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Reciprocal stratigraphy of the Campanian–Paleocene Western Interior of North America

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Abstract

The Claggett, Bearpaw, and Cannonball successions (Campanian–Paleocene) of the Western Interior foreland system were deposited during a series of regional second- and third-order transgressions and regressions, locally modulated by fourth- and lower-order cycles. The reciprocal flexural behaviour of the foreland lithosphere in response to orogenic cycles of loading and unloading results in out of phase sequences on opposite sides of the basin, which defines the concept of "reciprocal stratigraphies". The proximal and distal regions of the basin are separated by a few-kilometre wide "hinge zone" of proximal to distal facies change, with the mid point taken as a stratigraphic "hinge line", which defines the boundary between coeval marine transgressive and regressive systems tracts, or between nonmarine sequences and correlative sequence boundaries. Hinge lines have been mapped in Alberta, Saskatchewan, Montana and Wyoming for consecutive time-slices during the Campanian–Paleocene interval. Areally the hinge line traces a semielliptical pattern outlining the region of maximum foreland basin subsidence. It migrated at ~10 km/Ma to the north in response to Late Cretaceous–Paleocene orogenic dextral transpression. The hinge line also migrated eastward during the Campanian and Maastrichtian in response to thrust-sheet advance, and westward during the Paleocene in response to the redistribution of orogenic load associated with a transition from transpression to transtension, possibly accompanied by a visco-elastic deepening and narrowing of the foredeep.

Keywords: foreland system; reciprocal stratigraphy; hinge line; flexural tectonics; dynamic loading; foredeep migration

1. Introduction

1.1. Research objectives

We examine the large-scale stratigraphic patterns and regional sequence architecture that developed within the Campanian–Paleocene stratigraphic section of the Western Interior, during the Claggett (Early-Middle Campanian), Bearpaw (Late Campanian–Maastrichtian) and Cannonball (Paleocene) second-order transgressive–regressive cycles (Figs. 1 and 2). We also study: the relationship between Cordilleran tectonics and the associated foreland basin fill; the controls on the development of stratigraphic sequences and distribution of depositional environments; the differentiation between the foredeep and forebulge settings based on stratigraphic contrasts; the shift with time in the position of the stratigraphic hinge line; and the applicability of the

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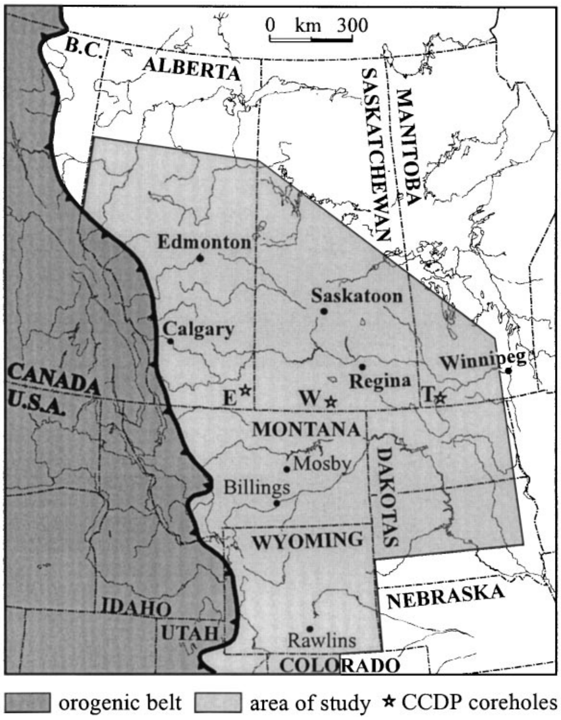


Fig. 1. Area of study. E, W and T: Elkwater, Wood Mountain and Turtle Mountain CCDP coreholes.

available hierarchical systems in the classification of sequence boundaries developed within retroarc foreland basins.

Our own data base and results derive mainly from strata accumulated within the Canadian part of the Western Interior (Western Canada Sedimentary Basin) during the Bearpaw and Cannonball cycles (Late Campanian–Paleocene interval). In addition, we also make use of the ammonite and shoreline data base of Gill and Cobban (1973) for the Claggett and Bearpaw cycles in the US portion of the Western Interior. This not only enlarges the regional perspective of our study, but also allows verification of our interpretation by matching results obtained from completely independent data and research techniques.

1.2. Flexural partitioning of foreland systems

Foreland basins form through the flexural deflection of the lithosphere in response to a combination of supra- and sublithospheric loads (Beaumont, 1981; Jordan, 1981; Jordan and Flemings, 1991; Sinclair et al., 1991; Sinclair and Allen, 1992; Watts, 1992; Beaumont et al., 1993; DeCelles and Giles, 1996). Four depozones are currently recognized within a foreland system, i.e. the wedge-top, foredeep, forebulge and back-bulge (DeCelles and Giles, 1996; Fig. 3). This partitioning of the foreland system develops in response to active loading in the orogenic belt. The magnitude of the flexural deflection is proportional to the degree of loading, as suggested in Fig. 3.

Fig. 4 illustrates the evolution of both flexural and surface profiles of a foreland system during successive stages of orogenic loading and unloading. The shape of the flexural deflection resembles a sine curve with the magnitude decreasing exponentially away from the orogenic load. The wavelength of the flexural deflection extends from the load's centre of weight to the depoaxis of the back-bulge (Fig. 4A). The flexural response of the foreland system to stages of orogenic unloading is a mirror-image of the flexural behaviour during loading stages (Beaumont et al., 1993; Fig. 4A). During both loading and unloading stages, the flexural foredeep and forebulge settings experience out of phase flexural responses relative to a flexural hinge line. The accommodation created through flexural deflection is gradually consumed by sedimentation, in a tendency to bring the surface profile of the foreland system back to the elevation of the adjacent craton (taken as a "datum" in Figs. 3 and 4A). Both the "datum" and the foreland surface profile adjust to the same base-level, reason for which they may all eventually become superimposed. As the surface profile approaches the base-level, the topographic elevation of the foreland system becomes alternatingly higher in the distal and proximal sectors during orogenic loading and unloading, respectively (Fig. 4B).

In the case of retroarc foreland basins, the foreland subsidence is controlled by the combined effect of static and dynamic loads (Catuneanu et al., 1997a). If the subduction taking place underneath the basin is rapid and/or at a shallow angle, the long-wave lithospheric subsidence triggered by dynamic/sub-lithospheric loading generates accommodation space and preservation potential in both sectors of the basin. If subduction takes place slowly or at steep angle, sediments are deposited and preserved only in the

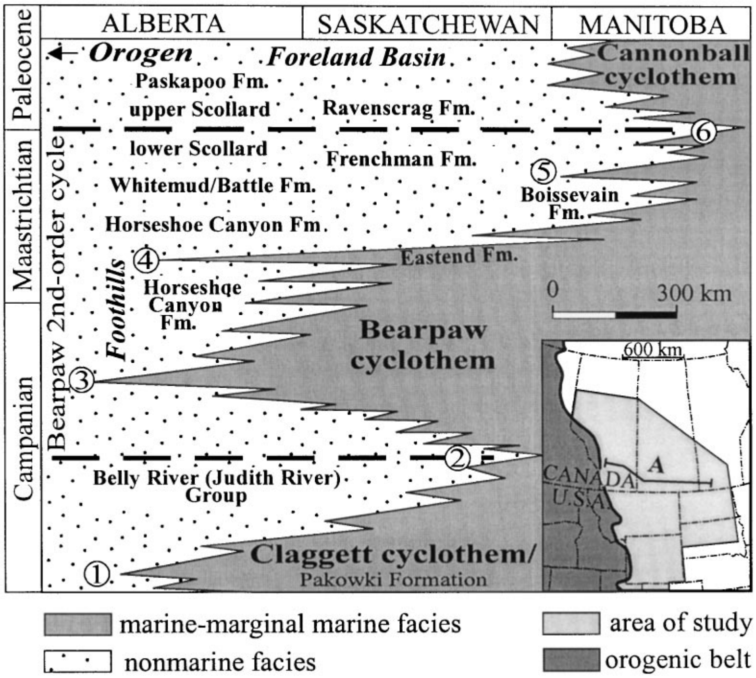


Fig. 2. Schematic cross-section of Latest Cretaceous and earliest Paleogene marine cyclothems (profile A in the inset map), without stratigraphic gaps shown. Pattern and age of the cyclothems are based on data derived in this study supplemented by Gill and Cobban (1973, Figs. 16 and 17) and Cherven and Jacob (1985, Fig. 3). 1. *Baculites mclearni*: maximum flooding surface of Claggett cycle (maximum transgression of Claggett Sea); 2. *Baculites scotti*: lower boundary of Bearpaw cycle (maximum regression of Claggett Sea); 3. *Baculites compressus*: maximum flooding surface of Bearpaw cycle (maximum transgression of Bearpaw Sea); 4. *Baculites grandis*: Drumheller Marine Tongue, a lower-order T–R cycle; 5. *Early Late Maastrichtian* (proximal, mostly subaerial unconformity—distal, Coulter Member of Pierre Formation or Breien Member of Hell Creek Formation in North Dakota); 6. *Late Late Maastrichtian-earliest Paleocene*: upper boundary of Bearpaw cycle (maximum regression of Bearpaw Sea).

foredeep, and an unconformity develops over the distal sector (Mitrovica et al., 1989; Waschbusch et al., 1996; Catuneanu et al., 1997a). The width of the foreland basin fill along dip is therefore a good indicator of the type of sub-lithospheric processes taking place during the foreland basin evolution.

The foreland system migrates with time in response to the redistribution of load within the orogenic belt. Cratonward migration of the foreland system is recorded during times of orogenic loading with the progradation of the orogenic front. Orogenward migration of the foreland system may be attributed to piggyback thrusting accompanied by a retrogradation of the centre of weight within the orogenic belt during orogenic loading, or to the retrogradation of the orogenic load through the erosion of the orogenic front during times of orogenic unloading (Catuneanu et al., 1998a). Superimposed on the effects of orogenic load redistribution, a long-term deepening and narrowing of the flexural foredeep may also occur due to the visco-elastic relaxation of the lithosphere (Beaumont et al., 1993). The amount of relaxation depends on the rheological properties of the lithosphere, and results in a gradual decrease in the flexural wavelength.

We focus our analysis on the stratigraphic consequences that may develop across the foreland system hinge line in response to the out of phase flexural tectonics of the proximal and distal regions. Mapping the position of the stratigraphic hinge line at consecutive time-steps may also allow conclusions regarding the relationship between the migration trends of the foreland system and the orogenic tectonics.

2. Geological background

2.1. Tectonic setting

The Western Interior is a classic example of a retro-arc foreland basin developed in front of the Western Cordillera of North America as a result of lithospheric flexure in response to tectonic loading in the adjacent convergent orogen (Jordan, 1981; Stockmal and Beaumont, 1987; Beaumont et al., 1993; Johnson and Beaumont, 1995). The compression generated through accretion of displaced terranes onto the Pacific margin of the North American continent controlled the amount of crustal shortening and thickening in the Cordilleran Orogen, and consequently the amount of subsidence in the foreland area. In addition, the orogenic belt also constituted a main source of sediments for the adjacent foredeep (Cant, 1989). The close relationship between the orogen and the foreland system requires a thorough understanding of both basin stratigraphy and orogenic tectonics.

Along-strike diachroneity of orogenic processes has been documented in the Canadian Cordillera (Cant and Stockmal, 1989; Price, 1994), demonstrating that stages of pre-loading, loading, and unloading have occurred simultaneously in different areas along the thrust-fold belt. Loading in one part of the orogen caused flexural subsidence in the neighbouring sector of the foreland basin contemporaneous with flexural uplift in other areas along the strike placed in a "pre-loading" or "unloading" stage (Beaumont et al., 1993). This implies that the foreland basin hinge line may not only be intersected along dip-oriented profiles but also along strike-oriented transects, which is an important conclusion as it explains why the foredeep should be expected to display a semielliptical shape rather than a linear one with the hinge line parallelling the orogenic front.

The stratigraphic section studied in this paper accumulated during Late Cretaceous–Paleocene dextral transpression in the Canadian Cordillera, dominated by strike-slip in the north and by compression in the

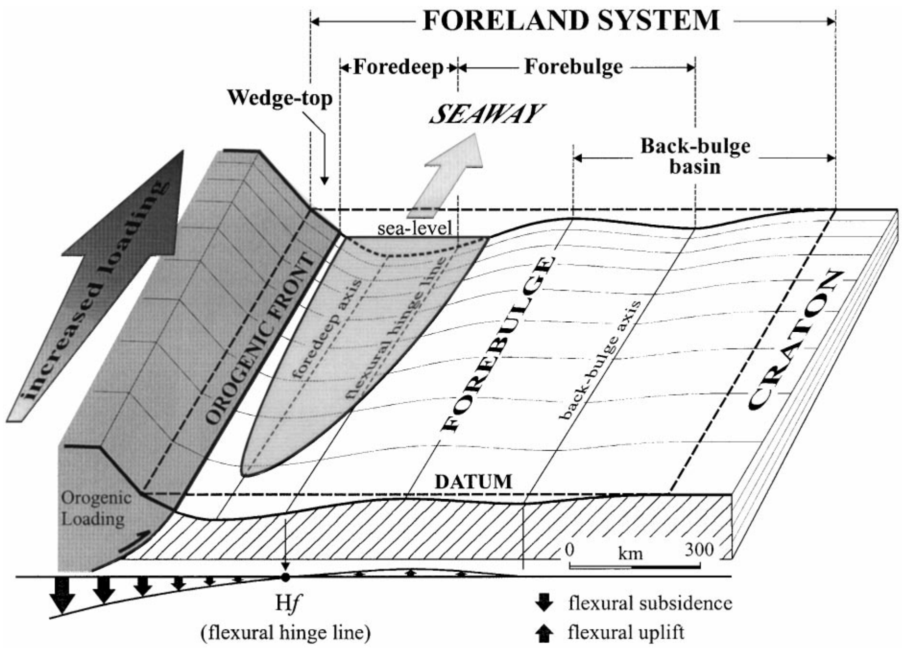


Fig. 3. Configuration of the foreland system during orogenic loading with strike variability. The magnitude of the flexural deflection is proportional to the degree of loading. Four depozones may be differentiated, i.e. wedge-top, foredeep, forebulge and back-bulge. We refer to the wedge-top and foredeep as the proximal sector, and to the forebulge and back-bulge as the distal sector. The proximal and distal sectors of the foreland system are separated by the flexural hinge line. The topographic elevation of the adjacent craton, approximated with a horizontal plane, is taken as a datum. The base-level of deposition within the foreland system 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.

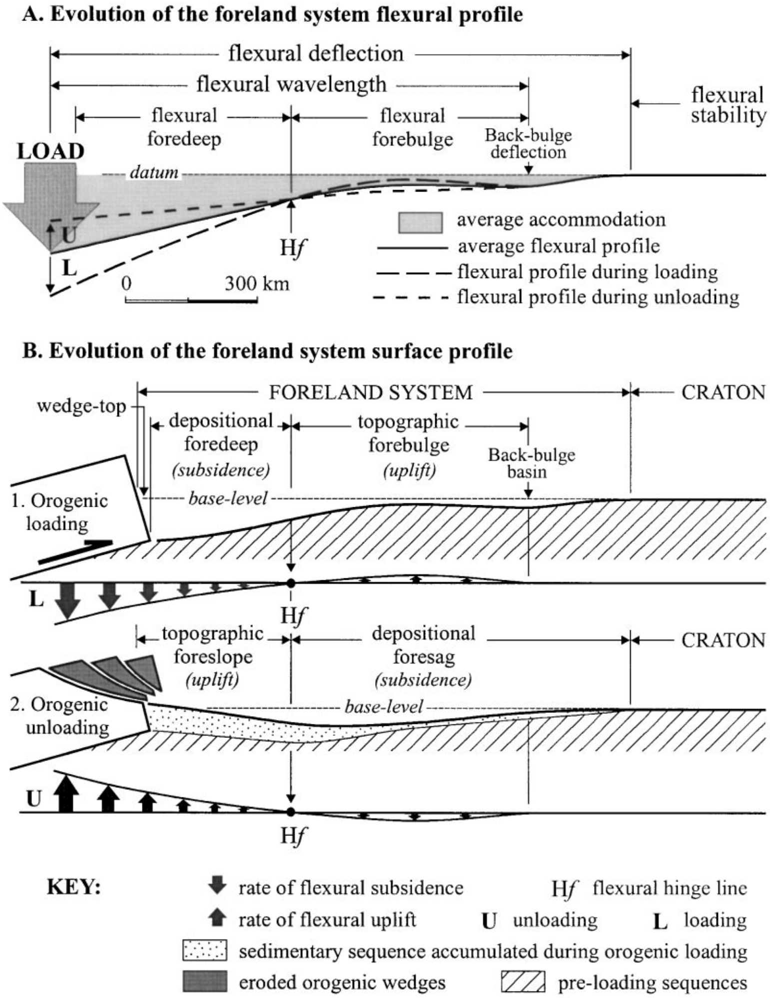


Fig. 4. Flexural and surface profiles illustrating the evolution of the foreland system during stages of orogenic loading and unloading (modified after Beaumont et al., 1993; Waschbusch et al., 1996; Catuneanu et al., 1997a; Catuneanu et al., 1999). Not to vertical scale. The flexural wavelength depends on the rheology and elastic thickness of the lithosphere, basement tectonics, and mass and distribution of the applied loads. The given horizontal scale suggests a continental lithosphere with high flexural rigidity. If the foreland system develops on a less rigid and fractured basement, the foredeep may be narrower than 150 km, such as in the case of the Alps molasse basins (Homewood et al., 1986; Crampton and Allen, 1995). (A) illustrates the flexural foredeep and forebulge undergoing out of phase subsidence and uplift in response to orogenic loading and unloading. During each flexural state, surface processes (sedimentation, erosion) tend to level the foreland topography to the elevation of the base-level (~datum), allowing the subsequent mirror-image rebound of the surface profile. In (B), the proximal foreland illustrates the depositional foredeep (loading case) and a topographic "foreslope" dipping away from the orogenic load (unloading case). The distal foreland represents the topographic forebulge (loading case) and the depositional foresag (unloading case). The depocenter of the foreland systems alternates between the depositional foredeep during orogenic loading, and the depositional foresag during orogenic unloading. The migration of the flexural hinge line due to the redistribution of orogenic load is not represented.

south (Price, 1994). Because the estimated crustal shortening in the south (150 km) is about three times greater than in the north (50 km), the tectonically induced slope determined a general southeastward sediment transport (Cant and Stockmal, 1989), from a distal-type (uplift-dominated bulge) to a proximal-type of tectonic setting (subsidence-dominated foredeep). The transition between the two tectonic regimes occurs at about 56N latitude (Price, 1994), which would roughly approximate a hinge line position for the Late Cretaceous–Paleocene.

2.2. Lithostratigraphy

Among the successive orogenward marine incursions of 10^7 year-episodicity that characterized the evolution of the Western Interior Seaway (Kauffman, 1969), the Claggett and Bearpaw transgressions were the last marine incursions to reach the present position of the foothills of Alberta (Fig. 2). In the last second-order transgressive–regressive cycle, corresponding to the incursion of the Cannonball seaway, the marine facies was restricted to the eastern portions of the Western Interior basin (Montana, North Dakota, southwestern Manitoba and possibly into southeastern Saskatchewan), with well-represented coeval nonmarine facies in the more proximal areas. As they are best represented in our area of study (Fig. 1), we focus our attention on the Claggett marine, Bearpaw marine, and post-Bearpaw nonmarine strata.

2.2.1. Pakowki/Claggett marine strata

The Pakowki Formation in southern Alberta is the correlative of the Claggett Formation of Montana and adjacent states. They overlie, respectively, the Milk River and Eagle formations with a fairly sharp contact and are transitionally overlain by the Belly River (Judith River) Group (Fig. 2). The Claggett consists largely of dark shales with interbedded sandstones, and has its type locality near the town of Judith in central Montana. Claggett shoreline positions and trends have been determined through ammonite occurrences and zonations (Gill and Cobban, 1973). Both the base and top of the Claggett Formation are highly diachronous (Gill and Cobban, 1973).

2.2.2. Bearpaw marine strata

The Bearpaw Formation is a westward-thinning lithostratigraphic unit consisting of marine silty shales and sandstones. The type section of the Bearpaw Formation is in the Bearpaw Mountains of north-central Montana, and it has been traced northward from Montana into adjacent parts of Alberta and Saskatchewan. Considering the overlying marginal-marine facies (Blood Reserve Sandstone, and Eastend and Foxhills formations) as part of the Bearpaw Sea-dominated strata, as did Nadon (1988), the Bearpaw is placed between two eastward-thinning dominantly nonmarine clastic wedges: the underlying Judith River Formation and Belly River Group, and the overlying St. Mary River/Horseshoe Canyon/Whitemud/Hell Creek formations (Fig. 2). Within the area of study, the Bearpaw Formation spans the time interval of up to 12 ammonite zones (*Baculites scotti-Baculites grandis*, Caldwell et al., 1993) and has diachronous boundaries. Maximum flooding occurs within the *Baculites compressus* zone (Gill and Cobban, 1973; Fig. 2).

2.2.3. Post-Bearpaw nonmarine strata

The table of formations in Fig. 5 summarizes nomenclature for the Maastrichtian and Paleocene of the Western Canada Basin. The lithostratigraphies reflect three aspects of the depositional environment. Fundamental to differences in the regional stratigraphies is the timing of the eastward withdrawal of the midcontinental (Bearpaw) seaway from the basin and the transgression and regression of the final seaway (Cannonball) along the eastern margin of the basin (Figs. 2 and 5). A second factor is regional differences in climate-controlled lithofacies. In southwestern Alberta, Maastrichtian and Paleocene rock units contain abundant caliche nodules and pedogenic hardpans whereas coal is absent except marginal to the Bearpaw seaway (Jerzykiewicz and Sweet, 1988). In central and north-central Alberta and southern Saskatchewan coal is common. The third factor is regional changes in the energy levels of transport systems and the availability of accommodation space.

3. Sequence stratigraphy and hierarchy

Sequence stratigraphic terminology used in this paper is illustrated in Fig. 6. The various types of sequences, systems tracts and bounding surfaces are defined in relationship to the relative sea-level (combined effect of tectonics and eustasy) and transgressive–regressive (combined effect of relative sea-level and sediment supply) curves. Within nonmarine successions we use the time gaps in the stratigraphic record to delineate the position of sequence boundaries, either referred to as “subaerial unconformities”, where evidences of subaerial exposure and erosion are

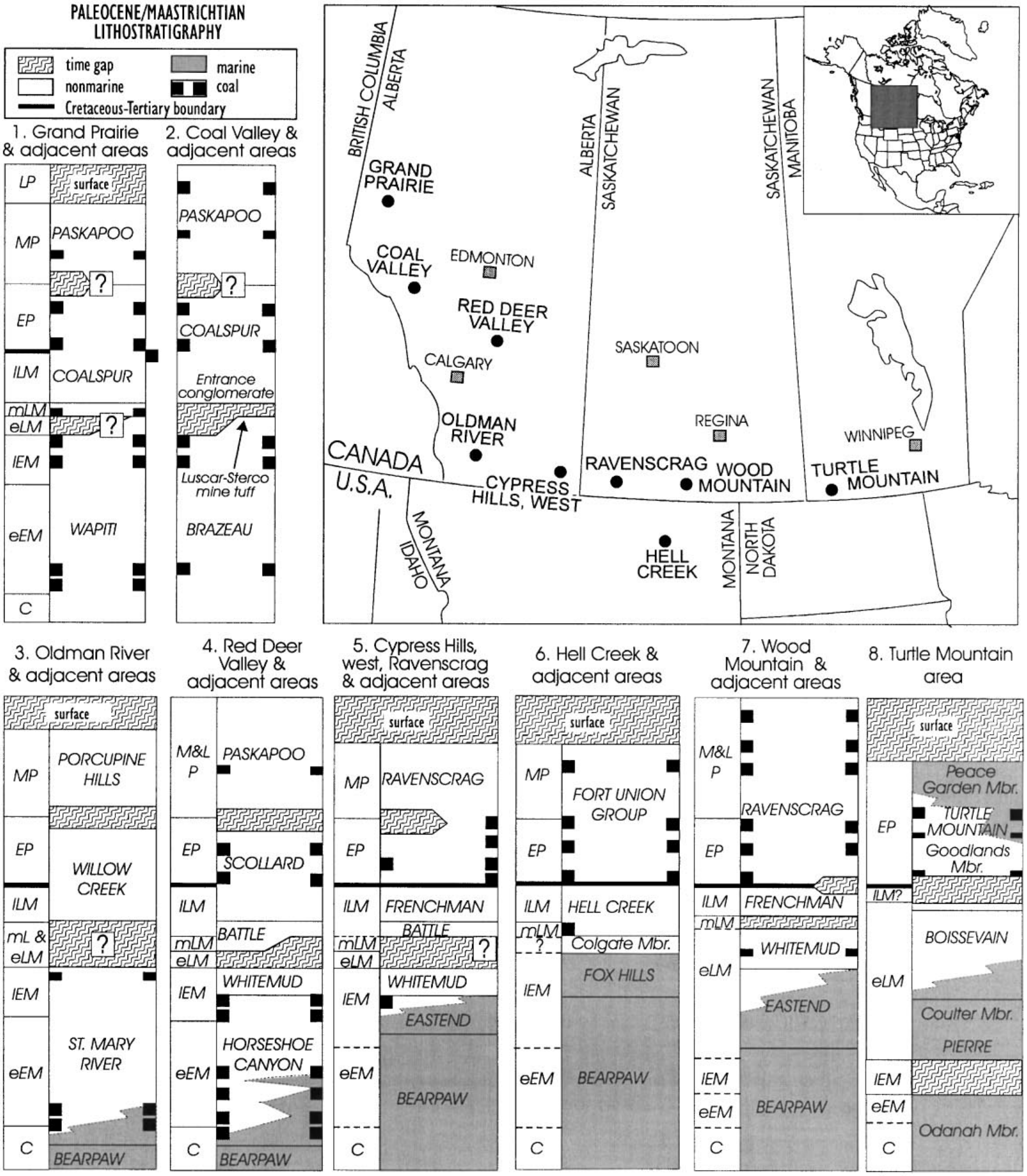
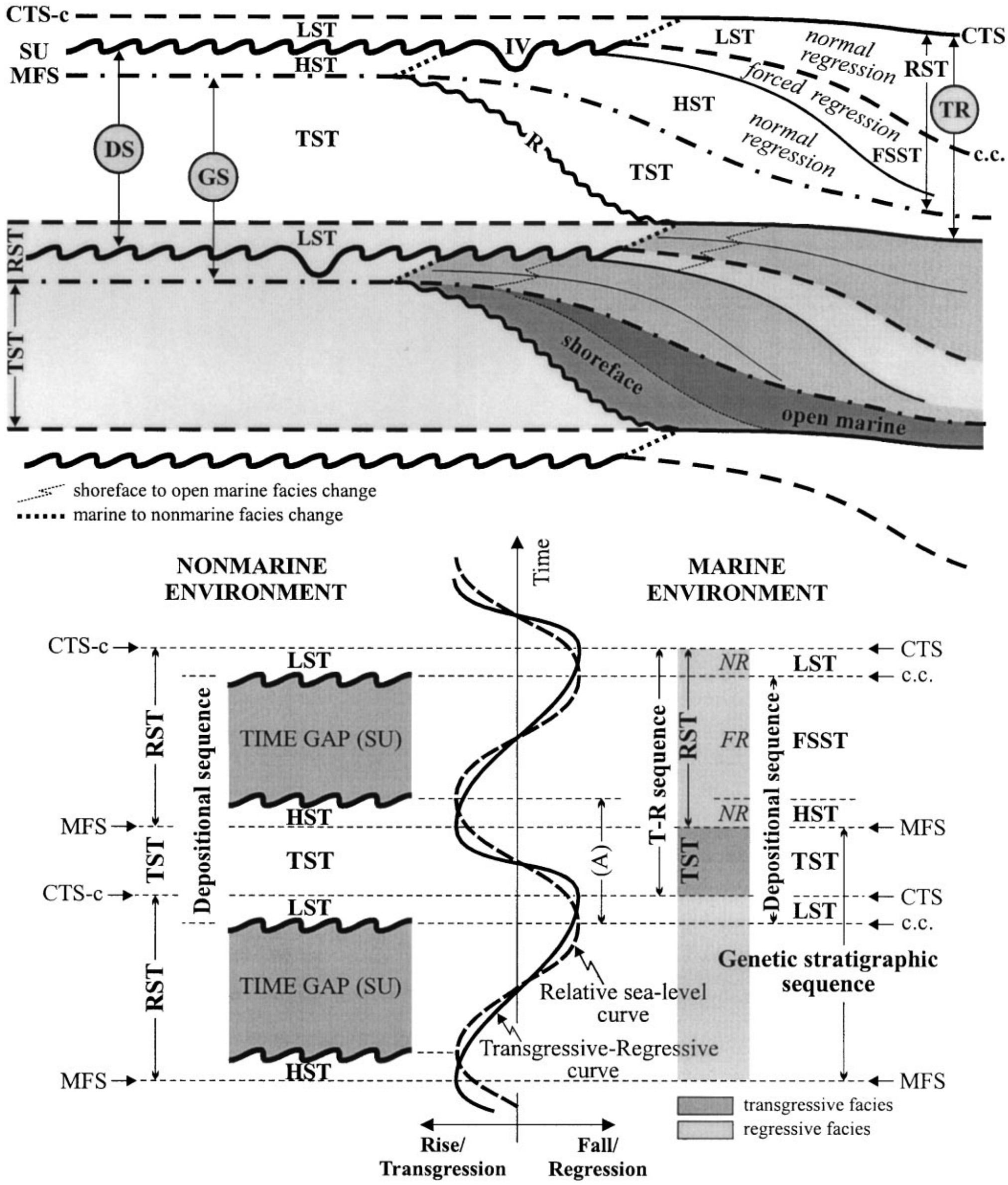


Fig. 5. Stratigraphies present in the Maastrichtian and Paleocene of the Western Canada Basin based primarily on: Column 1. Dawson et al. (1994); 2. Jerzykiewicz and McLean (1980), Jerzykiewicz and Sweet (1986, 1988) and Demchuk (1990); 3. Unpublished data (ARS) and as for two; 4. Gibson (1977), Lerbekmo et al. (1992), unpublished data (ARS); 5. Furnival (1946), Nambudiri and Binda (1991), Lerbekmo (1985) and unpublished data (ARS); 6. Shoemaker (1966), Archibald (1982), Robertson (1974) and unpublished data (ARS); 7. Furnival (1946), Byers (1969) and unpublished data (ARS); 8. Bamburak (1978) and unpublished data (ARS). The inferred ages are according to the above authors modified as necessary by the results of this and concurrent studies. Not to vertical scale.

apparent; or "paraconformities", where there is no direct evidence of erosional processes. Within marine successions we use the conformable transgressive surfaces as sequence boundaries, because they have field expression and can also be recognized on well logs (Embry, 1995). These two types of bounding surfaces are temporally offset by the duration of the lowstand systems tract overlying the subaerial unconformity (regression during early relative sea-level rise, due to sedimentation outpacing the rate of relative rise; Fig. 6). However, when erosion associated with the marine ravinement surface entirely removes the basinward portion of the nonmarine lowstand systems tract, the conformable transgressive surface may be in the lateral continuation of the subaerial unconformity.

A sequence hierarchy involves the separation of different orders of stratigraphic surfaces based on



the interpreted magnitude of base-level change that generated the surface. The higher the order of the surface the larger the base-level shifts involved in its generation. The highest order boundaries are designated as first-order and lower order ones are then designated as second, third, fourth, etc. The reader is referred to Embry (1993, 1995) for suggested criteria of differentiating various orders. For this study we have used the areal extent of the surface and the estimated base-level shift from facies analysis to determine the order of a surface.

4. Reciprocal stratigraphy of Bearpaw strata

4.1. Data base and results

The marine transgressive and regressive strata of the Bearpaw second-order cycle (Late Campanian–Early Maastrichtian) have been studied through facies analysis at the regional scale of the Western Canada Foreland Basin, in outcrop and subsurface (area of study shown on the inset map in Fig. 7). The reader is referred to Catuneanu (1996) and Catuneanu et al. (1997b) for a complete account of Bearpaw stratigraphy.

Our previous work showed that the Bearpaw Formation accumulated during a succession of third-order transgressive–regressive (T–R) cycles which led to the deposition of fining-upward transgressive and coarsening-upward regressive facies. These facies were identified as transgressive systems tracts (TST) and regressive systems tracts (RST), respectively, separated by conformable transgressive surfaces and maximum flooding surfaces (Catuneanu et al., 1997b). Two distinct types of third-order T–R sequences were recognized in the Bearpaw Formation, and each is found in a geographically distinct area. Proximal-type sequences are characterized by a thin TST and thick RST, and they occur in the southwestern part of the study area, proximal to the orogenic belt. Distal-type sequences are characterized by a thick TST and a thin RST and occur to the northeast in a distal sector. The available time control (laterally extensive bentonite beds and ammonite zonation) indicates that the conformable transgressive surfaces of the proximal sequences correlate with the maximum flooding surfaces of the distal sequences and vice versa. In other words, the thin proximal TST and distal RST, as well as the thicker proximal RST and distal TST have been found to be coeval (Catuneanu et al., 1997b; Fig. 7).

The change of facies between the proximal and distal sequence types takes place within a relatively narrow zone, less than 10 km wide (within the size of a township), which we term a hinge zone. The mid point of the hinge zone can be arbitrarily taken as a stratigraphic hinge line. By mapping the areal extent of the proximal and distal types of sequences, the hinge line can be traced trending SE–NW across the Alberta–Saskatchewan border (Catuneanu et al., 1997b).

4.2. Interpretation

The reciprocal architecture of Bearpaw T–R sequences is best explained by a tectonic control on the foreland basin stratigraphy through the flexural compensation of the lithosphere in response to

Fig. 6. Types of sequences, bounding surfaces and systems tracts, defined in relation to the base-level and transgressive–regressive curves (from Catuneanu et al., 1998b). The depositional sequence boundary (i.e. subaerial unconformity and its marine correlative conformity) is generated at the end of relative sea-level (base-level) fall (Posamentier et al., 1988; Van Wagoner et al., 1988). The genetic stratigraphic sequence boundary (i.e. maximum flooding surface) is taken at the top of marine and nonmarine transgressive facies (Galloway, 1989). The T–R sequence boundary is taken at the top of marine regressive facies (i.e. the conformable transgressive surface, Embry, 1995); within nonmarine facies, the T–R sequence is chosen to coincide with the depositional sequence, although the RST may extend above the SU, due to the difficulty in field recognition of the CTS-correlative. When the most basinward portion of the nonmarine LST is completely eroded by the ravinement surface during transgression, the CTS may be mapped in the lateral continuation of the SU in spite of the fact that the two surfaces are temporally offset. Abbreviations: DS = depositional sequence; GS = genetic stratigraphic sequence; T–R = transgressive–regressive sequence; TST = transgressive systems tract; RST = regressive systems tract; LST = lowstand systems tract; HST = highstand systems tract; FSST = falling stage systems tract; SU = subaerial unconformity; c.c. = correlative conformity; CTS = conformable transgressive surface; CTS-c = CTS-correlative (i.e. the nonmarine correlative of the marine CTS); MFS = maximum flooding surface; R = ravinement surface; IV = incised valley; (A) = creation of accommodation space (base-level rise); NR = normal (sediment supply-driven) regression; FR = forced (base-level fall-driven) regression.

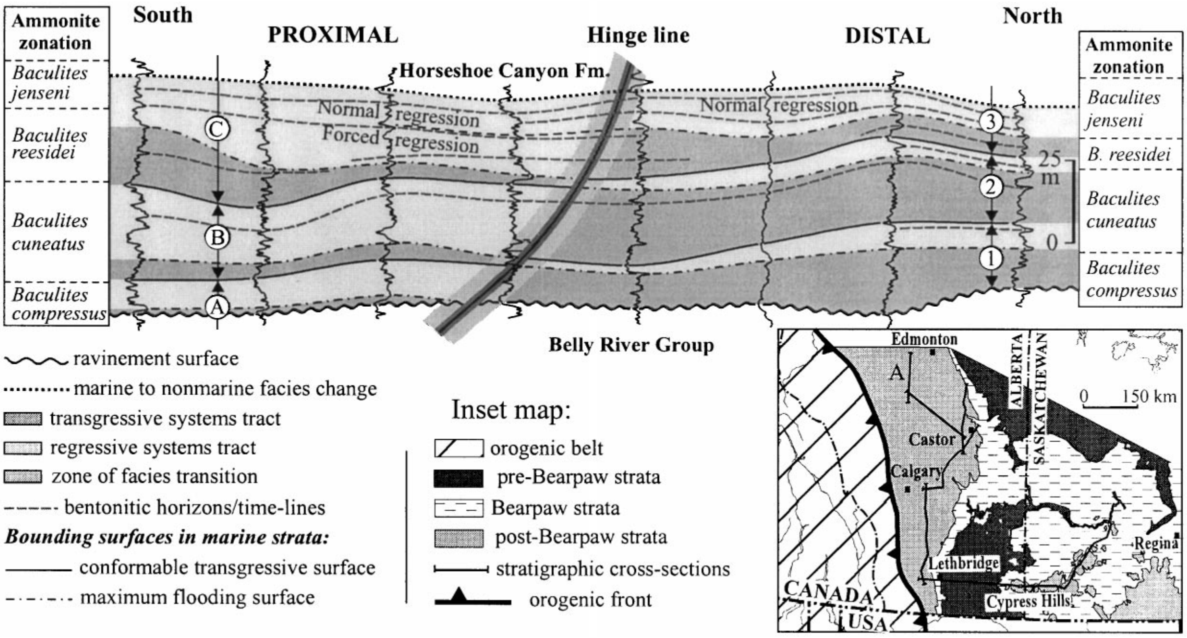


Fig. 7. Simplified stratigraphic cross-section of correlation of Bearpaw strata, south-west of Edmonton (profile A in the inset map), illustrating the facies changes in the transition (hinge) zone between the proximal and distal foreland settings. A complete account of Bearpaw stratigraphy, as well as the other stratigraphic profiles shown in the inset map, are provided in Catuneanu (1996) and Catuneanu et al. (1997b). A–C: proximal sequences; 1–3: distal sequences. Correlation based on gamma ray logs.

successive cycles of orogenic loading and unloading, i.e. opposite flexural tectonics between the proximal and distal sectors of the basin relative to the stable hinge line that separates them (Fig. 4). The interplay between flexural tectonics and dynamic subsidence is discussed in detail by Catuneanu et al. (1999). During orogenic compressional loading (pulse stage), coeval accumulation of proximal TST and distal RST takes place, driven by high versus low rates of base-level rise, respectively (Catuneanu et al., 1999). During erosional/extensional unloading in the orogenic belt (quiescence stages), the reverse situation occurs leading to the simultaneous deposition of proximal RST and distal TST. The fact that within distal sequences the RST are thinner, although they include coarser sediment, suggests that the time for TST deposition (orogenic unloading stage) was much longer than the time for RST accumulation (orogenic loading stage). Further discussions on the necessary conditions for the generation of distal-type sequences, dominated by thick TST, as well as on the position of sediment sources for proximal and distal deposition, are provided in Catuneanu et al. (1997b).

4.3. Boundary hierarchy within Bearpaw strata

The conformable transgressive surfaces and maximum flooding surfaces mapped within the Bearpaw second-order cyclothem may be classified into a fourfold hierarchy based on their areal extent (Catuneanu, 1996), which is interpreted to reflect the magnitude of base-level change that resulted in boundary generation (Embry, 1995). Third-order sequence boundaries develop throughout the entire proximal or distal sectors, and have correlative maximum flooding surfaces on the other side of the hinge line. Fourth-order boundaries may also develop in both sectors of the basin with correlative flooding surfaces over the hinge line, but are associated with smaller changes in base-level and therefore bound thinner sequences, which may have discontinuous areal development. Fifth- and sixth-order boundaries are confined to the proximal sector, with no distal correlatives. The

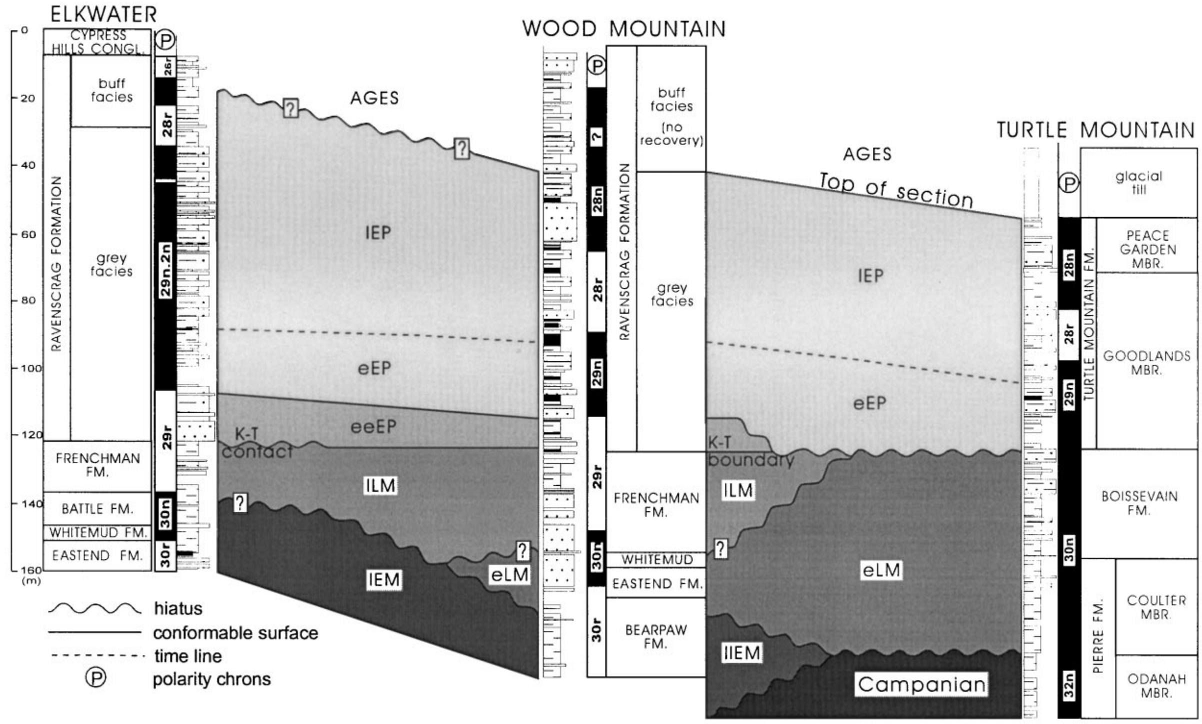


Fig. 8. Stratigraphic correlations established between the three Canadian Continental K–T project coreholes. Campanian age based on dinoflagellates (D.J. McIntyre, personal communication); *Late* Early Maastrichtian (lEM) on the presence of the *Scollardia trapaformis* Zone assemblage of Srivastava (1970) and *latest* early (llEM) on the presence of *S. trapaformis* in combination with *Wodehouseia spinata*; *Early* Late Maastrichtian on the combined presence of *W. spinata*, *Porosipollies porosus* and *Mancicorpus notabile*, *Late* Late Maastrichtian (lLM) on the combined presence of *W. spinata*, *Myrtipites scabratus*, and *Aquilapollenites delicatus* var. *collaris*; earliest Paleocene (eeEP) on the combined presence of *W. spinata* and *Aquilapollenites reticulatus* as described in Sweet (1978); Early Paleocene on the presence of *W. fimbriata* within polarity chron 29n; *Late* Early Paleocene (lEP) on the presence of *Momipites* spp., *Tricolporopollenites kruschii* and the associated polarity chrons; and Middle Paleocene on the presence of the *Aquilapollenites spinulosus* Zone assemblage of Demchuk (1990) in associated outcrop. Modified after Catuneanu et al. (1995); polarity chrons after Lerbekmo, personal communication.

interplay between tectonics and sediment supply best explains the stratigraphic patterns of the third- to sixth-order sequences. The relative importance between tectonics and sediment supply favours tectonics in the case of third- and fourth-order sequences, and gradually changes in favour of sediment supply towards the sixth-order cycles, which may be completely autocyclic and do not require base-level changes (Catuneanu, 1996).

5. Reciprocal stratigraphy of post-Bearpaw strata

5.1. Data base and results

Three coreholes, completed under the Canadian Continental Drilling Program (CCDP) (Fig.1; Elkwater in southeastern Alberta, Wood Mountain in south-central Saskatchewan and Turtle Mountain in southwestern Manitoba) provided lithostratigraphies, palynostratigraphies and polarity chronologies for Paleocene and latest Cretaceous strata (Fig. 8). Previously reported palynological assemblages (Snead, 1969; Srivastava, 1970; Sweet, 1978; Demchuk, 1990) and magnetostratigraphies (Lerbekmo, 1985; Lerbekmo and Coulter, 1985; Lerbekmo et al., 1992) have been added to the chronological framework for these coreholes. In composite these have allowed the distinction of the *Early* and *Late* Early Maastrichtian, *Early, Middle* (only in outcrop) and *Late* Late Maastrichtian, earliest Paleocene, *Early* and *Late* Early Paleocene, Middle

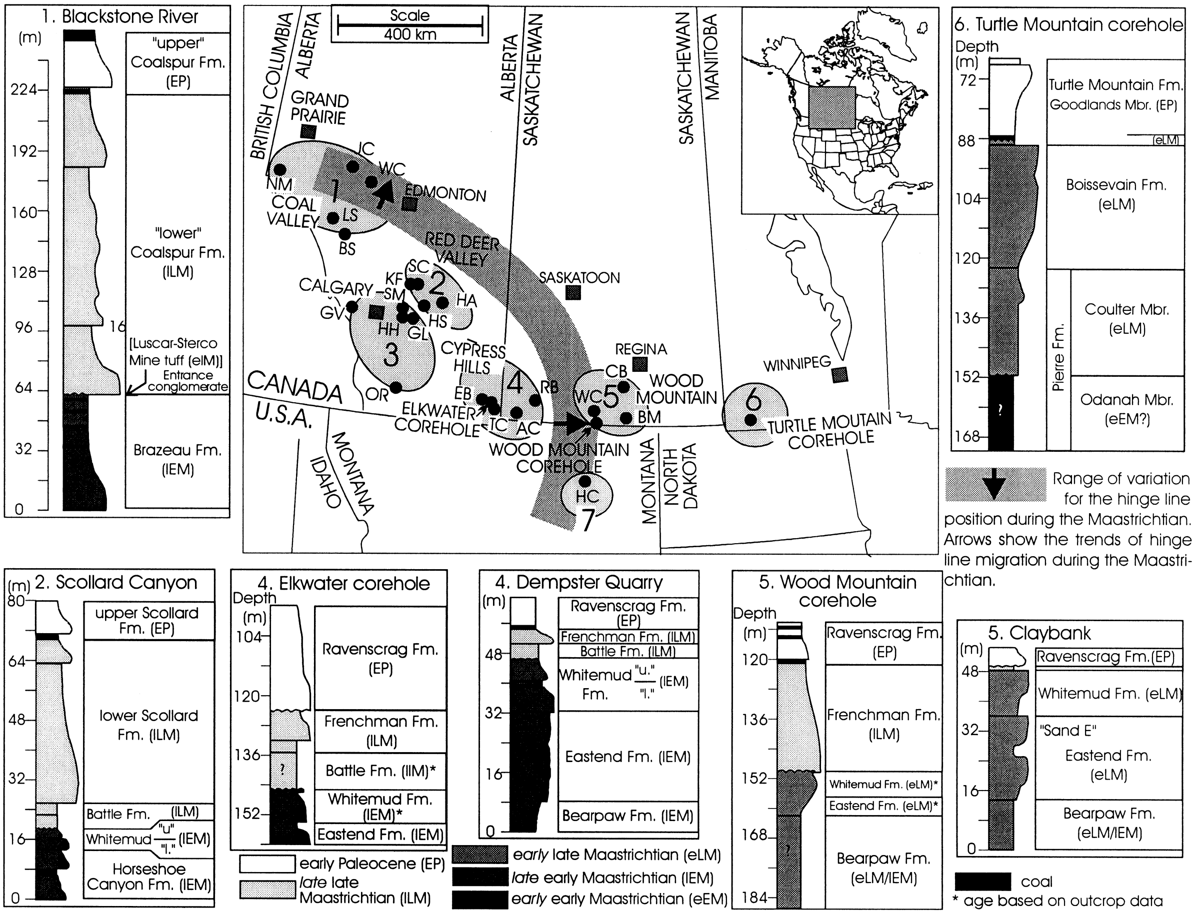


Fig. 9. Distribution of Whitemud/Battle sections and their correlatives grouped into seven areas (grey ovals). 1. Coal Valley area: Blackstone River (BS) reference section, modified after Jerzykiewicz (1992, Fig. 5); LS: Luscar Sterco mine; NM: Nose Mountain; JC: Judy Creek; and WC: Whitecourt. 2. Red Deer Valley and vicinity: Scollard Canyon (SC) reference section, modified after Gibson (1977, Fig. 3c); KF: Knudsen's Farm; HA: Hand Hills; HS: Horseshoe Canyon. 3. Calgary and vicinity: no designated reference section; SM: Strathmore; GL: Gliechen; HH: Hammer Hill; GV: Grand Valley Creek; and OR: Oldman River section near Fort Macleod. 4. Cypress Hills area: Elkwater corehole and Dempster Quarry (DQ) reference sections (the latter modified after Kupsch, 1957); EB: Eagle Butte; TC: Thelma Creek; AC: Adam Creek; RB: Ravenscrag Butte. 5. Wood Mountain area: Wood Mountain corehole and Claybank (CB) reference sections (the latter modified after A.R. Byers, 1959); WC: Wood Mountain Creek area; BM: Big Muddy. 6. Turtle Mountain area, Turtle Mountain corehole. 7. Hell Creek area (HC); locality north of Jordan, Montana (not illustrated).

Paleocene, and Late Paleocene (Figs. 5 and 8). Palynological assemblages diagnostic for these stratigraphic subdivisions are indicated in the description for Fig. 8.

Our surface and subsurface data come from localities that can be grouped into seven areas (ovals in Fig. 9): 1. Central Foothills and adjacent Plains area; 2. Red Deer Valley and vicinity; 3. Calgary and vicinity; 4. Cypress Hills area; 5. Wood Mountain area; 6. Turtle Mountain area; and 7. Hell Creek area. Beyond differences in thicknesses and lithofacies between areas or reference sections, which can be related to local paleo-environmental conditions or variations in sediment supply, we focused on the identification and dating of the major stratigraphic hiatuses present in the post-Bearpaw nonmarine succession. These hiatuses can be related to allocyclic controls such as changes in base-level. Four important

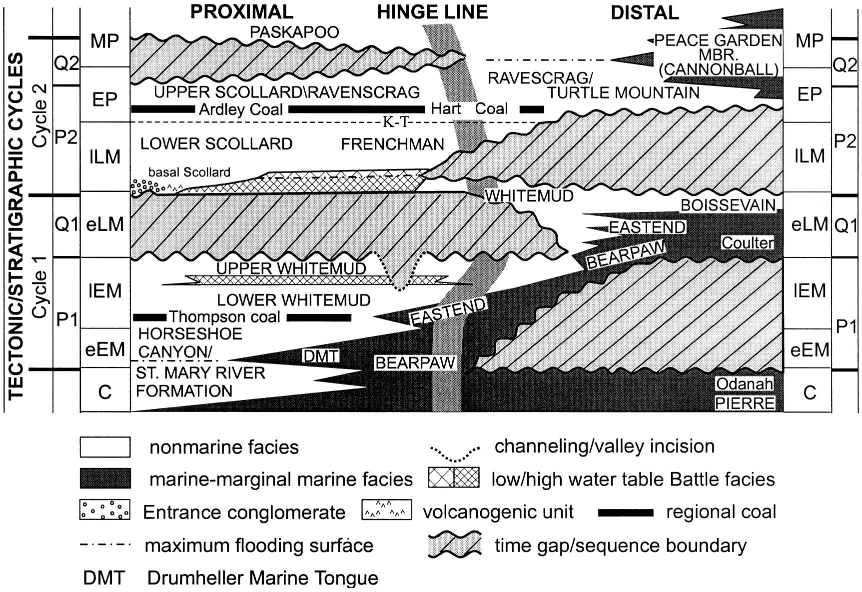


Fig. 10. Summary diagram illustrating an interpretation of the regional stratigraphic relationships, from southwestern Alberta (proximal) to southwestern Manitoba (distal). P, Q: orogenic pulse/quiescence stages. Modified after Catuneanu et al. (1995).

stratigraphic gaps have been identified in the Maastrichtian–Paleocene section of the Western Canada basin, which are confined to either a western (proximal: south-central Alberta and southwestern Saskatchewan) or an eastern (distal: south-central Saskatchewan and Manitoba) sector (Fig. 10). The proximal sector includes two major stratigraphic gaps corresponding to the *Early* Late Maastrichtian and *Late* Early to *Early* Middle Paleocene, but has a relatively continuous record of the Early Maastrichtian and the *Late* Late Maastrichtian to *Middle* Early Paleocene intervals. The distal sector is characterized by major stratigraphic gaps corresponding to the Early Maastrichtian and the *Late* Late Maastrichtian—earliest Paleocene interval, but has a good record of the *Early* Late Maastrichtian and Early Paleocene. Three out of four stratigraphic gaps, corresponding to the *Early* Late Maastrichtian (proximal), *Late* Late Maastrichtian–earliest Paleocene (distal) and *Late* Early-*Early* Middle Paleocene (proximal), developed under subaerial conditions, and qualify as "subaerial unconformities". The distal Early Maastrichtian stratigraphic gap qualifies as "paraconformity", because there is no direct evidence of subaerial exposure associated with this surface. This paraconformity has been identified chronologically as separating two lithologically similar members, the Odanah and Coulter of the marine Pierre Formation (Figs. 8 and 10).

The two sectors displaying reciprocal stratigraphic patterns are separated by a hinge zone of proximal to distal facies change, within which the four stratigraphic gaps are gradually replaced by correlative sedimentary wedges (Fig. 10). The mid point of the transitional hinge zone may be taken as a stratigraphic hinge line. As shown in Fig. 10, the hinge line for the post-Bearpaw nonmarine succession falls in the continuation of the hinge line mapped for the underlying Bearpaw Formation. In plan view, an average position for the Maastrichtian hinge line is shown in Fig. 9.

5.2. Interpretation

We associate the major stratigraphic gaps with stages of base-level fall (negative accommodation) in the areas of occurrence, whereas their correlative sedimentary wedges are interpreted to indicate stages of base-level rise (positive accommodation) in the depositional areas. A simple explanation for simultaneous base-level fall and rise within the foreland basin is provided by the model of flexural response of the foreland lithosphere to stages of orogenic loading and unloading (Fig. 4). Orogenic loading generates subsidence (base-level rise) in the proximal sector simultaneous with uplift (base-level fall) in the distal sector, which seems to be the case during the Early Maastrichtian and the *Late* Late Maastrichtian–earliest Paleocene interval (P1 and P2 pulses in Fig. 10). Orogenic unloading generates proximal uplift (base-level fall) and distal subsidence (base-level rise), which would explain the *Early* Late Maastrichtian and the *Late* Early to *Early* Middle Paleocene stratigraphies (Q1 and Q2 quiescence stages in Fig. 10). In both cases, the hinge line represents the stable alignment within the basin relative to which the flexural subsidence and uplift take place. As a result, the hinge line separates the depositional areas from the coeval non-depositional/erosional areas, which makes this inflexion point on the flexural profile the base-level for the nonmarine environments of the post-Bearpaw succession.

5.3. Boundary hierarchy within post-Bearpaw strata

Two orders of sequence boundaries have been distinguished within the post-Bearpaw nonmarine strata. Third-order boundaries are represented by the major proximal or distal stratigraphic gaps discussed above, as they provide the basic subdivision into sequences of the nonmarine portion of the Bearpaw and Cannonball second-order cycles. These boundaries are correlative with well-developed sedimentary wedges on the other side of the hinge line. The stratigraphy of the proximal third-order sequences is complicated by fourth-order boundaries associated with smaller stratigraphic gaps and therefore interpreted to correspond to stages of lower magnitude base-level fall that occurred on the overall background of the third-order base-level rise. The fourth-order boundaries tend to develop only in the proximal sector because the magnitude of flexural tectonics decreases exponentially away from the orogenic load. Cycles of orogenic loading–unloading in the orogenic belt are assumed to control both third- and fourth-order sequences, with a corresponding difference in magnitude between them.

The four third-order sequence boundaries represented in Fig. 10 are associated with significant stratigraphic gaps. Their correlative third-order maximum flooding surfaces on the other side of the hinge line are represented either by extensive coal seams in the proximal sector (i.e. the Thompson and the Ardley coals) generated during times of highest water table relative to the depositional surface, or by peaks of marine transgression in the distal sector where the much lower sediment influx allowed the preservation of the Western Interior seaway. The fourth-order sequence boundaries are associated with less (or no) time gaps, and tend to be represented by paleosols (i.e. the top of the lower Whitemud Formation, or the top of the Battle Formation, Fig. 10).

6. Reciprocal stratigraphies from ammonite data, Claggett and Bearpaw formations

To trace the hinge line position within the US portion of the Western Interior, we use the comprehensive data base published by Gill and Cobban (1973). Based on ammonite occurrence and zonation, they reconstructed the strandline position of the Western Interior seaway, south of 49° N latitude, for consecutive time steps corresponding to the uppermost Cretaceous ammonite zones. The transgressive or regressive character of each individual strandline was also indicated. Most of the studied strandlines display simultaneous transgressive and regressive trends in different parts of the basin, which led the authors to conclude that eustasy cannot be invoked as a main controlling factor on the Western Interior transgressive and regressive events. However, no explanation involving the flexural properties of the foreland basin underlying lithosphere was available at that time.

Using the concept of the hinge line, we are now able to consistently explain the observed features of the uppermost Cretaceous Western Interior strandlines. We consider in our example the Campanian–Early Maastrichtian interval, which corresponds to the second-order marine transgressions and regressions of the Claggett and Bearpaw seaways. Based on the principle that the foreland system hinge line separates sectors characterized by opposite flexural tectonics at any given time, it follows that the inflexion point of the flexural profile can be mapped as the limit at which most strandlines change their relative direction of motion. The accuracy of this interpretation could be affected by the possible strike and dip variability of

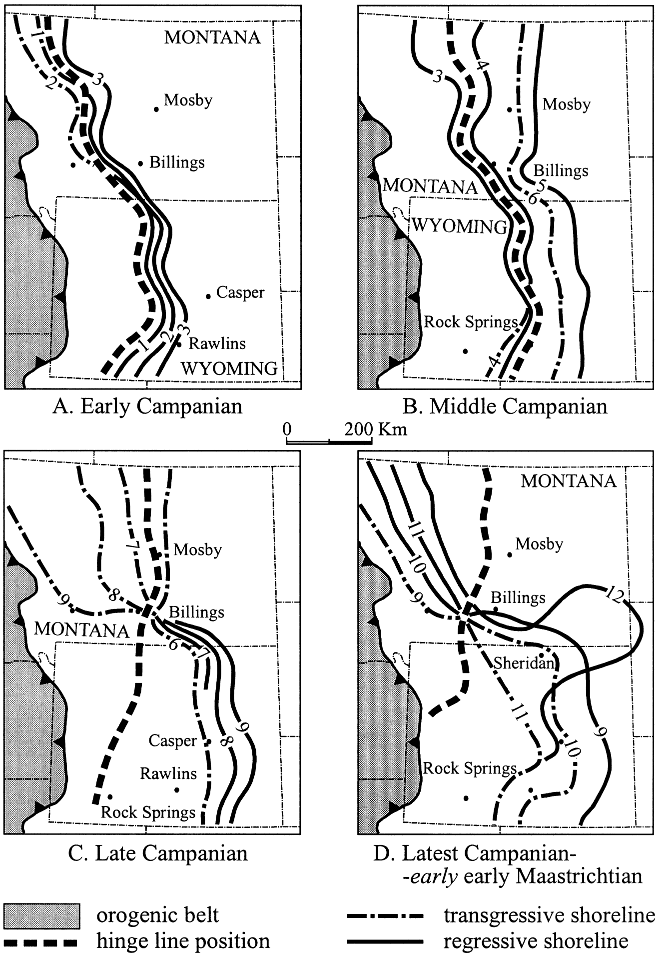


Fig. 11. Example of the use of ammonite occurrence and zonation in tracing the foreland basin hinge line (strandlines from Gill and Cobban, 1973). 1: *Baculites obtusus*; 2: *Baculites mclearni*; 3: *Baculites asperiformis*; 4: *Baculites perplexus*; 5: *Baculites gregoryensis*; 6: *Baculites scotti*; 7: *Didymoceras nebrascense-Didymoceras stevensoni*; 8: *Exiteloceras jenneyi-Didymoceras cheyennense*; 9: *Baculites compressus-Baculites cuneatus*; 10: *Baculites reesidei*; 11: *Baculites jenseni-Baculites eliasi*; 12: *Baculites baculus.*

sediment supply that may occur within the basin (Martinsen and Helland-Hansen, 1995), especially when the interpretation is based on the strandline features within a very narrow stratigraphic interval. In our case there is a remarkable consistency in the locus of change between transgressive and regressive strandlines during a time interval covering 17 ammonite zones, suggesting a dominant allocyclic control on the strandline character.

As a limit between coeval transgressive and regressive strandlines, we mapped the hinge line position at the levels of Early Campanian (*Baculites obtusus–B. asperiformis*, generally corresponding to the second-order Claggett transgression), Middle Campanian (*Baculites asperiformis–B. scotti*, second-order Claggett regression), Late Campanian (*Baculites scotti–B. compressus*, overall second-order Bearpaw transgression) and latest Campanian–*Early* Early Maastrichtian (*Baculites cuneatus–B. baculus*, second-order Bearpaw regression) (Fig. 11). We interpret continuous transgressive or continuous regressive strandlines to reflect their position entirely on one side of the hinge line, although exceptions may occur in the case of eustatically driven transgressions (which

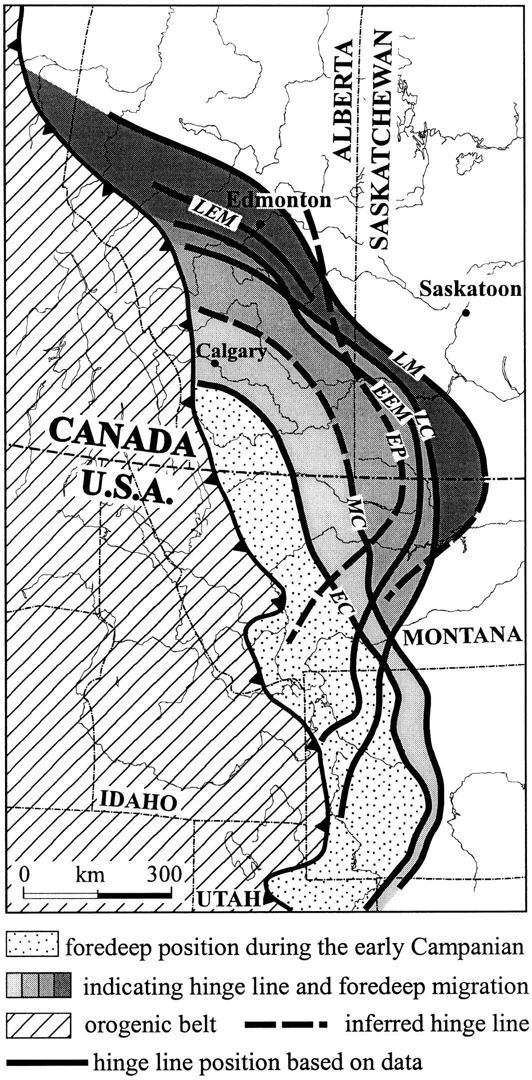


Fig. 12. Hinge line position and migration during the Campanian–Paleocene regime of dextral transpression. EC: Early Campanian; MC: Middle Campanian; LC: Late Campanian; eEM: *Early* Early Maastrichtian; lEM: *Late* Early Maastrichtian; LM: Late Maastrichtian; EP: Early Paleocene.

would produce a transgressive strandline crossing the hinge line—no example within the considered stratigraphic interval), or sediment supply-driven "normal" regressions, which we consider is the case of the *Baculites baculus* regressive strandline (Fig. 11D) as a same age normal regression terminated the Bearpaw deposition in the Western Canada Basin (Catuneanu, 1996). A particularly good correlation is obtained for the Late Campanian, as the hinge line mapped on the US side of the Western Interior based on ammonite zonation perfectly follows as a continuation of the same age hinge line mapped within the Canadian side of the Western Interior based on the facies analysis of Bearpaw sequences.

7. Hinge line position during the Campanian–Paleocene interval

Fig. 12 presents the configuration of the hinge line position for consecutive time-slices of the Campanian–Paleocene interval, compiled from the three data sets.

7.1. Early Campanian

The hinge line position is largely based on ammonite occurrences (Fig. 11A), and the foredeep is placed almost entirely on the US side of the Western Interior. The continuation of the hinge line into the Western Canada Basin was based on mapping the boundary between the correlative Milk River Formation and the Eagle shoulder (McLean, 1971). The position of the Early Campanian hinge line south of Calgary (Fig. 12) is also confirmed by the fundamental shift in the age of "Chungo" strata between the Highwood River (Virgelle Formation, Milk River Group, of Santonian age) and the Bow River (Campanian) (Sweet and Braman, 1989, with ages adjusted to conform with Leahy and Lerbekmo, 1995).

7.2. Middle Campanian

The hinge line position south of 49° N latitude is based on the interpretation of Gill and Cobban's (1973) data base (Fig. 11B). The continuation of the hinge line north of the US–Canada border is uncertain (dashed line in Fig. 12). It has been placed at the mid-distance between the traces of the Early Campanian and Late Campanian hinge lines. Further study of the Belly River (Judith River) Group is required for stratigraphic evidence. Nevertheless, the general pattern of northward hinge line migration with time supports this interpretation.

7.3. Late Campanian and Early Early Maastrichtian

Evidence for the hinge line position was obtained by facies analysis of the Bearpaw sequences on the Canadian side of the Western Interior (Catuneanu et al., 1997b), and by ammonite occurrence and zonation on the American side (Fig. 11C and D).

7.4. Late Early Maastrichtian, Late Maastrichtian and Early Paleocene

The hinge line position has been traced by palynostratigraphic mapping of stratigraphic hiatuses and correlative sedimentary wedges. For example, a *Late* Early Maastrichtian assemblage (*Scollardia trapaformis* zone, in Srivastava, 1970; Jerzykiewicz and Sweet, 1988), characteristically found in the area included here in the proximal sector, occurs as far north as the Grand Prairie region of north-central Alberta (Dawson et al., 1994) thus placing the northwestern extent of the *Late* Early Maastrichtian hinge line through the Edmonton area. A parallel but farther north position for the Late Maastrichtian hinge line is indicated by the presence of *Late* Late Maastrichtian strata in the Judy Creek area north of Edmonton (Sweet and Braman, 1992) where the *Scollardia trapaformis* zone assemblage is absent. The position of the Late Maastrichtian hinge line in the Wood Mountain area (south-central Saskatchewan) is constrained by the Wood Mountain corehole (Figs. 8 and 10), where the Late Maastrichtian stratigraphy is of proximal type, as opposed to the Early Maastrichtian and Paleocene distal stratigraphy types. The continuation of the Late Maastrichtian hinge line south of the US–Canada border is uncertain (dashed line in Fig. 12). The Early Paleocene hinge line is essentially unconstrained (dashed line in Fig. 12). However, similarities in palynological assemblages between the Wood Mountain and Turtle Mountain coreholes allows us to trace its position to the west of the Wood Mountain corehole.

8. Hinge line migration: a result of orogenic tectonics

During the Campanian–Paleocene interval the Cordilleran orogen was dominated by a tectonic regime of dextral transpression (Price, 1994). A dextral strike-slip displacement of 350 km is estimated to have occurred between 90 and 58 Ma (Price, 1994), which would imply a gradual northward migration with time of the locus of orogenic load emplacement, at an average rate of about 10 km/Ma. At the same time, the area of most active compression, tectonic loading and crustal shortening in the orogenic belt determines the locus of the associated foredeep. This may explain the northward migration of the foredeep throughout the latest Cretaceous–Early Paleocene time (Fig. 12).

The ratio between the along-strike and down-dip diameters of the foredeep at any given time can provide an indication about the shape of the emplaced thrust load: the bigger the ratio, the more elongated along-strike was the body of thrust sheets. In our case study, considering that the position of the Late Cretaceous orogenic front was about 165 km behind the present position (Price, 1994), the foredeeps identified for different time-slices of the Campanian–Paleocene interval display relatively small ratios, being close to semicircular shapes. This indicates relatively localized areas of maximum tectonic loading for any of the considered time-slices. If the step-by-step advance of the thrust had occurred isochronously along the entire orogenic belt, then the foredeep would have been extended along the entire orogenic front and the hinge line would have never met the boundary between the orogen and the foreland basin.

Besides the steady northward component of hinge line migration, which is explained by the tectonic regime of dextral transpression, contemporaneous overall cratonward shift during the Campanian and Maastrichtian, as well as orogenward shift during the Paleocene were also recorded (Fig. 12). The cratonward migration of the foredeep indicates a corresponding eastward advance of the thrust sheets and orogenic front. The westward retreat of the Paleocene hinge line suggests a cessation of the basinward progradation of orogenic load. In this case, the orogenward migration of the foreland system may either be attributed to piggyback thrusting accompanied by a retrogradation of the centre of weight within the orogenic belt, or to the retrogradation of the orogenic load through the erosion of the orogenic front during times of quiescence (Catuneanu et al, 1998a). It is also possible that visco-elastic lithospheric relaxation may have enhanced the trend of orogenward migration of the hinge line through the deepening and narrowing of the foredeep. The interpretation of the Paleocene stage as a time of overall orogenic quiescence is in agreement with the transition from dextral transpression to dextral transtension known to have occurred in the Cordilleran belt towards the end of the Paleocene (Price, 1994).

9. Discussion

The reciprocal stratigraphies recorded at the level of third- and fourth-order cyclicity indicates the foreland system sequences are controlled by orogenic cycles of loading and unloading in the fold-thrust belt, as explained by flexural models (Peper et al., 1992; Beaumont et al., 1993). When the foreland basin is occupied by an interior seaway (Bearpaw case), sediments may continue to accumulate on both sides of the hinge line until the final regression of the sea. In this case, the hinge line can be traced at the limit between coeval marine transgressive and regressive systems tracts.

Within a nonmarine environment (post-Bearpaw case), the area exposed to base-level fall develops a stratigraphic hiatus (sequence boundary). Its location alternates from the forebulge to the foredeep, respectively, during orogenic loading and unloading. In this case, the hinge line can be mapped at the limit between sequence boundaries and their time-equivalent sedimentary wedges (sequences).

The stratigraphic hinge line that we trace based on reciprocal stratigraphies is not necessarily superimposed on the flexural hinge line (i.e. inflexion point between coeval flexural uplift and subsidence) of the basin. The position of the stratigraphic hinge line relative to the flexural hinge line depends on the balance between the rates of (1) flexural tectonics (supra-lithospheric loading); (2) long-wavelength base-level rise (sub-lithospheric loading and/or eustasy); and (3) sedimentation, in the region of the flexural hinge line. When the entire foreland system is subject to base-level rise due to long-wavelength processes, the position of the stratigraphic hinge line depends on the balance between subsidence and sedimentation in the flexural hinge line area (the case of the Bearpaw marine strata). In the case of the post-Bearpaw nonmarine sequences, the stratigraphic hinge line is controlled by the balance between flexural uplift and dynamic subsidence in the region of the flexural hinge line, separating areas of falling and rising base-levels (see Catuneanu et al., 1999 for detailed discussion on this particular topic).

10. Conclusions

1. Reciprocal stratigraphies refer to out-of-phase stratigraphic sequences developed in response to opposite flexural tectonics between the foredeep and forebulge settings of a foreland system. The genesis and styles of reciprocal stratigraphies depend on the interplay between flexural tectonics, long-wavelength base-level rise and sedimentation.
2. The reciprocal stratigraphies developed between the two sectors of the foreland system are separated by a hinge zone of facies transition. The mid point of the hinge zone may be defined as stratigraphic hinge line. The position of the stratigraphic hinge line relative to the flexural hinge line depends on the balance between subsidence and sedimentation in the region of the flexural hinge line (Bearpaw case), or between flexural uplift and dynamic subsidence adjacent to the flexural hinge line (post-Bearpaw case).
3. As opposed to the flexural profile that shows a more elevated forebulge relative to the foredeep at all times (Fig. 4A), the elevation of the surface profile alternates, being higher in the forebulge during orogenic loading and higher in the foredeep during unloading, as a result of the interplay between tectonics and sedimentation (Fig. 4B).
4. The depocenter of the foreland system alternates between the depositional foredeep, during orogenic loading, and the depositional foresag, during orogenic unloading (Fig. 4).
5. The hinge line migrates through time in response to the redistribution of orogenic load. A change in the locus of the main compressional stress along the strike of the thrust-fold belt determines a parallel migration of the foredeep depocenter. The thrust sheet advance towards the basin causes a corresponding cratonward migration of the hinge line, whereas a cessation of the basinward progradation of load allows for the visco-elastic deepening and narrowing of the foredeep, i.e. orogenward hinge

line migration. Orogenward hinge line migration is also controlled by the retrogradation of the centre of mass in the orogenic belt, which may occur through piggyback thrusting without the progradation of the orogenic front, or through the erosion of the orogenic front during orogenic unloading.

6. The foredeep of the Western Interior system migrated to the north during the Campanian–Paleocene interval at a rate of about 10 km/Ma in response to the regime of dextral transpression in the Cordilleran belt. The semicircular shape of the foredeep at consecutive time steps suggests localized areas of thrusting/loading. In parallel to the strike migration of the foredeep, a dip-oriented shift is also recorded, i.e. towards the craton during the Campanian and Maastrichtian, and towards the orogen during the Paleocene (Fig. 12).

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