blair & Bilodeau, 1988; Heller *et al.*, 1988). During periods of crustal shortening and orogenic loading, increased rates of subsidence trap coarse-grained sediment in areas directly adjacent to the orogenic load, whereas fine-grained material is deposited throughout the major part of the basin. Erosional reduction of the orogenic load during periods of tectonic quiescence causes flexural rebound and reduced accommodation in the proximal foreland basin, which in turn allows coarse-grained material to prograde into the distal part of the basin. Episodes of crustal shortening and orogenic growth are thus

correlated with periods of fine-grained sedimentation throughout most of the foreland basin, and periods of tectonic quiescence are correlated with influxes of coarse-grained sediment into the foreland basin ('subsidence-driven progradation' of Paola *et al.*, 1992). The two-phase model has been used in numerous recent papers to explain formation-scale alternations in lithofacies and regional thickness patterns (e.g. Heller & Paola, 1989; in addition to many others). A potential problem with this model for foreland basin systems is that thrust loading may not be as episodic as commonly perceived from the vantage point of the foreland. Several recent studies of long-term thrust-belt kinematic histories (e.g. Burbank *et al.*, 1992; Jordan *et al.*, 1993; DeCelles, 1994) have shown that orogenic shortening and loading are probably more continuous in time, as might be predicted from the general steadiness of plate convergence velocities and from critical taper theories of orogenic wedges (Chapple, 1978; Davis *et al.*, 1983; Stockmal, 1983). Whereas thrusting in the frontal part of the orogenic wedge may indeed be episodic, frontal thrusts are characteristically very thin-skinned, often blind, and therefore have little effect on the overall distribution of loading. Significant crustal thickening is concentrated above large crustal ramps in the hinterland. In addition, the 'quiescent' periods between frontal thrusting events are often marked by taper-restoring events such as out-of-sequence or synchronous thrusting, backthrusting, and, especially in the hinterland part of the wedge, penetrative internal shortening. Thus, the idea that kinks in subsidence history curves represent individual thrusting events may be somewhat misleading.

Another problem with the simple two-phase model was raised by Damanti's (1993) analysis of depositional systems in the actively subsiding Bermejo foreland basin of western Argentina. He showed that the major control on distribution of coarse sediment in the foreland basin is the size distribution of catchment basins in the adjacent thrust belt. Large catchments provide high-volume sediment fluxes that distribute coarse sediment across much of the foreland basin, whereas small catchments produce minor alluvial fans that coalesce laterally into a narrow belt of coarse sediment along the topographic front of the thrust belt. Both types of depositional systems coexist under the same tectonic and climatic conditions. Large antecedent drainages in thrust belts (DeCelles, 1988; Schmitt & Steidtmann, 1990) may therefore overwhelm even rapidly subsiding foredeep depozones ('flux-driven' progradation of Paola *et al.*, 1992).

The predictions of the two-phase model for foreland basin stratigraphy also conflict with evidence from the wedge-top depozones of foreland basin systems which indicates that, during periods of thrust displacement in the frontal thrust belt, the wedge-top depozone is uplifted, deformed and at least partially eroded (e.g. Wiltschko & Dorr, 1983; Burbank *et al.*, 1988; DeCelles, 1994). Sediment thereby produced must be transported into the foredeep depozone. Whether this sediment is

coarse or fine depends upon a number of independent factors (e.g. source rock type, climate and amount of penetrative strain in the source rocks), but the key point is that, during thrusting, sediment must be supplied to the more distal part of the foreland basin system because no accommodation exists in the wedgetop. The two-phase model for foreland basins is inappropriate because the original basin geometry is not realistic; the 'doorstop' wedge geometry used in these models is more appropriate for a half-graben than a foreland basin. In fact, the two-phase model works well in half-graben settings (e.g. Leeder & Gawthorpe, 1987; Mack & Seager, 1990). A better way to parameterize the transverse geometry of a foreland basin system would be as a somewhat asymmetrical prism that tapers both toward and away from the orogenic belt (e.g. Beaumont *et al.*, 1992), with subsidence increasing away from the orogen to a maximum along the proximal edge of the foredeep depozone, and decreasing from there toward the craton (Fig. 10C). As expected, basins modelled like this display in-phase deformation and sediment progradation on their 'orogenic' sides (Paola *et al.*, 1992, fig. 7).

A final example of how the presently accepted concept of a foreland basin leads to potentially erroneous or oversimplified interpretations is in the application of sequence-stratigraphic models to foreland basin fills. Because subsidence is presumed to increase continuously toward the orogen, the proximal part of the foreland basin ('Zone A' of Posamentier & Allen, 1993) is considered to be an area in which the rate of subsidence always outpaces the rate of falling eustatic sea level. The result is an absence of type 1 unconformities on the proximal side of the basin (e.g. Swift *et al.*, 1987; Jordan, 1995). Tectonic events, which drive increased rates of subsidence, are thought to generate transgressive surfaces, highstand systems tracts and nearly continuous deltaic and coastal plain deposition in the part of the basin closest to the thrust belt. Again, this model artificially truncates the foreland basin system on its thrust-belt side and does not account for the presence of regional unconformities in wedge-top depozones that are fully integrated with foredeep depozones. The examples most often cited in the sequence stratigraphy paradigm for foreland basins are the superbly exposed, Albian-Palaeocene fluvial and deltaic deposits of east-central Utah (e.g. Swift *et al.*, 1987). These deposits are generally within the foredeep depozone of the western interior foreland-basin system. The late Cenomanian to early Campanian part of the succession lacks major unconformities and consists mainly of distal deltaic and offshore marine facies (e.g. Molenaar & Cobban, 1991). In contrast, overlying middle Campanian to Palaeocene strata are dominated by sandy fluvial facies and contain major unconformities of Campanian and Maastrichtian-Palaeocene age (Fouch *et al.*, 1983). The same unconformities appear to be traceable westward into the wedge-top part of the system in central Utah (Lawton, 1982, 1985; DeCelles *et al.*, 1995). This pattern can be explained as a transition from

the distal foredeep depozone (late Cenomanian to early Campanian) to the proximal foredeep and distal wedge-top depozones (middle Campanian to Palaeocene). The 'missing' unconformities exist; they just are not present in the distal foredeep depozone (Gardner, 1995). Subsidence does not increase continuously toward the proximal side of the foreland basin system, and concepts of sequence stratigraphy developed on passive margins are applicable to foreland basins, albeit with more attention given to the role of tectonic subsidence.

## CONCLUSIONS

Foreland basin systems comprise four depozones that result from the primary flexural response to topographic loading: wedge-top, foredeep, forebulge and back-bulge. Superimposed on the primary flexural response is flexure due to subducted slab-pull (in peripheral forelands; Royden, 1993) or flexure due to regional viscous coupling of the overriding plate with circulating mantle-wedge material (in retroarc forelands; Gurnis, 1992). Each depozone has a peculiar pattern of subsidence and uplift in response to tectonic driving forces related to the adjacent orogen and subduction system and potential interference of flexural responses of differing wavelengths.

The internal architecture, sedimentology and structure of each depozone are distinctive. The wedge-top comprises all sediment that accumulates above the active orogenic thrust wedge, and is characterized by textural immaturity (especially in nonmarine systems), progressive deformation, widespread tectonic unconformities and regional thinning toward the orogenic belt. Explicit inclusion of the wedge-top depozone in the definition of a foreland basin system requires that basin models be geometrically parameterized (in transverse cross-sections) with a doubly tapered prismatic shape, rather than a wedge shape. The distal part of the wedge-top merges with the proximal part of the foredeep depozone. The foredeep depozone constitutes the thickest part of the foreland basin system; it characteristically thickens abruptly toward the thrust belt, has spatially divergent chronozones, and is deposited by a variety of distal fluvial, lacustrine, deltaic and marine depositional systems. Although foredeep depozones are relatively well understood, they should be considered in the context of adjacent wedge-top and forebulge depozones in order to provide unambiguous information about orogenic processes.

Forebulge and back-bulge depozones have received limited treatment in the literature, in spite of the fact that they can provide important information about orogenic processes. Because neither forebulge nor back-bulge depozones exhibit the features that are routinely used to associate the foredeep depozone with tectonic processes in the orogenic belt, they commonly are overlooked as indicators of thrust-belt activity. Both depozones are typically thin relative to the distal part of the wedge-top and the foredeep depozones. Regional uncon-

formities, well-developed palaeosols, distal fluvial, aeolian and shallow-marine deposits (including both fine-grained siliciclastic and carbonate rocks), and regionally subparallel chronozones are all typical features of forebulge and back-bulge depozones. Because the four types of depozone migrate with the thrust belt, correct interpretation of tectonic processes based on strata in foreland basin systems must be founded on the recognition of the depozone context of the sediment during deposition, rather than the ultimate spatial configuration of the sediment with respect to the thrust belt.

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