Journal of African Earth Sciences 38 (2004) 225–242

Journal of African Earth Sciences

www.elsevier.com/locate/jafrearsci

Geological Society of Africa Presidential Review, No. 7

Retroarc foreland systems—evolution through time

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Abstract

Retroarc foreland systems form through the flexural deflection of the lithosphere in response to a combination of supra- and sublithospheric loads. Supracrustal loading by orogens leads to the partitioning of foreland systems into flexural provinces, i.e. the foredeep, forebulge, and back-bulge. Renewed thrusting in the orogenic belt results in foredeep subsidence and forebulge uplift, and the reverse occurs as orogenic load is removed by erosion or extension. This pattern of opposite vertical tectonics modifies the relative amounts of available accommodation in the two flexural provinces, and may generate out of phase (reciprocal) proximal to distal stratigraphies. Coupled with flexural tectonics, additional accommodation may be created or destroyed by the superimposed effects of eustasy and dynamic (sublithospheric) loading. The latter mechanism operates at regional scales, and depends on the dynamics and geometry of the subduction processes underneath the basin. The eustatic and tectonic controls on accommodation may generate sequences and unconformities over a wide range of time scales, both > and $<10^6$ yr.

The interplay of base level changes and sediment supply controls the degree to which the available accommodation is consumed by sedimentation. This defines the underfilled, filled, and overfilled stages in the evolution of a foreland system, in which depositional processes relate to sedimentation in deep marine, shallow marine, and fluvial environments respectively. Each stage results in typical stratigraphic patterns in the rock record, reflecting the unique nature of flexural and longer-wavelength controls on accommodation. Predictable shifts in the balance between flexural tectonics and dynamic loading allow subdivision of the first-order foreland cycle into early and late phases of evolution dominated by flexural tectonics, and a middle phase dominated by system-wide dynamic subsidence. The early phase dominated by flexural tectonics corresponds to an early underfilled foredeep and a forebulge elevated above base level, whose erosion and rapid progradation results in the formation of the forebulge (basal) unconformity. The middle phase dominated by dynamic subsidence corresponds to a stage of system-wide sedimentation, when the forebulge subsides below the base level and the foredeep goes from a late underfilled to a filled state. The late stage dominated by flexural tectonics corresponds to the first-order overfilled stage of foreland evolution, when fluvial sedimentation is out of phase across the flexural hingeline of the foreland system.

Keywords: Foreland systems; Orogenic loads; Dynamic subsidence; Flexural provinces; Underfilled forelands; Filled forelands; Overfilled forelands

Contents

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0899-5362/$ - see front matter
doi:10.1016/j.jafrearsci.2004.01.004

1. Introduction

Within the context of plate tectonics, foreland systems are associated with convergent plate margins, where fold-thrust (orogenic) belts form along the edge of the overriding continent (Fig. 1). These mountain ranges represent supracrustal loads that press the lithosphere down on both sides of the convergent plate margin, generating accommodation via flexural deflection, i.e. foreland basins (Beaumont, 1981; Jordan, 1981). The newly created depozones are referred to as proarc (or simply pro-) foreland basins, where placed in front of the orogenic belt—on the descending (pro-lithosphere) plate, or as retroarc (retro-) foreland basins, where placed behind the orogenic belt—on the overriding (retro-lithosphere) plate (Fig. 1). One important difference between the proarc and retroarc foreland settings is that the retro-lithosphere is subject to the manifestation of additional mechanisms that create accommodation, related to sublithospheric forces. Sublithospheric "loading" of the overriding plate is primarily caused by the drag force generated by viscous mantle corner flow coupled to the subducting plate, especially where subduction is rapid and/or takes place at a shallow angle beneath the retroarc foreland basin (Mitrovica et al., 1989; Gurnis, 1992; Holt and Stern, 1994; Burgess et al., 1997). This corner flow-driven "dynamic" loading generates accommodation at continental scales, with subsidence rates decreasing exponentially with distance away from the orogen in a cratonward direction (Fig. 1). In addition to dynamic subsidence, the gravitational pull of the subducting slab also contributes to the sublithospheric loading of the retro-lithosphere (Fig. 1).

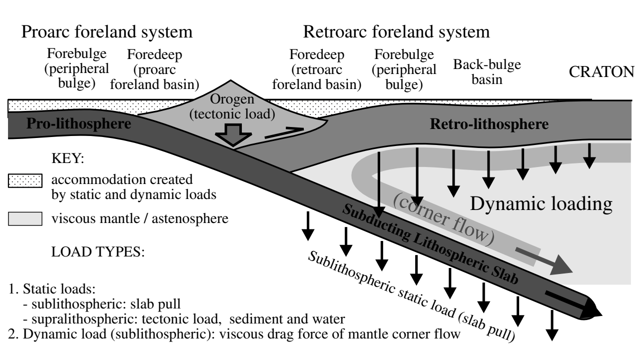

Fig. 1. Proarc and retroarc foreland systems—tectonic setting and controls on accommodation (modified from Catuneanu et al., 1997a). Foreland systems form by the flexural deflection of the lithosphere under a combination of supra- and sublithospheric loads. See text for details.

Excepting for dynamic loading, all other types of supra- and sublithospheric subsidence mechanisms relate to the gravitational pull of "static" loads represented by the subducting slab, the orogen, or the sediment–water mixture that fills the foreland accommodation created by lithospheric flexural deflection (Fig. 1). The static tectonic load of the orogen and the sublithospheric dynamic loading are most often invoked as the primary subsidence mechanisms that control accommodation and sedimentation patterns in retroarc foreland settings (Beaumont et al., 1993; DeCelles and Giles, 1996; Pysklywec and Mitrovica, 1999; Catuneanu et al., 1999a, 2002). The static load of the sediment–water mixture is only of secondary importance in the formation of foreland basins, because accommodation by flexural deflection must be created first, before sediments can start to accumulate.

Even though only one of several subsidence mechanisms, tectonic loading by orogens provides the defining feature of foreland systems, i.e. their partitioning into flexural provinces: foredeep (foreland basin), forebulge (peripheral bulge) and back-bulge (Fig. 1). As a matter of semantics, the foreland system refers to the sum of three flexural provinces, whereas the foreland basin only refers to the foredeep flexural province. Along the flexural profile, the uplift of the forebulge is virtually synchronous with the subsidence of the foredeep, and is caused by the rapid lateral displacement of sublithospheric viscous mantle material as a result of lithospheric downwarp beneath the orogen and the adjacent foredeep (C. Beaumont, pers. com., 2002). The issue of whether the forebulge region may or may not receive and preserve sediments depends on the interplay of flexural tectonics and other mechanisms that control accommodation, and will be tackled in subsequent sections of the paper.

It can be concluded that foreland systems form by the flexural deflection of the lithosphere in response to a combination of supra- and sublithospheric loads. The magnitude of this deflection varies with the amount and distribution of loads, as well as with the physical attributes of the lithosphere that is subject to flexural deformation. This paper summarizes the mechanisms that control the formation and evolution of retroarc foreland systems, as well as their diagnostic stratigraphic signatures. Field examples are mainly provided by the case studies of the Western Canada and southern African Main Karoo basins, but other Precambrian and Phanerozoic foreland systems are also discussed for comparison.

2. Controls on accommodation

Accommodation in retroarc foreland systems is primarily controlled by tectonic forces, as tectonism is responsible for the formation of this basin type in the first place. Subsidence in these settings, whether related to static or dynamic loads, is always differential, with rates generally increasing towards the associated orogenic belt. This gives the basin fill an overall wedge-shaped geometry, which in turn demonstrates the reality of tectonic tilt and hence the dominance of the tectonic control.

The factors that modify the amounts of accommodation in the basin may be classified on the basis of their regional significance. One set of "basin-scale controls" includes allogenic factors that act in a predictable manner at the scale of the basin, and may be used to model the overall evolution of the basin. Additional controls may locally modify the amounts of accommodation, and affect only specific areas within the basin.

2.1. Basin-scale controls

The most important basin-scale controls on accommodation include flexural tectonics related to tectonic loads, dynamic subsidence related to subduction-induced corner flows, and sea level changes. Eustatic fluctuations are global in nature and may affect depositional processes in all types of basins; flexural tectonics characterizes both types of foreland basins, in proarc and retroarc settings; dynamic subsidence is diagnostic for retroarc foreland systems. The following discussion focuses on the two main tectonic controls on accommodation in a retroarc foreland setting, namely flexural tectonics and dynamic subsidence.

Fig. 2 illustrates the separate and combined effects of flexural tectonics and dynamic loading. Flexure of the retro-lithosphere under orogenic loading results in the partitioning of the foreland system into foredeep, forebulge and back-bulge flexural provinces. This lithospheric deflection resembles the shape of a sine curve, with the amplitude attenuated with distance. The amounts of foredeep subsidence and forebulge uplift, which define the vertical scale of the flexural profile, are proportional to the mass of the applied orogenic load, and inversely proportional to the flexural rigidity of the lithosphere. The attenuation with distance of the sinusoidal flexural profile is very rapid, such that the magnitude of forebulge uplift is generally at least 20 times less than the amount of foredeep subsidence (Crampton and Allen, 1995). Accordingly, while the flexural downwarp of the foredeep is usually in a range of kilometers, the magnitude of forebulge uplift is generally less than 200 m (Crampton and Allen, 1995; Catuneanu and Sweet, 1999).

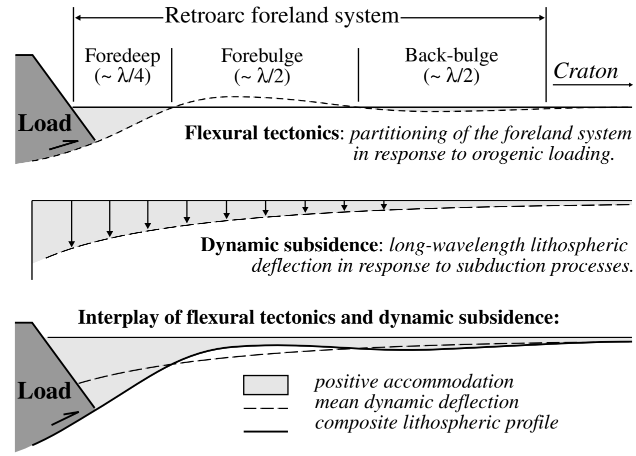

Fig. 2. Interplay of flexural tectonics and dynamic subsidence in retroarc foreland systems. These two basin-scale controls on accommodation are independent of each other, being related to static and dynamic loading respectively, and may record amplitude fluctuations over different time-scales. The balance between their rates controls the direction and magnitudes of base level changes in the different flexural provinces of the foreland system—see Catuneanu et al. (1999a) for a full discussion. λ = flexural wavelength.

The wavelength of the sinusoidal flexural profile, defining the horizontal scale of the foreland system, varies with the basin, depending primarily on the rheology and thickness of the underlying lithosphere (Watts, 1992; Beaumont et al., 1993). For comparison, an effectively elastic plate (infinite relaxation time) would generate a wider foreland basin than a visco-elastic plate (finite relaxation time), and so would a thicker, older or less deformed plate (with higher flexural rigidity and longer relaxation time). Computer modeling of foreland systems shows that for an effectively elastic lithosphere, the foredeep may reach a few hundred kilometers in width, up to more than 400 km for thicker plates (Johnson and Beaumont, 1995). An example of a foreland system developed on continental lithosphere with high flexural rigidity is the Western Canada Basin, where the hingeline separating the foredeep from the forebulge has been mapped as far as 350 km away from the orogenic front (Catuneanu et al., 1997b). Similar distances describe the scale of the Karoo foreland system (Main Karoo basin, southern Africa), which also formed on thick and old (high flexural rigidity) Precambrian crust, largely of the Kaapvaal craton (Catuneanu et al., 1998). Where the foreland system develops on a less rigid and fractured lithosphere (modeled as visco-elastic), the foredeep may be much narrower, such as in the case of the Alpine "molasse" basins with a width of less than 150 km (Homewood et al., 1986; Crampton and Allen, 1995). Young and therefore less rigid plates also generate narrow foredeeps, which is generally the norm with the older, Precambrian foreland systems that formed on newly cratonized continental lithosphere. A good example is the Late Archean Witwatersrand foredeep in South Africa (Kaapvaal craton), with a width of only about 130 km (Catuneanu, 2001).

One important feature of the flexural profile of a foreland system, irrespective of actual size, is the relative proportion between the extent of flexural provinces, as measured along dip (Fig. 2). The uplifted peripheral bulge and the back-bulge depozone are both significantly wider than the foredeep, each measuring about half of the wavelength of the sinusoidal flexural profile. This may be explained by comparing the lithospheric flexural profile with a transversal wave that is attenuated with distance (looses energy) away from the source. Along the direction of propagation of such a wave, each positive or negative deflection corresponds to half of its wavelength. The foredeep is also part of a half wavelength portion of the flexural "wave", but as the other part of the same downwarp is occupied by orogenic structures, the foredeep is invariably the narrowest flexural province, extending only about a quarter of the wavelength along dip (Fig. 2). Following Turcotte and Schubert's (1982) modeling of the flexural response of an elastic plate floating above a fluid mantle substrate, Crampton and Allen (1995) also reached the conclusion that the wavelength of the deflection is a constant for the entire flexural profile of the foreland system, assuming that the plate is homogeneous.

According to Turcotte and Schubert's (1982) theoretical considerations, expanded by DeCelles and Giles (1996) to include the flexural subsidence of the back-bulge basin, the wavelength of the flexural deflection of an elastic plate under loading depends on the flexural parameter of the lithosphere (α), and equals $2\pi\alpha$ (for an infinite plate) or $3\pi\alpha/2$ (for a broken plate). Based on these formulae, the horizontal width of flexural provinces measured along the dip of an infinite plate is $\pi\alpha/2$ for the foredeep, and $\pi\alpha$ for both the forebulge and the back-bulge regions. These formulae may be extrapolated to broken plates, by introducing a multiplication coefficient of 3/4. The flexural parameter varies with the rheology of the lithosphere and the contrast in density between the mantle and the basin fill, and may range from tens to hundreds of kilometers.

If flexural tectonism related to orogenic loading was the only mechanism controlling accommodation in the foreland system, the uplifted forebulge area would never receive and preserve sediments. Under such circumstances, the peripheral bulge would be subject to erosion during the entire evolution of the basin. This is the case with a number of foreland systems, including the Alpine molasse basins, or the Witwatersrand Basin. In such cases, the depositional areas are generally restricted to the foredeep and the back-bulge depozones (e.g., see top cross-section in Fig. 2 for a conceptual illustration of this principle; also, see Catuneanu, 2001, for a case study). Other foreland systems however accumulate sediments across all flexural provinces, resulting in the formation of foreland-fill wedges close to 1000 km wide. What accounts for these differences?

Independent of flexural tectonics, dynamic subsidence generates accommodation at continental scales, with rates decreasing exponentially with distance from the subduction zone (Fig. 2). This additional long-wavelength lithospheric deflection may potentially bring the entire foreland flexural profile below the base level, as illustrated by the composite lithospheric profile in Fig. 2. Under these circumstances, the forebulge region may receive sediments for as long as the rates of dynamic subsidence exceed the rates of flexural uplift due to orogenic loading. The balance between the rates of dynamic loading and flexural tectonics is the key in determining the direction and magnitude of base level changes in the different flexural provinces of the foreland system, and implicitly the stratigraphic architecture across the foreland system (see Catuneanu et al., 1999a, for a full discussion). For example, the dominance of dynamic loading over the effects of tectonic loading implies base level rise across the entire system, even though with contrasting rates across the flexural hingeline that separates the foredeep from the forebulge, and, depending upon sediment supply as well, the basin may likely be dominated by marine sedimentation. In contrast, a stage of flexural uplift outpacing the rates of dynamic subsidence leads to the formation of unconformities, commonly restricted to the flexural province that is subject to uplift, and the basin tends to be dominated by nonmarine sedimentation. The major transgressive–regressive cycles observed in most foreland systems may in part be related to this changing balance between dynamic and static loading, and of course also in part by fluctuations in sediment supply (Catuneanu et al., 1999a).

The composite lithospheric profile in Fig. 2 (average flexural profile in Fig. 3) changes through time in response to fluctuations in the amount of orogenic loading. Such orogenic cycles of thrusting (loading) and quiescence (erosional or extensional unloading) may operate over a wide range of time scales, both > and <1 My (Cloetingh, 1988; Cloetingh et al., 1985, 1989; Peper et al., 1992; Catuneanu et al., 1997b; Catuneanu and Sweet, 1999). Renewed thrusting (loading) in the orogenic belt generates subsidence in the foredeep and uplift of the forebulge, and the reverse occurs during orogenic

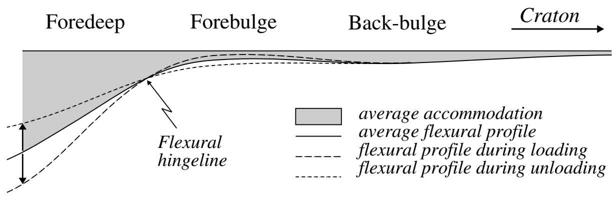

Fig. 3. Flexural response to orogenic loading and unloading (modified from Catuneanu et al., 1997a). Renewed thrusting in the orogenic belt (loading) results in foredeep subsidence and forebulge uplift. The reverse occurs during stages of orogenic quiescence (erosional or extensional unloading): the foredeep undergoes uplift as a result of isostatic rebound, compensated by subsidence of the forebulge.

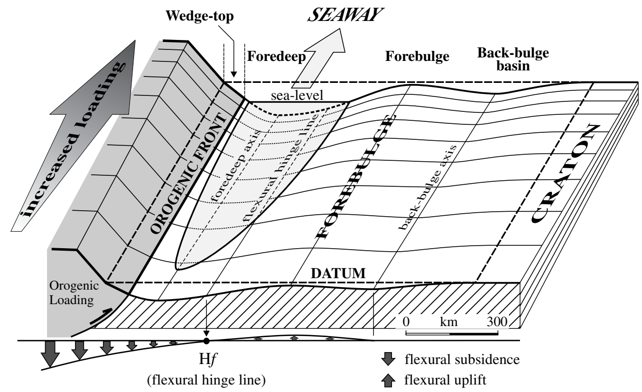

Fig. 4. Configuration of a foreland system during orogenic loading with strike variability (modified from Catuneanu et al., 2000). The topographic elevation of the adjacent craton, approximated with a horizontal plane, is taken as a datum. The apex of the forebulge is below the elevation of the adjacent craton because of the effect of dynamic loading (not shown). The base level of deposition within the foreland system ($\sim$ sea level of the interior seaway) may be in any position (below, above, or superimposed) relative to the datum, although surface processes on the craton (sedimentation, erosion) tend to adjust the datum to the base level. Note that the magnitude of forebulge uplift increases with the amount of foredeep subsidence, and implicitly with the amount of tectonic loading.

unloading: i.e., isostatic rebound of the foredeep, compensated by subsidence of the forebulge. Ongoing research of Pleistocene and Holocene glacio-isostatic cycles of crustal adjustment shows that the rise of the forebulge is virtually synchronous with the subsidence of the foredeep, indicating rapid flow and pressure equilibration of the sublithospheric viscous mantle as a result of changes in supracrustal loading (C. Beaumont, pers. comm., 2002).

This flexural behavior of the foreland system in response to orogenic cycles of loading and unloading generates contrasting base level changes across the flexural hinge line (Fig. 3). As inferred above, this contrast may only be in terms of rates (high vs. low subsidence rates, explaining, for example, the manifestation of coeval transgressions and normal regressions in shallow seas across the hingeline), or in terms of direction of base level changes (rise vs. fall, explaining, for example, the coeval formation of depositional sequences and stratigraphic hiatuses across the hingeline). These contrasts in the direction and/or the magnitude of base level shifts across the hingeline define the key diagnostic feature of foreland systems, and are fundamental to understanding the stratigraphic architecture of this basin type.

It is also important to note that the topographic profile of foreland systems does not necessarily have to parallel the shape of the lithospheric flexural profile. For example, if the entire amount of accommodation shown in Fig. 3 is consumed by sedimentation, the landscape topography may be approximately flat, without showing the presence of the flexural forebulge. From there, isostatic rebound of the foredeep coupled with subsidence of the forebulge (stage of orogenic quiescence) may lead to a topography that is the mirror image of the lithospheric flexural profile. In this scenario, the foredeep region of the foreland system may be topographically more elevated than the subsiding forebulge region, which becomes a sag, even though the lithospheric profile still has a flexural forebulge (Fig. 3). These aspects are tackled in more detail in the section that deals with the stratigraphy of the retroarc foreland systems.

Figs. 2 and 3 are simple two-dimensional representations, which only explain the regional trends along dip-oriented profiles. In a three-dimensional view however, the geology is complicated by the strike variability in orogenic loading, which triggers an axial tilt in the basin as illustrated in Fig. 4. This strike variability in orogenic loading is generally the norm rather than the exception, as it is highly unlikely that a thrust sheet would have exactly the same mass everywhere along the strike of the fold-thrust belt. This differential loading along strike generates a tilt in the direction of increased loading, which in turn controls the direction of shoreline transgressions and regressions, as well as the flow of fluvial systems.

2.2. Additional controls

In addition to the basin-scale controls on the evolution of the foreland system, which provide the basis for regional modeling (e.g., Figs. 1–4), the amounts of available accommodation may be modified by the manifestation of local factors, such as differential uplift and subsidence of basement blocks (i.e., basement tectonics), dissolution or displacement of salt deposits that may be present in the subsurface, or differential compaction. These secondary controls on accommodation generate stratigraphic "anomalies", i.e. departures from the predicted geometry of sedimentary sequences that fill the basin.

Basement tectonics, triggered by the reactivation of crustal faults, is probably the most significant of these secondary controls on accommodation. Differential subsidence and uplift of basement blocks may generate mini-basins and arches (horst structures) within the foreland system, with an apparent random distribution that depends on the nature of the basement and the way the intra-plate stress fields propagate and affect the zones of weakness in the underlying crust. The smooth and symmetrical sinusoidal flexural profiles modeled in Figs. 2 and 3 are idealized, based on the assumption that the underlying basement is homogeneous. This is rarely the case in reality, as basements tend to be heterogeneous, composed of blocks of different ages and lithologies, and hence with different rheological properties.

A consequence of having a heterogeneous basement underlying the foreland system is that the expected relative proportions along dip between flexural provinces may be distorted. This is because the position of flexural hingelines that separate the foredeep from the forebulge, or the forebulge from the back-bulge basin, is potentially controlled by the boundaries between basement blocks with different rheologies, which may not necessarily fit the inflexion points of the theoretical sine curve. For example, the hingeline that outlines the Late Campanian–Early Maastrichtian foredeep of the Western Canada foreland system is virtually superimposed, in map view, on the limit between the Eyehill High and the Medicine Hat Block/Vulcan Low Archean basement provinces (Catuneanu et al., 1997b). Similarly, the hingeline between the foredeep and the forebulge of the Karoo Basin follows closely the boundary between the Namaqua-Natal Belt and the Kaapvaal Craton in the underlying basement. In the same basin, the limit between the forebulge and the back-bulge depozone is controlled by the contact between the Bushveld and the Pietersburg blocks of the Kaapvaal Craton, which was reactivated during the evolution of the basin (Catuneanu et al., 1999b). Fig. 5 illustrates the situation of the Karoo Basin during the Late Carboniferous, where the boundaries between flexural provinces, and implicitly the distribution of syn-depositional glacial facies of the Dwyka Group, are primarily controlled by basement structures. In this example, the width of the forebulge is greater than expected from the flexural modeling of a homogeneous plate.

Besides the influence of basement tectonics and heterogeneity on the location of flexural hingelines, and implicitly on the extent of flexural provinces, reactivation of crustal faults by foreland system tectonism may also control thickness and facies trends within individual flexural provinces. As documented in a series of case studies in the Western Interior foredeep of Canada and the United States (e.g., Hart and Plint, 1993; Plint et al., 1993; Pang and Nummedal, 1995; Donaldson et al., 1998), flexure over reactivated crustal faults resulted in varying rates of subsidence across basement block

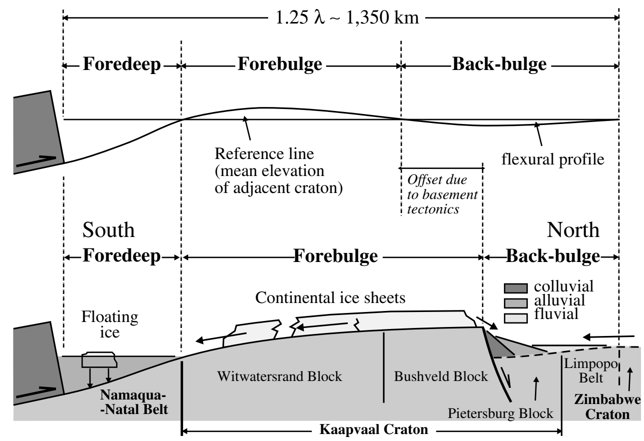

Fig. 5. Flexural and basement controls on the distribution of Late Carboniferous Dwyka glacial facies in the Karoo Basin. The foredeep subsided below the sea level, hosting sedimentation from floating ice (dropstones within a marine silty matrix). The forebulge was elevated above the sea level, supporting the formation of continental ice sheets. The back-bulge area was fault-bounded to the south, hosting a half-graben style of sedimentation with colluvial, alluvial, and fluvial facies. The offset between the theoretical flexural profile and the actual field cross-section is caused by the syn-depositional reactivation of the fault boundary between the Bushveld and the Pietersburg basement blocks. λ = flexural wavelength.

boundaries, explaining localized incision, paleogeographic trends, and thickness and facies patterns that parallel terrane boundaries in the basement. The importance of basement tectonics on sedimentation was challenged by Aitken (1993), who noted that many basement features are only occasionally reflected, if at all, in the sedimentary cover overlying the basement. The lack of correlation between some basement structures and the overlying sedimentary record may however be caused by selective reactivation of crustal faults, and/or by rapid "healing" of the underlying topography by the oldest sedimentary sequences of the basin.

3. Stratigraphy of retroarc foreland systems

Retroarc foreland systems have a distinct stratigraphic architecture relative to any other basin type, which reflects the contrasts in the direction and/or the magnitude of base level shifts across flexural higelines. The pattern of opposite flexural tectonics between the foredeep and the forebulge (Fig. 3) modifies the relative amounts of available accommodation in the two flexural provinces, generating out of phase (reciprocal) proximal to distal stratigraphies (Catuneanu et al., 1997a,b, 1999a, 2000; Catuneanu and Sweet, 1999).

Two styles of reciprocal stratigraphies have been defined in relation to the pattern of base level changes across the foreland system (Catuneanu et al., 1999a). One style refers to the case where the contrast in base level changes is only in terms of rates (high vs. low subsidence rates across the hingeline), and consists of a conformable succession of correlative transgressive and normal regressive systems tracts. This case requires continuous basin-wide sedimentation, with the rates within the range of variation of the rates of base level rise. A second style of reciprocal stratigraphies refers to the case where base level changes across the flexural hingeline take place in opposite directions, resulting in sequences correlative to age-equivalent stratigraphic hiatuses (sequence boundaries) in relation to coeval rising and falling base level respectively. The shift from one style to another in the evolution of a basin is linked to the change in the balance between the rates of flexural tectonics and the rates of longer-wavelength controls on accommodation, such as dynamic subsidence and sea-level changes (see Catuneanu et al., 1999a, for a full discussion). Cyclic changes within the framework of this balance may result in major (second-order) transgressive–regressive shoreline shifts within the foreland system, and implicitly in changes between dominant marine or nonmarine sedimentation across the basin.

The interplay of base level changes and sediment supply controls the degree to which the available accommodation is consumed by sedimentation. This defines the underfilled, filled and overfilled stages in the evolution of the foreland system, in which depositional processes are dominated by deep marine, shallow marine, or fluvial sedimentation, respectively (Sinclair and Allen, 1992). The change from underfilled to overfilled stages is best observed in the foredeep, because the forebulge may be subject to erosion in the absence of (sufficient) dynamic loading, or, at most, it may accommodate shallow marine to fluvial environments even when the foredeep is underfilled. An example of such a succession of stages is illustrated in Fig. 6, from the case study of the Karoo Basin.

It can be noted that the evolution of any foreland basin is somewhat predictable, starting with an underfilled phase (“flysch” style of sedimentation), and ending with an overfilled phase (“molasse” style of sedimentation) (Crampton and Allen, 1995). Early forelands tend to be underfilled because the onset of tectonic (orogenic) loading is generally undercompensated by sediment supply, due to the low elevation of the young orogenic structures (Allen et al., 1986; Covey, 1986; Stockmal et al., 1986; Desegaulx et al., 1991; Sinclair and Allen, 1992; Crampton and Allen, 1995). At the same time, the forebulge associated with the earliest stage of foreland system evolution is always subject to erosion, and hence recorded as an unconformity at the base of the foreland basin-fill (basal/forebulge unconformity; Crampton and Allen, 1995; Catuneanu, 2001; Fig. 6). This is because the onset of dynamic loading lags in time behind the initiation of subduction and tectonic loading, as it takes time for the subducting slab to reach far enough beneath the overriding plate to generate a viscous corner flow. Once dynamic subsidence is high enough to outpace the flexural uplift of the forebulge, the entire foreland system may be lowered below the base level (Fig. 2). The change in forebulge status from an erosional area to a depositional area may take place during the underfilled stage of basin evolution, as in the case of the Karoo Basin. The configuration of the earliest Karoo foreland system accounts for a forebulge elevated above the base level during the Dwyka time (Fig. 5; “early underfilled” stage in Fig. 6), followed by a time of system-wide sedimentation during the Ecca time (“late underfilled” stage in Fig. 6; Catuneanu et al., 1998, their Fig. 13). This transition may be explained by the initiation of dynamic loading associated with the subduction of the paleo-Pacific plate beneath Gondwana. In this example, the lag time between the onset of subduction and tectonic loading in the Namurian (Smellie, 1981; Johnson, 1991; Mpodozis and Kay, 1992; Visser, 1992), and the onset of dynamic loading at the beginning of the Permian was at least 40 My, corresponding more or less with the entire duration of the Late Carboniferous (Fig. 6).

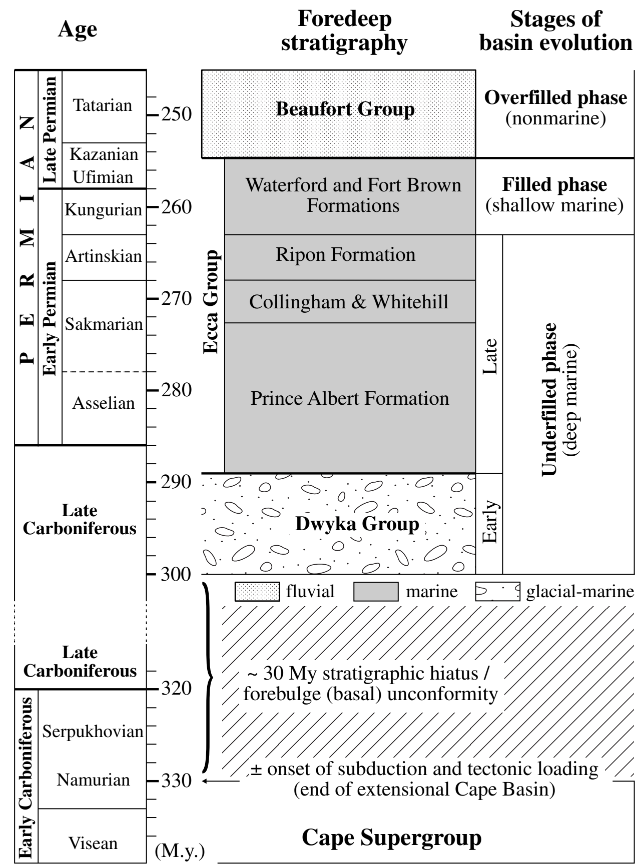

Fig. 6. Early Carboniferous to Late Permian evolution of southern Africa, showing the change from extensional to compressional and flexural tectonic regimes, as well as the stages in the evolution of the Karoo foredeep.

Overfilled phases of dominantly fluvial sedimentation mark the late stages of foreland system evolution, when sediment supply is high and both orogenic and dynamic loading subside towards the end of the compressional regime in the adjacent fold-thrust belt. Fig. 7 summarizes the overall evolution of a retroarc foreland system, from underfilled to overfilled. The initial underfilled nature of the foredeep is generally the norm, as the earliest tectonic loads are likely placed below the sea level (not shown in Fig. 7; Stockmal et al., 1986; Desegaulx et al., 1991; Sinclair and Allen, 1992) and hence subject to little erosion. As sediment supply increases through time with the gradual uplift of the fold-thrust belt, sedimentation catches up with the available accommodation and the deep seas become shallow seas, which eventually regress to make room for a dominantly continental environment.

This first-order cycle of foreland evolution and sedimentation correlates with a cycle of overall increase and decline in the rates of dynamic subsidence (e.g., Pysklywec and Mitrovica, 1999), as the rates of subduction increase from zero at the onset of compression, and decrease back to zero towards the transition from compression to extension. It is thus expected that flexural uplift outpaces dynamic subsidence in the earliest and latest stages of basin evolution, with the dynamic loading dominating the intermediate stage of basin-wide shallow marine sedimentation (Fig. 7). This first-order scenario only describes general trends, and second-order (and superimposed on these, higher frequency) fluctuations in the balance between the rates of flexural tectonics, dynamic subsidence, sea level shifts and sedimentation are common and explain the complexity of changes in depositional systems across the foreland system with time. Forward modeling of such changes has been used to simulate synthetic stratigraphic models of foreland system development under specific conditions of sediment supply and accommodation (e.g., Flemings and Jordan, 1989; Jordan and Flemings, 1991; Sinclair et al., 1991). The sections below provide examples of underfilled, filled and overfilled stratal stacking patterns from the case studies of the Karoo and Western Canada foreland systems (Figs. 8 and 9).

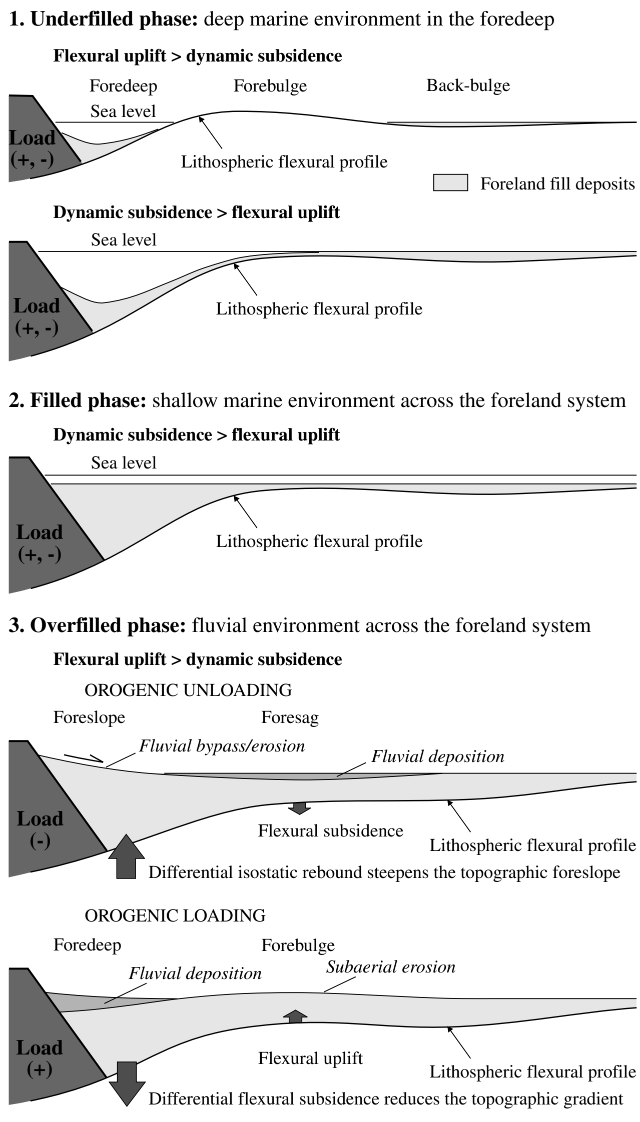

Fig. 7. Underfilled, filled and overfilled stages of foreland system evolution. The fill of the foreland system corresponds to a first-order cycle of changing balance between flexural tectonics and dynamic subsidence. Note the difference between bathymetric/topographic profiles and the lithospheric flexural profile. (+, −) refer to increases and decreases in orogenic load, respectively.

3.1. Underfilled forelands

Underfilled foreland systems are defined by rapidly increasing water depths in the foredeep, as accommo-

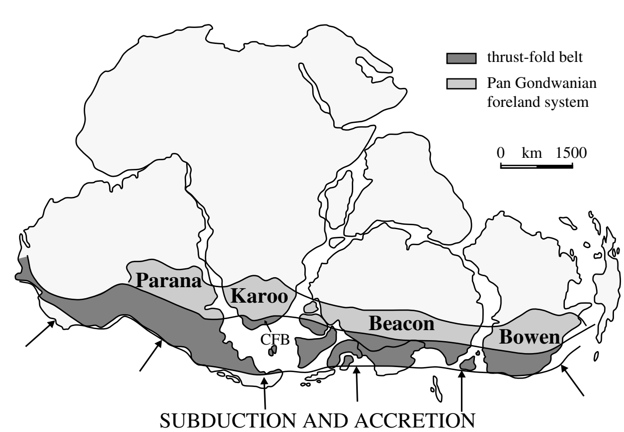

Fig. 8. Paleogeographic reconstruction of Gondwana, showing the subduction from south to north of the paleo-Pacific plate beneath the supercontinent. As a result of compression and terrain accretion, a ca. 6000 km long fold-thrust belt formed along the southern margin of Gondwana, with an associated retroarc foreland system to the north. Following the breakup of Gondwana, portions of this foreland system are now preserved in South America, South Africa, Antarctica and Australia. CFB = Cape Fold Belt.

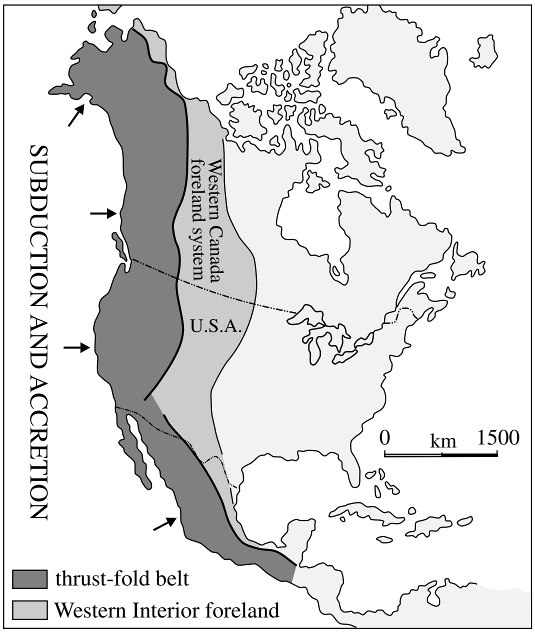

Fig. 9. The Western Interior foreland system, in relation to the associated fold-thrust belt and the convergent margin of North America.

dation is created faster relative to the rates of sedimentation. This results in a relatively deep water environment, with water depths in a range of a few hundred meters. Underfilled conditions characterize the early stages of basin development, when tectonic loads are below sea level or have a low elevation above sea level. The initial subsidence of the foredeep is accompanied by the flexural uplift of the forebulge above the base level, which leads to the formation of the basal (forebulge) unconformity. This unconformity has the significance of a first-order sequence boundary, as it separates the sedimentary fill of an extensional basin, below, from foreland basin deposits above (Fig. 6). Gradual increases in the rates of dynamic loading with time may lead to the lowering of the forebulge below the base level, and hence to basin-wide sedimentation. These two stages are illustrated in the top two diagrams of Fig. 7. This predictable trend of evolution allows us to separate an early underfilled stage characterized by foredeep sedimentation and forebulge emergence (flexural uplift > dynamic subsidence), followed by a late underfilled stage of basin-wide sedimentation (dynamic subsidence > flexural uplift) (Fig. 7).

The early underfilled stage is exemplified by the case study of the Dwyka Group in the Karoo Basin (Fig. 5). The onset of emplacement of tectonic loads in the Carboniferous led to subsidence and the establishment of a glacial-marine environment in the foredeep, accompanied by the elevation of a peripheral bulge above sea level. The latter region, with a wavelength controlled in part by basement heterogeneities (see discussion above), allowed for the formation of continental ice sheets (Visser, 1991).

The late underfilled stage applies to the underfilled portion of the Ecca Group (Prince Albert to Ripon Formations; Fig. 6), when sedimentation extended across the entire foreland system, and the foredeep accumulated pelagic to gravity flow sediments. At a first-order level, dynamic subsidence outpaced the rates of flexural uplift, leading to the lowering of the peripheral bulge below the sea level and the manifestation of basin-wide transgressions of the interior seaway (Fig. 10). At a higher frequency level, fluctuations in sediment supply and in the balance between the rates of flexural tectonics and dynamic loading may result in forced regressions on the side of the basin that is subject to flexural uplift, coeval with transgressions or normal regressions of the opposite shoreline of the interior seaway (Fig. 11). The latter is the case with the age-equivalent Ripon and Vryheid formations in the Karoo Basin, when submarine fans in the foredeep formed at the same time as the progradation of fluvial–deltaic sequences over the forebulge (Catuneanu et al., 2002).

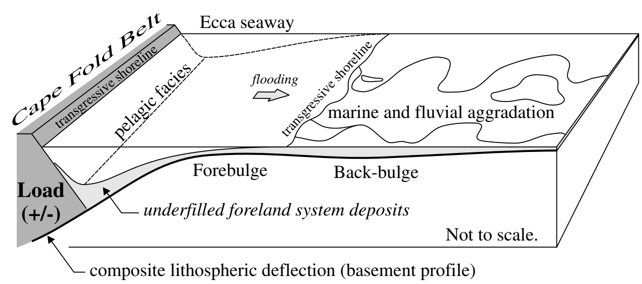

Fig. 10. Late underfilled stage in the evolution of a foreland system, where increased dynamic subsidence results in the lowering of the peripheral bulge below the sea level (modified from Catuneanu et al., 2002). Transgressions of both proximal and distal shorelines of the interior seaway may be recorded as accommodation outpaces the low rates of sediment supply.

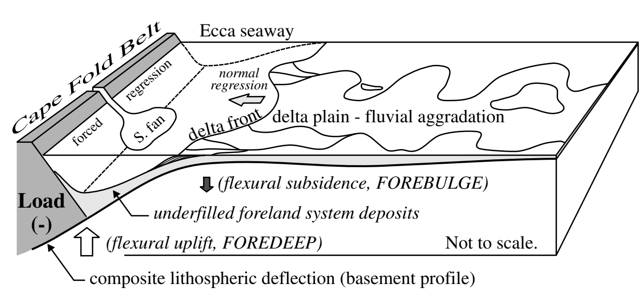

Fig. 11. Late underfilled stage in the evolution of a foreland system, where temporary dominance of flexural tectonics may result in the manifestation of coeval forced and normal regressions of the proximal and distal shorelines of the interior seaway (modified from Catuneanu et al., 2002). In this example, submarine fans in the foredeep (“S. fan” in the diagram) and fluvial-deltaic sequences over the forebulge prograde at the same time into the interior seaway.

This has significant implications for the petroleum exploration of underfilled foreland sequences, as both the proximal turbidites and the distal deltas have the potential to form good hydrocarbon reservoirs.

3.2. Filled forelands

The gradual increase in the amount of sediment supply through time in response to tectonic uplift in the fold-thrust belt leads to a shallowing of the earlier underfilled interior seaway and the establishment of a shallow marine environment across the foreland system. This defines the filled stage of foreland system evolution, where accommodation and sedimentation are more or less in balance (Fig. 7). Maintaining a system-wide shallow marine environment requires that a number of conditions are fulfilled, including (1) a forebulge lowered below sea level, hence (2) dynamic subsidence (or sea level rise) outpacing the rates of flexural uplift, which in turn implies (3) base level rise across the entire foreland system, and (4) sedimentation rates that are within the range of variation of the rates of base level rise.

The rates of dynamic subsidence are expected to peak during this intermediate stage of foreland system evolution, as discussed above, and hence the overall rates of base level rise are potentially at a maximum during filled stages. This means that sedimentation rates also reach a peak during these stages, and therefore the basin is most active from both tectonic and depositional points of view.

A case study for a filled foreland system succession is provided by the Bearpaw Formation of the Western Canada Sedimentary Basin (Fig. 12). This shallow marine succession corresponds to the last basin-wide incursion of the Western Interior seaway, which flooded the entire foreland system from Manitoba to the Foothills of Alberta. The overall (second-order) transgression took place over ca. 3 My (*Baculites scotti* to *Baculites compressus* ammonite zones; Obradovich, 1993), whereas the overall regression lasted for about 9 My (*Baculites compressus* till the end of the Cretaceous; Catuneanu et al., 2000). Sedimentation during this ca. 12 My second-order transgressive-regressive cycle was generally continuous, as the succession is relatively conformable. Several third-order transgressive–regressive sequences have been mapped within the Bearpaw Formation, each corresponding to a flexural cycle of orogenic loading and unloading (Catuneanu et al., 1997b, 2000). The stratigraphic architecture shows coeval accumulation of transgressive facies in the foredeep and normal regressive facies in the forebulge area during stages of orogenic loading, and the converse situation for stages of orogenic unloading (Fig. 13). This defines a style of reciprocal stratigraphies that is typical for filled forelands (Catuneanu et al., 1999a).

A full discussion of the balance between the rates of flexural tectonics, dynamic subsidence and sedimentation that allows for the formation of this style of reciprocal stratigraphies is provided by Catuneanu et al. (1999a). The base level rises across the entire foreland system as dynamic subsidence outpaces the rates of flexural uplift, but with higher rates in the region that is subject to flexural subsidence and with lower rates in the region that is subject to flexural uplift. Given a sediment supply that is within the range of variation of the rates of base level rise, the higher subsidence areas are associated with transgressions, and the lower subsidence areas are associated with normal regressions. If this balance is altered in the favor of sedimentation, i.e. sedimentation > higher subsidence rates, then a basin-wide normal regression takes place, and a transition is made towards an overfilled foreland system. This is the case with the final regression of the Bearpaw seaway (Fig. 13), which may be attributed to a decrease in the

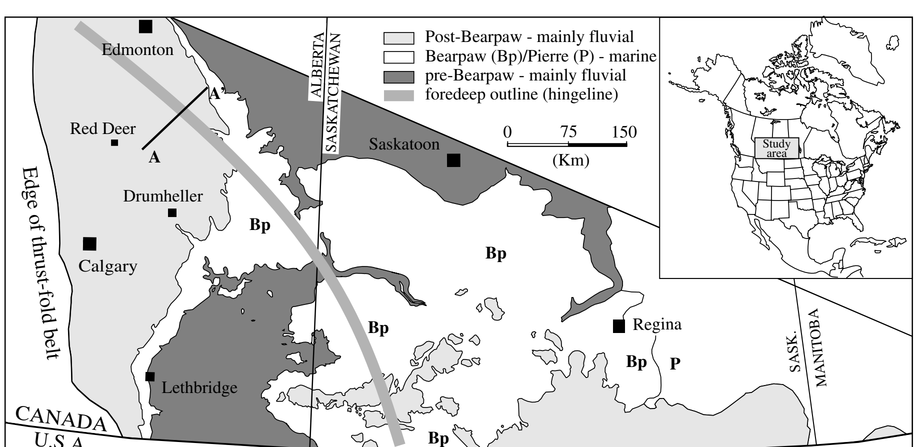

Fig. 12. Geological map showing the location of the Bearpaw Formation in Alberta and Saskatchewan (modified from Catuneanu and Sweet, 1999). Cross-section A–A′ is shown in Fig. 13.

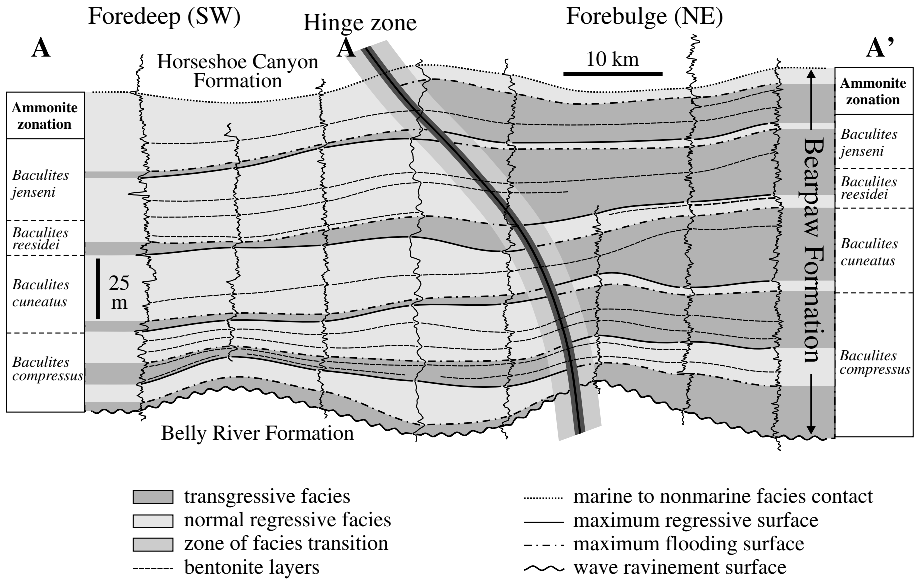

Fig. 13. Stratigraphic cross-section of correlation of gamma ray logs (A–A′ in Fig. 12), showing the distribution of transgressive and normal regressive facies of the Bearpaw Formation in a time framework (modified from Catuneanu et al., 1997b). The underlying and overlying lithostratigraphic formations (Belly River and Horseshoe Canyon) are mainly composed of fluvial facies. The marine succession is relatively conformable and shows a style of reciprocal stratigraphies typical for filled foreland systems, controlled by flexural cycles of orogenic loading and unloading.

rates of dynamic subsidence. If the lower subsidence rates outpace the sedimentation rates, then a basin-wide transgression takes place. A variety of situations can therefore be envisaged and modeled, explaining the variability observed in case studies. As a general trend, the long-wavelength controls on base level rise (dynamic subsidence and/or sea level rise) are dominant at the onset of filled foreland stages, explaining the basin-wide transgressions of the shallow interior seaways, and subside toward the end of the filled stages, allowing the transition to overfilled forelands.

3.3. Overfilled forelands

Overfilled foreland systems are dominated by nonmarine environments, and reflect stages in the evolution of the basin when sediment supply outpaces the available accommodation (Fig. 7). At a first-order level, the overfilled stage represents the final phase in the evolution of a foreland system, when the rates of dynamic subsidence decrease below the rates of flexural uplift. This shift in the balance between the two main controls on foreland accommodation results in a half-system style of sedimentation, where only one flexural province (the one subject to flexural subsidence) receives sediments at any given time. At higher frequency levels, foreland systems may reach an overfilled state any time sedimentation exceeds the rates of base level rise, which may happen during the dominance of either flexural tectonics or dynamic subsidence. In the latter case, basin-wide nonmarine sequences may be generated.

An example of typical overfilled regional architecture is illustrated in Fig. 14, from the case study of the Western Canada foreland system. In contrast to the underlying marine Bearpaw Formation, which is relatively conformable and displays a filled style of reciprocal stratigraphies, the post-Bearpaw nonmarine

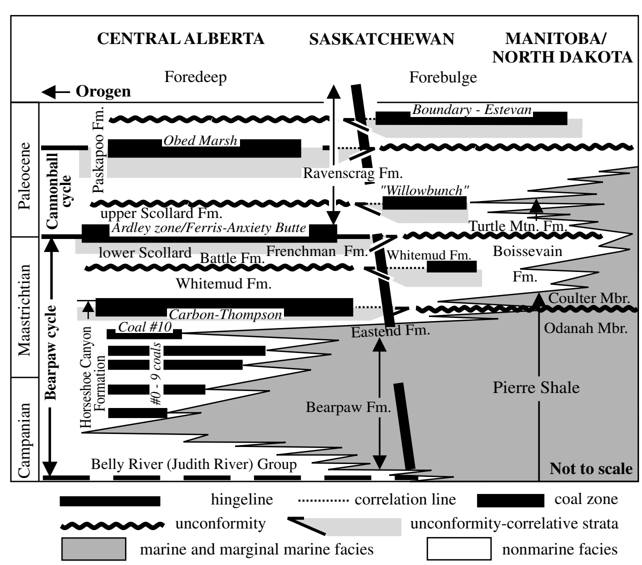

Fig. 14. Generalized dip-oriented cross-section through the Western Canada foreland system, showing the overfilled style of reciprocal stratigraphies in the post-Bearpaw fluvial succession (modified from Catuneanu et al., 1999a). Note that each major unconformity is restricted to the one side of the basin that was subject to flexural uplift at the time.

succession is marked by the presence of significant unconformities. These unconformities are associated with stratigraphic hiatuses of at least 1 My, and are restricted to the flexural province that was subject to uplift at the time of unconformity formation. This demonstrates the fact that flexural tectonics outpaced the rates of dynamic subsidence, and defines the overfilled style of reciprocal stratigraphies in which each unconformity has an age-equivalent depositional sequence on the opposite side of the flexural hingeline (Fig. 14). Flexural tectonism is therefore the dominant control on the overfilled foreland stratigraphy—stages of orogenic loading result in foredeep sequences and coeval forebulge unconformities (sequence boundaries), whereas stages of orogenic unloading lead to the formation of sequence boundaries in the foredeep and age-equivalent sequences in the forebulge area (Figs. 7, 14). Even though of secondary importance in terms of rates, dynamic subsidence is still active during this overfilled stage, as evidenced by the preservation of proximal and distal depositional sequences. In the absence of dynamic loading, each newly accumulated foredeep or forebulge sequence would be eroded during subsequent flexural uplift.

The vertical profiles of foredeep and forebulge fluvial sequences are primarily a function of the changes through time in the type of rivers that bring sediment from the source areas. In turn, the style of fluvial systems is controlled by changes through time in topographic gradients. The reason these gradients change is because both flexural subsidence and isostatic rebound/uplift during stages of orogenic loading and unloading take place with differential rates. For example, the topographic slope of the foredeep area ("foreslope" in Fig. 7) becomes increasingly steeper during orogenic unloading, as the rates of isostatic rebound are highest in the fold-thrust belt and gradually decrease towards the subsiding forebulge. During such a stage, the foreslope is subject to bypass and/or erosion and the subsiding forebulge area ("foresag" in Fig. 7) receives coarser and coarser sediments brought by rivers that are characterized by increasingly higher energy levels (Catuneanu and Sweet, 1999). During orogenic loading, differential flexural subsidence in the foredeep, with the highest rates in the center of loading, reduces the foreslope gradient, which lowers the energy level of fluvial systems through time (Fig. 7). This leads to the accumulation of fining-upward foredeep fluvial sequences, as shown in Fig. 15. Note that the fining-upward trend characterizes each individual sequence, in direct response to lowering slope gradients, but the overall vertical profile of the overfilled foredeep may be coarsening-upward due to the progradation through time of the orogenic front (Fig. 15; Catuneanu and Elango, 2001).

In addition to the differential subsidence along dip, subsidence may also be differential along strike due to

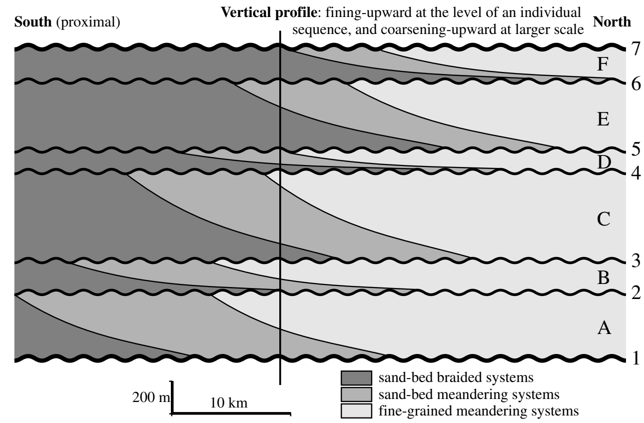

Fig. 15. Internal architecture of the Balfour Formation along a dip oriented profile through the overfilled Karoo foredeep (modified from Catuneanu and Elango, 2001). Changes in fluvial style within each sequence generate fining-upward trends, as a response to decreasing slope gradients through time. At a larger scale, the overall vertical profile is coarsening-upward due to the gradual progradation of the orogenic front. A–F = third-order fluvial sequences; 1 and 7 = second-order sequence boundaries; 2–6 = third-order sequence boundaries.

the variability in orogenic loading. This leads to changes in syn-depositional tilt, hence to changes in paleoflow directions both within a sequence (Catuneanu et al., 2003) and across sequence boundaries (Catuneanu and Elango, 2001). In spite of the differential rates of subsidence and uplift that characterize the evolution of any foredeep basin, sequence boundaries observed at outcrop scale do not necessarily show angular relationships because the slope gradients may only vary within a ±2% range, which is however enough to generate changes in fluvial styles during the deposition of each sequence.

As the duration of loading stages in the fold-thrust belt is generally much shorter in time relative to the duration of quiescence stages, the overfilled foredeep stratigraphy only preserves a fraction of the geological record, with most of the time being absorbed within sequence boundaries. A generalized model of out-of-phase foreland sedimentation during an overfilled stage, showing vertical trends of foredeep and forebulge fluvial sequences, is illustrated in Fig. 16.

The seesaw patterns of sedimentation described in this section are similar to the "antitectonic" model for foreland system development proposed by Heller et al. (1988). This model also predicts the most active proximal sedimentation during thrust loading, followed by deposition in the forebulge area during post-orogenic tectonic rebound, when the foredeep is subject to erosion. The topographic profile of the overfilled foreland system changes significantly during these loading and unloading stages (Fig. 7), causing predictable shifts in the types of fluvial drainage systems (i.e., axial vs. transversal). Orogenic loading induces a proximal axial

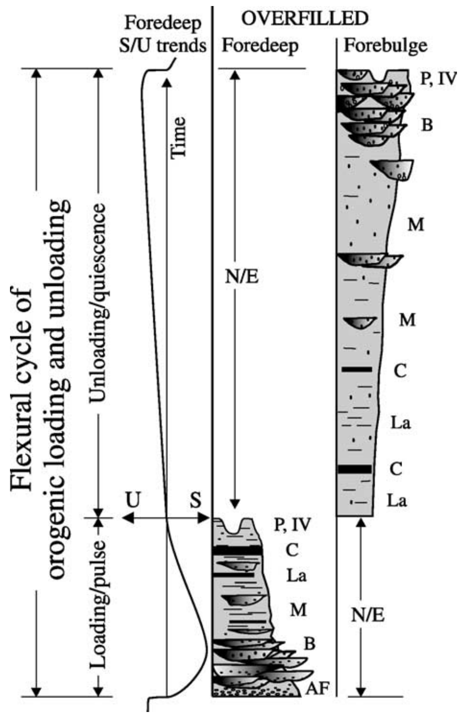

Fig. 16. Generalized model of fluvial sedimentation in overfilled foreland systems dominated by flexural tectonism (modified from Catuneanu and Sweet, 1999). Stages of orogenic loading are shorter in time relative to stages of orogenic unloading. The former result in the accumulation of fining-upward foredeep sequences, the latter lead to the deposition of coarsening-upward forebulge (topographic foresag) successions. Sedimentation across the flexural hingeline is out-of-phase, which defines the overfilled style of reciprocal stratigraphies. Note that the vertical axis indicates time—due to the increase in subsidence rates in a proximal direction, the foredeep sequences are actually thicker than the forebulge ones. Abbreviations: U = uplift; S = subsidence; N/E = non-deposition or erosion; AF = alluvial fans; B = braided fluvial systems; M = meandering fluvial systems; La = lacustrine systems; C = coal seams; P = paleosols; IV = incised valleys.

(longitudinal) drainage system which is channelized along the subsiding foredeep that appears ‘‘underfilled’’ relative to the uplifted orogen and forebulge that flank it on both sides (Jordan, 1995). The actual direction of flow depends on the direction of tilt in the basin, which in turn is controlled by the strike variability in orogenic loading. Such axial drainage systems follow the axis of the foredeep, and receive tributaries from both the orogen and the forebulge (Jordan, 1995). Isostatic rebound during orogenic unloading modifies the topographic profile in such a way that the dominant drainage system becomes transversal across the proximal foreslope (Jordan, 1995). In this case, the dominant flow may only become longitudinal along the axis of the topographic foresag (Fig. 7). Such a change from a proximal to a distal axial flow has been documented for the Pliocene section of the Himalayan foreland system of India, based on the shifts in the position of the Ganges river system (Burbank, 1992).

4. Discussion and conclusions

4.1. First-order foreland cycle

The life span of a retroarc foreland system starts and ends with events of inversion tectonics at the associated continental margin, from extension to compression and from compression to extension respectively. For example, in the case of the Karoo, the change from a divergent continental margin to a convergent plate margin took place towards the end of the Early Carboniferous (Smellie, 1981; Johnson, 1991; Mpodozis and Kay, 1992; Visser, 1992), and the end of the compression and hence foreland system evolution was marked by the initiation of the Gondwana breakup in the Middle Jurassic (ca. 183 My ago; Duncan et al., 1997). The first-order foreland cycle thus lasted for about 117 My (300–183 My). As extension on one side of the globe needs to be compensated by compression elsewhere, the subduction and accretion of terrains leading to the initial deformation of the Western Canadian Cordillera started in the early Middle Jurassic (ca. 180 My ago) and lasted until the onset of extension in the Early Eocene (ca. 55 My ago; Monger, 1989). The Western Canada foreland system cycle thus lasted for about 125 My (180–55 My).

Accommodation during such first-order foreland cycles is primarily controlled by the interplay of orogen-driven flexural tectonics and subduction-induced dynamic subsidence. The rates of flexural uplift of the peripheral bulge during stages of orogenic loading may be considered as approximately constant during the evolution of the basin, assuming that each thrust sheet brings about the same mass of supracrustal load. In contrast, the rates of dynamic subsidence increase following the onset of subduction, and decrease towards the end of the compressional cycle (Fig. 17). This allows the inference of a general trend in the evolution of accommodation across the foreland system, with (1) early dominance of flexural tectonics (underfilled foredeep with the formation of the forebulge unconformity); (2) mid-cycle dominance of dynamic subsidence (system-wide sedimentation, with the forebulge lowered below base level: late underfilled and filled stages); and (3) late dominance of flexural tectonics (overfilled stage, with sedimentation restricted to one flexural province at a time) (Fig. 17). Cessation of dynamic loading following the end of the foreland cycle leads to continental scale uplift (Fig. 17; e.g., Pysklywec and Mitrovica, 1999).

While the rates of flexural uplift may be considered constant during the first-order foreland cycle (Fig. 17), the periodicity of orogenic pulses changes through time following the same pattern as the curve that describes the change in the rates of dynamic subsidence. In other words, flexural cycles are longer in time in the early and late stages of foreland basin evolution (less frequent terrain accretion events at the convergent margin in

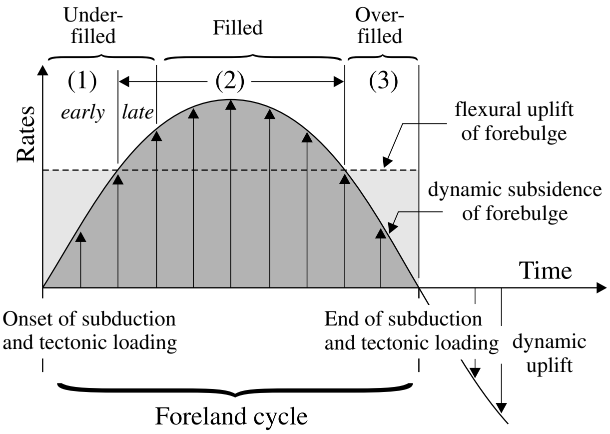

Fig. 17. Relative balance between the rates of flexural uplift and dynamic subsidence during a first-order foreland cycle, based on the case studies of the Karoo and Western Canada foreland systems. While the rates of flexural uplift may be considered constant during the first-order foreland cycle (i.e., each thrusting event brings approximately the same mass of supracrustal load), the periodicity of orogenic pulses changes through time following the same pattern as the curve that describes the change in the rates of dynamic subsidence—see text for more details. The basal (forebulge) unconformity forms during stage (1) of early underfilled state, when the rates of flexural uplift outpace the rates of dynamic loading. Stage (2) reflects an overall dominance of dynamic loading over flexural tectonism, whereas stage (3) corresponds to the decline in dynamic loading, when the system is again dominated by flexural tectonism. Stage (2) tends to start with a late underfilled state, followed by the filled phase of equilibrium between the rates of sedimentation and base level rise. The cessation of dynamic loading at the end of the foreland cycle causes long-wavelength lithospheric rebound, i.e. dynamic uplift.

response to lower subduction rates), and occur with a higher frequency in the middle stage when subduction and dynamic loading are most active. For example, flexural cycles in the filled Western Canada foreland system occurred over time scales of less than 1 My (Catuneanu et al., 1997b), whereas the overfilled stage of the basin recorded flexural cycles over 1 My in duration (Catuneanu and Sweet, 1999). The underfilled and overfilled stages of the Karoo Basin also recorded a flexural cyclicity of over 1 My in duration (Catuneanu et al., 1998).

This general scenario of evolution of a foreland system is supported by the case studies of the Karoo and Western Canada basins. Simplified versions of this scenario may also occur, when the rates of dynamic subsidence are never high enough to outpace the rates of flexural uplift. In such cases (e.g., the Witwatersrand Basin of South Africa; Catuneanu, 2001), the forebulge region does not preserve a stratigraphic record, and sedimentation is restricted to the foredeep and the back-bulge flexural provinces. Irrespective of the relative rates of dynamic subsidence, a common theme among all foreland systems emerges, which is that at least the early and late stages of evolution are always dominated by flexural tectonics. For this reason, basal (forebulge) unconformities always form at the onset of tectonic loading, while the foredeep is underfilled because of the low sediment supply. As the orogenic wedge rises above the sea level and sediment supply increases, the foredeep becomes filled and then overfilled towards the final phases of foreland basin evolution.

4.2. Higher-frequency foreland cycles

Superimposed on the first-order foreland cycle, shorter-term changes in the balance between accommodation and sedimentation, and implicitly in the depositional systems that dominate the foreland system, occur as a result of the interplay of flexural tectonics, dynamic subsidence, eustasy and sediment supply. This explains the manifestation of second- or lower-order transgressions and regressions of the interior seaways, which in effect represent high-frequency changes between filled (shallow marine) and overfilled (nonmarine) conditions.

System-wide sedimentation of wedges that show a uniform depositional character (e.g., progradational or retrogradational) is possible when sedimentation consistently outpaces or is outpaced by the rates of base level rise across the entire foreland system. This describes a situation where the rates of sedimentation and base level rise are off-balance, varying within ranges that do not overlap. Often though, the rates of sedimentation and base level rise vary within the same range, and this delicate balance leads to the formation of reciprocal (out-of-phase) stratigraphies across flexural hingelines (Catuneanu et al., 1999a). Two styles of reciprocal stratigraphies have been defined, for the filled and overfilled stages of basin evolution, and both refer to sequences with a cyclicity controlled by flexural tectonism. The filled style of reciprocal stratigraphies requires base level rise across the entire foreland system (dynamic subsidence or sea level rise > flexural uplift), and the coeval backstepping and progradation of the proximal and distal shorelines of the interior seaway (Fig. 13). The overfilled style of reciprocal stratigraphies requires the dominance of flexural tectonism over the long-wavelength controls on accommodation (dynamic subsidence or sea level rise), which results in a half-system type of sedimentation (Figs. 14 and 16).

The recognition of reciprocal stratigraphies requires good time control, and can be used to reconstruct the history of thrusting and off-loading in the adjacent fold-thrust belt. Each stratigraphic sequence showing a reciprocal pattern across the flexural hingeline corresponds to a flexural cycle of orogenic loading and unloading, and the periodicity of such cycles may be both > and <1 My, as shown by modeling and case studies. Reciprocal stratigraphies can also be used to map the geographic location of flexural hingelines, as well as to monitor their migration through time. Based on such stratigraphic criteria, flexural hingelines often seem to be controlled by underlying basement structures, which explains the mismatch that may be observed between the expected (theoretical) wavelengths and the real extent of flexural provinces in the field (Fig. 5).

4.3. Migration of foreland systems through time

Foreland systems prograde or retrograde through time in response to the redistribution of orogenic loads (Beaumont et al., 1993; Crampton and Allen, 1995; DeCelles and Giles, 1996). As a general trend, foreland systems prograde (shift towards the craton) during early stages of evolution, and tend to retrograde during their late stages. The progradation of a foreland system is a direct response to the progradation of the orogenic load as thrusting proceeds (Fig. 18). The rates of progradation depend on the dynamics of the orogenic belt, and the actual distance of progradation ranges from less than 100 km (Catuneanu, 2001) to 200 km or more (see Catuneanu et al., 2000, for a discussion on rates and amounts of progradation in the case of the Western Canada foreland system). The retrogradation of a foreland system may be attributed to piggyback thrusting accompanied by a retrogradation of the center of weight within the orogenic belt during orogenic loading, or to the retrogradation of the orogenic load via the erosion of the orogenic front during times of orogenic unloading (Catuneanu et al., 1998). The retrogradation of the center of weight in the fold-thrust belt may also be caused by the redistribution of load in relation to stages of extension or transtension, as recorded in the final phase of evolution of the Canadian cordillera (Price, 1994; Catuneanu et al., 2000).

Independent of tectonic loads, the retrogradation of flexural hingelines may also be caused by the visco-elastic relaxation of the lithosphere through time (Fig. 19; Beaumont et al., 1988, 1993). As indicated by modeling, a lithosphere that is subject to supracrustal loading tends to relax stress, leading to the deepening and narrowing of the foredeep with time. This type of visco-elastic relaxation operates over time scales that are larger relative to the cyclicity of flexural loading–unloading cycles (Beaumont et al., 1988, 1993).

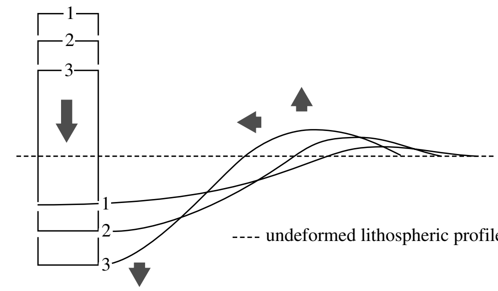

Fig. 19. Flexural response of a visco-elastic lithosphere that relaxes stress (modified from Beaumont et al., 1988, 1993). The diagram shows static loading in the fold-thrust belt, which leads to foredeep subsidence and forebulge uplift. At the same time, the foredeep becomes deeper and narrower with time (time steps 1–3) in response to the visco-elastic relaxation of the lithosphere.

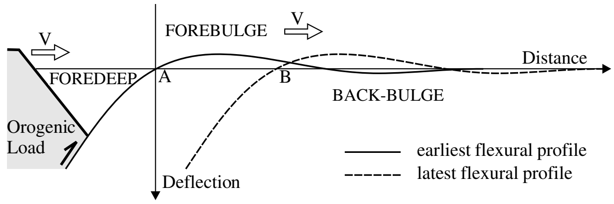

Fig. 18. Progradation through time of a foreland system in response to the progradation of orogenic load (modified from Beaumont et al., 1993; Crampton and Allen, 1995; DeCelles and Giles, 1996). "V" refers to the rate of progradation. Vertical scale is exaggerated for clarity.

An example of a migrating foredeep basin is offered by the case study of the Western Canada foreland system (Fig. 20). The location of flexural hingelines at different time steps in the evolution of the basin was mapped based on stratigraphic criteria (Catuneanu et al., 2000). The tectonic regime of dextral transgression is known from independent structural studies in the cordilleran belt (e.g., Price, 1994). The foredeep prograded to the east during the Campanian–Maastrichtian interval in response to the progradation of orogenic load, and retrograded in the Paleocene as a result of the tectonic load redistribution that accompanied the transition from compression to extension in the fold-thrust belt. Visco-elastic relaxation of the lithosphere may have also contributed to the observed retrogradation pattern. At the same time, the tectonic regime of dextral transgression led to an oblique direction of thrusting, which resulted in the northward migration (along strike) of the foredeep basin. It is also important to note that foredeeps tend to have a limited lateral extent along strike, as flexural hingelines curve around the locus of maximum tectonic loading at any given time (Fig. 20).

As a matter of first-order trends, it is conceivable that the progradation of a foreland system tends to be most rapid during its early stages of evolution, when the mass of the tectonic load is still low and hence the rates of thrusting are highest. In contrast, the rates of visco-elastic relaxation are expected to increase with time in response to repeated stress fluctuations (gradual weakening of the lithosphere) and increased sedimentary load in the basin. These opposite trends are illustrated in Fig. 21, and explain why most foreland systems are similar in terms of migration patterns, recording initial progradation and final retrogradation during their evolution.

The first-order progradation of the forebulge (stage 1 in Fig. 21) results in the formation of the forebulge (basal) unconformity, commonly found at the base of

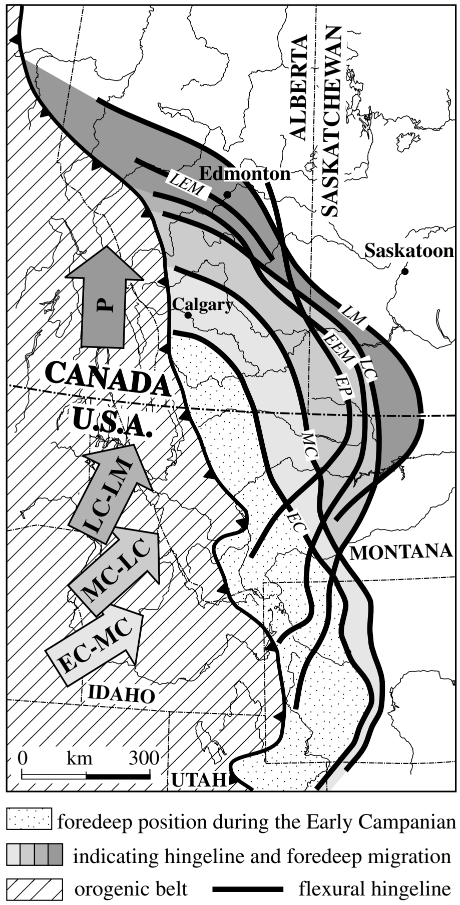

Fig. 20. Foredeep position during consecutive time slices of the Early Campanian–Early Paleocene interval in the evolution of the Western Canada foreland system (modified from Catuneanu et al., 2000). The arrows in the orogenic belt indicate the main direction of load shift during the Early Campanian–Middle Campanian (EC–MC), Middle Campanian–Late Campanian (MC–LC), Late Campanian–Late Maastrichtian (LC–LM), and Paleocene (P) intervals. This stage was dominated by dextral transgression in the Canadian cordillera, which resulted in the northward migration of the foredeep. Trends of progradation and retrogradation of the flexural hingeline may also be observed along dip.

foredeep deposits (Fig. 6). The age-equivalent foredeep strata of this hiatus are generally overthrusted and incorporated within the structures of the orogenic belt. In the case of the Karoo Basin, the early shift of the forebulge during the 330–300 My interval was likely in excess of 450 km, as no sediments of this age are preserved in the Cape Fold Belt or in the undeformed basin to the north. As documented by stratigraphic studies of the preserved Karoo Basin, the northward migration of the forebulge continued after the Late Carboniferous with another ca. 200 km during the Permian (Catuneanu et al., 1998). A decrease in progradation rates may therefore be inferred, from at least 15 km/My during the Carboniferous, to maximum 5 km/My towards the end of the Permian.

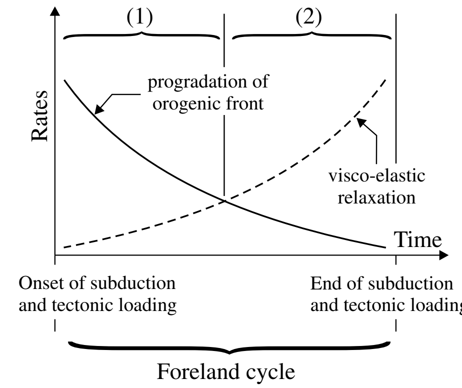

Fig. 21. First-order trends in the evolution of retroarc foreland systems. The rate of orogenic front progradation decreases through time as the mass of the orogen increases. The visco-elastic relaxation of the lithosphere increases through time, leading to an accelerated narrowing and deepening of the foredeep. As the balance between these two opposite trends changes through time, the foreland cycle may be subdivided into (1) a first-order stage of forebulge progradation, when the forebulge (basal) unconformity forms; and (2) a first-order stage of forebulge retrogradation, when the foredeep accumulates increasingly thicker sequences with time.

The first-order retrogradation of the forebulge (stage 2 in Fig. 21) in response to the gradual increase in the rates of visco-elastic relaxation is accompanied by a corresponding increase in the rates of creation of accommodation in the foredeep. As a result, the younger foredeep sequences tend to be thicker, as observed in the Western Canada Basin. In this case study, the thickening upward trend of foredeep sequences records a change of one order of magnitude, from the 10^0–10^1 m thick sequences of the filled stage (Catuneanu et al., 1997b) to the 10^1–10^2 m thick sequences of the overfilled stage (Catuneanu and Sweet, 1999).

4.4. Phanerozoic vs. Precambrian foreland systems

Foreland systems irrespective of age are the product of the same allogenic mechanisms, which control their formation and evolution through time. Similarities are therefore expected in terms of shapes, stages of evolution, and stratigraphic architectures. Two important differences are however related to the age of the underlying lithosphere relative to the age of the basin, and the dynamics of plate tectonic processes at the time of basin formation.

One of the oldest and best preserved Precambrian retroarc foreland systems in the world is the Late Archean (ca. 3.0–2.8 Gy) Witwatersrand Basin of South Africa. The sedimentary fill of the foredeep shows the same wedge-shape geometry as all Phanerozoic counterparts, as well as the typical transition from underfilled (West Rand Group) to overfilled (Central Rand Group) stages (Catuneanu, 2001). The wavelength of the flexural profile is however much shorter (about half) compared to Phanerozoic basins like the Karoo or the Western Canada, suggesting a different lithospheric rheology at the time of formation. This rheological contrast is due to the fact that the Witwatersrand Basin developed on a young (newly cratonized), hotter and hence less rigid, underlying lithosphere (Catuneanu, 2001). This relatively short wavelength may therefore be a distinctive feature of all Precambrian foreland systems, as they formed on young visco-elastic plates with a finite relaxation time.

The second major factor that influenced the evolution of Precambrian foreland systems is the erratic character of plate tectonic dynamics in its early stages (Eriksson and Catuneanu, 2004). The interaction between plate tectonics and mantle plumes was a unique first-order control on Precambrian continental drift, and caused the rates of extension or compression to vary significantly within a wide range, generally much larger than the one known for the Phanerozoic. The rates of subduction are proportional to the magnitude of dynamic loading, and hence are directly relevant to the evolution of any retroarc foreland system. These rates are generally considered to have been higher in the Precambrian, due to the higher geothermal gradients and the more rapid convection of sublithospheric mantle cells. This would tend to lead to increased rates of dynamic subsidence in the Precambrian retroarc foreland settings. Mantle plumes however also interfered with the dynamics of the early Earth, and often led to the "suspension" of continents over the uprising mantle material (Eriksson and Catuneanu, 2004). This suspension would temporarily freeze plate tectonic processes, leading to a decrease in the rates of subduction and hence a cessation of dynamic loading. This seems to be the case with the Witwatersrand foreland system, where the forebulge region was probably never lowered below the base level, being represented in the rock record by an unconformity (Catuneanu, 2001).

Acknowledgements

Financial support during the completion of this work was provided by the University of Alberta and NSERC Canada. My knowledge and ideas about the geology and evolution of foreland systems has benefited considerably from discussions and collaboration with Andrew Miall, Chris Beaumont, Glen Stockmal, Jerry Mitrovica, Art Sweet, and many others. Feedback from Andrew Miall, Art Sweet and Pat Eriksson helped to improve earlier versions of this manuscript. I also wish to thank Pat Eriksson for his exceptional editorial support.

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