Maastrichtian–Paleocene foreland-basin stratigraphies, western Canada: a reciprocal sequence architecture[1]

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Abstract: Palynological and magnetostratigraphic chronostratigraphic correlations of lower Maastrichtian to Paleocene strata along an east–west Western Canada Basin transect allow for the recognition of a reciprocal sequence architecture in nonmarine strata. Reference sections include three Canadian Continental Drilling Program Cretaceous–Tertiary Boundary Project core holes and outcrops in Alberta, southern Saskatchewan, and north-central Montana. The spatial and temporal position of the third-order sequences provides evidence for the correlation of proximal sector regional disconformities and sedimentary wedges with distal sector sedimentary wedges and regional disconformities, respectively. The boundary between the two sectors is represented by a hingeline, which separates the foreland-basin "syncline" from the "peripheral bulge." The stratigraphies defined by reciprocal third-order sequences are complicated by fourth-order boundaries, developed within proximal sedimentary wedges and with no correlative distal strata. These results support tectonic control on foreland-basin sedimentation. A model for interpreting the various types of sequences in terms of foreland-basin evolution, vertical tectonics, and orogenic cycles is provided. It is argued that nonmarine sequence boundaries (times of maximum uplift in the foreland region) may be expressed as disconformities, incised valleys, top of mature paleosol levels, or base of fluvial channels, whereas nonmarine equivalents of marine maximum flooding surfaces (times of maximum basinal subsidence) may be indicated by extensive coal seams and (or) lacustrine sediments.

Résumé : Les corrélations palynologiques, magnétostratigraphiques et chronostratigraphiques des strates échelonnées du Maastrichtien inférieur au Paléocène, le long du transect est-ouest du bassin de la Cordillère occidentale au Canada, permettent la reconnaissance d'une ordonnance architecturale de séquences réciproques dans les strates continentales. Les coupes de référence incluent trois trous de forage carottés du «Canadian Continental Drilling Program Cretaceous–Tertiary Boundary Project», et des affleurements localisés en Alberta, dans le sud de la Saskatchewan et dans le centre-nord du Montana. La position spatiale et temporelle des séquences de troisième ordre plaide pour la mise en corrélation des discordances régionales et des dépôts sédimentaires en coin dans le secteur proximal avec les dépôts sédimentaires en coin et les discordances régionales dans le secteur distal, respectivement. Une ligne de jonction représente la limite entre les deux secteurs, qui sépare le bassin d'avant pays «en forme de cuvette» du «bombement périphérique». Les stratigraphies définies par les séquences de troisième ordre réciproques sont embrouillées par les limites de quatrième ordre, développées au sein des dépôts sédimentaires en coin proximaux sans la présence de strates distales corrélatives. Ces résultats appuient la thèse d'un contrôle tectonique sur le bassin sédimentaire d'avant-pays. On présente ici un modèle interprétant les divers types de séquences en termes d'évolution d'un bassin d'avant-pays, de tectonique verticale et de cycles orogéniques. Nous tentons de démontrer que les discordances, les vallées entaillées, les horizons sommitaux de paléosols matures ou la base des chenaux fluviaux peuvent aider à définir les limites des séquences continentales (périodes de soulèvement maximum dans la région de l'avant-pays), tandis que les couches de charbon étendues et (ou) les sédiments lacustres représentent plutôt les équivalents continentaux des aires d'inondation maximale (périodes de subsidence maximum du bassin).

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Introduction

The present study was stimulated by challenging questions regarding the regional correlation of disconformity-bounded stratigraphic units encountered in the three Canadian Continental Drilling Program (CCDP) Cretaceous–Tertiary (K–T) Boundary Project core holes (Elkwater, Wood Mountain, and Turtle Mountain) and associated outcrop reference sections in the south-central parts of Alberta, southern Saskatchewan, and southwestern Manitoba (Fig. 1). These lower Maastrichtian – middle Paleocene sequences are analyzed in the context of deposition occurring within the tec-

Received May 14, 1997. Accepted February 9, 1998.

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Fig. 1. Regional map of the study area showing the subcrop pattern of post-Bearpaw strata east of the foothills belt and the location of the three CCDP reference core holes. The contact between Bearpaw–pre-Bearpaw strata and the Pierre Formation in southeastern Saskatchewan is only nomenclatural. The outcrop pattern is after North and Caldwell (1970), MacDonald and Broughton (1980), and Jackson et al. (1981).

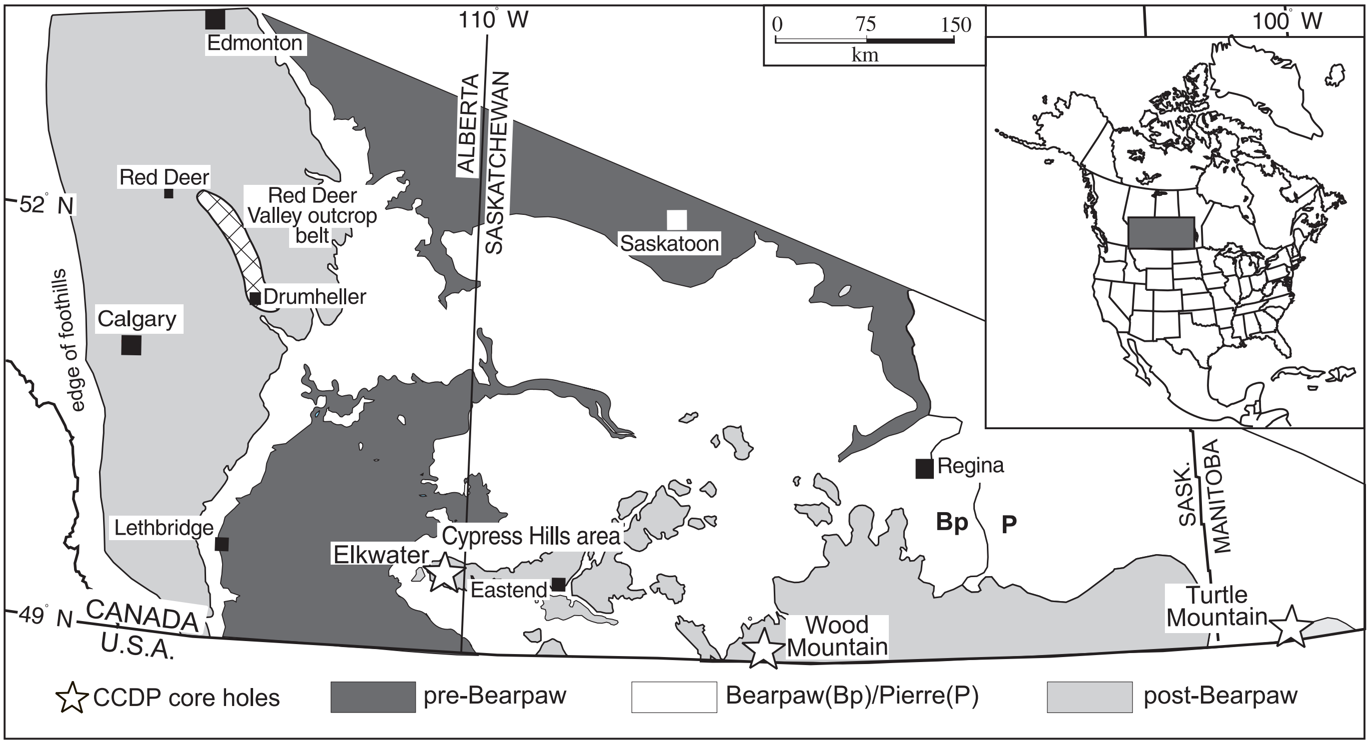
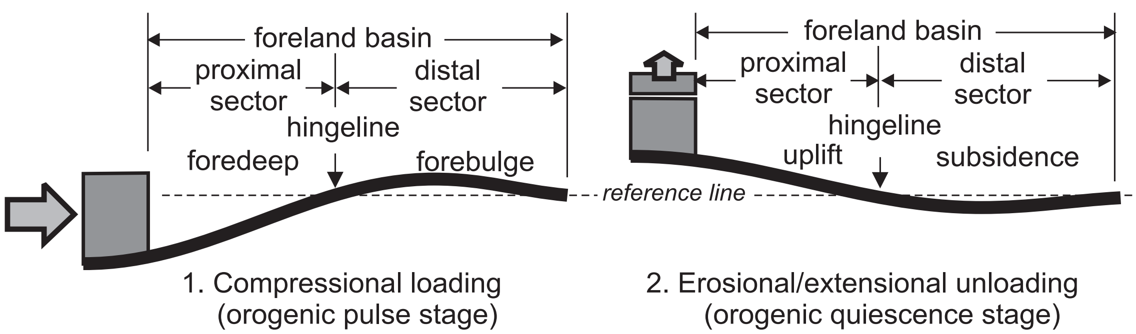


Fig. 2. Evolution of the surface profile, accounting for the combined effect of tectonics and sedimentation. Note that during each flexural state, surface processes (sedimentation, erosion) tend to level the flexural profile (not shown), allowing the mirror-image rebound of the surface profile.

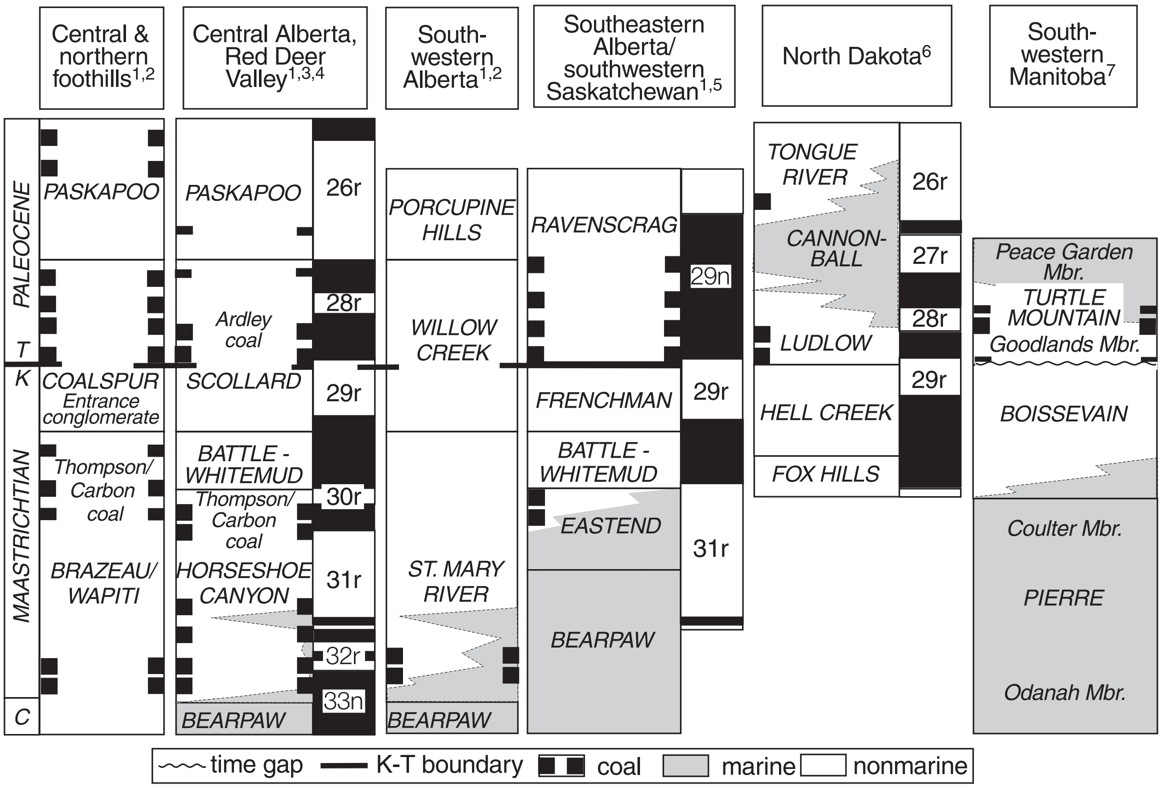


tonically controlled western Canada foreland basin (McLean and Jerzykiewicz 1978; Cant and Stockmal 1989; Stockmal et al. 1992; Beaumont et al. 1993).

In the first attempts at basin-wide correlations using information gained from the CCDP core holes, it became apparent that disconformities in the eastern portion of the study area were correlative with sedimentary wedges in the western part and vice versa, following a pattern compatible with the two-phase stratigraphic model of foreland-basin sequences presented by Heller et al. (1988). Coincidentally, Catuneanu was developing evidence, in a primarily subsurface, geophysical log based study of the marine upper Campanian to lower Maastrichtian Bearpaw Formation, for a fundamental division of the foreland basin into proximal (adjacent to the orogen) and distal (adjacent to the craton) sectors with reciprocal stratigraphies separated by a hingeline or zone of inflection between opposite (reciprocal) vertical movements within the two sectors through time (Catuneanu et al. 1997) (Fig. 2).

Previously, Beaumont et al. (1993, their Fig. 1) had theorized that foreland basins are composed of a foreland syncline (foredeep) and peripheral bulge (forebulge). Increased loading in the orogen during an orogenic pulse was predicted to result in the subsidence of the proximal part of the foreland basin and a contemporaneous uplift distally to form a forebulge. Separating orogenic pulses were times of orogenic quiescence (Beaumont et al. 1993), during which erosional off-loading in the orogen was modelled as resulting in uplift of the proximal sector and subsidence of the forebulge region or distal sector. Catuneanu et al. (1997)

Fig. 3. Selected Maastrichtian and Paleocene lithostratigraphies in the Western Canada Basin. Except for the K–T contact datum, there is no intent to imply a direct correlation between each column. Sources: 1, Dawson et al. 1994; 2, Jerzykiewicz 1997; 3, Lerbekmo and Coulter 1985; 4, Lerbekmo et al. 1992; 5, Lerbekmo 1985; 6, Hartman and Kihm; 7, Bamburak 1978. The Eastend Formation is considered marine based on a marine fauna in its lower part (Russell 1943) and dinoflagellates in its upper part. Not to vertical scale.



considered such reciprocal proximal and distal vertical movements as responsible for the generation of correlative transgressive and regressive systems tracts on adjacent sides of a “hingeline” in the marine Bearpaw Formation.

Our objective is to demonstrate the applicability and use of the principle of reciprocal stratigraphies in explaining basin-wide stratigraphic relationships of nonmarine and interfingering marine and nonmarine lower Maastrichtian to middle Paleocene sequences intersected in the three CCDP core holes. In addition, the following topics have been addressed: (1) the controlling factors (tectonics versus eustasy) on foreland-basin deposition; (2) regional patterns in the foreland-basin evolution; and (3) vertical profile models as keys in interpreting stages in the evolution of foreland-basin sequences.

Sequence stratigraphic background

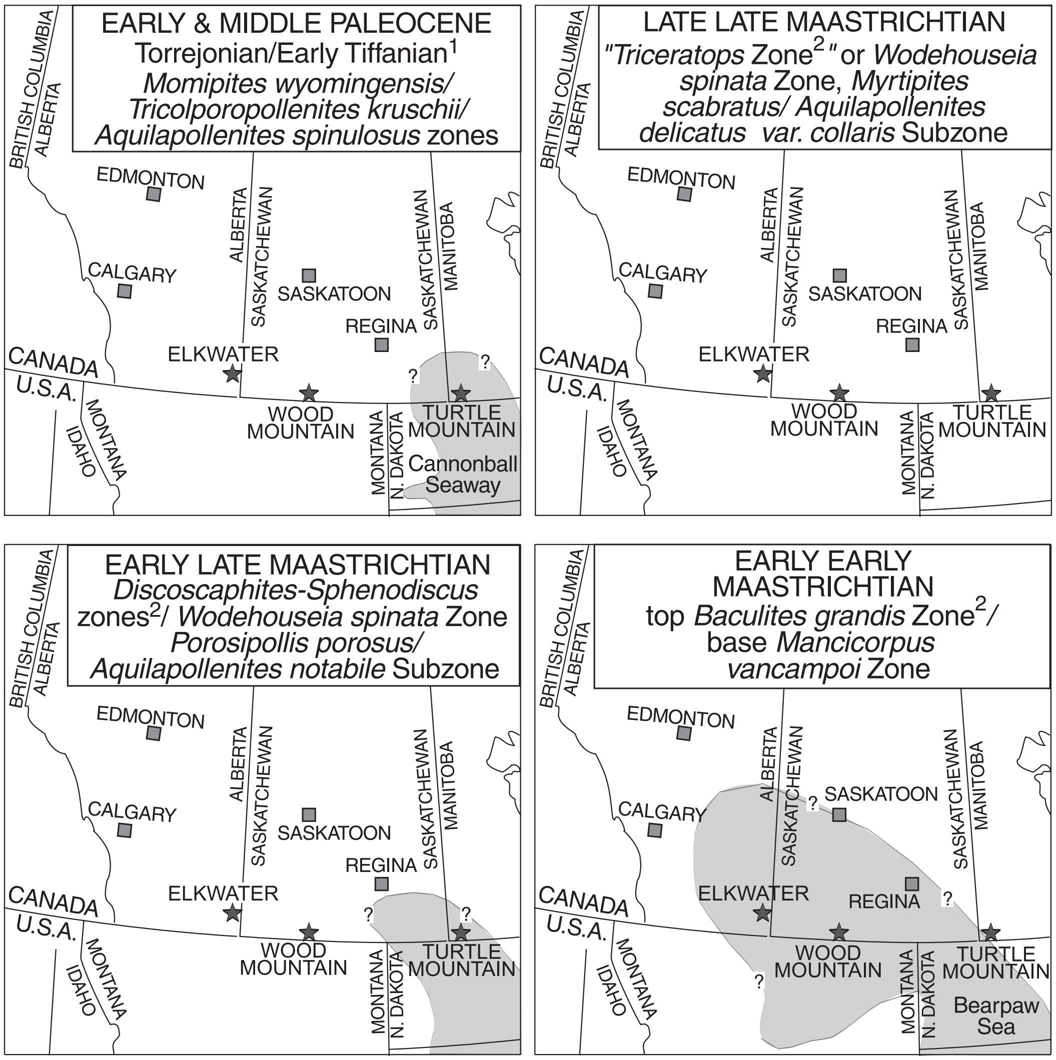
A sequence stratigraphic analysis of a nonmarine or interfingering nonmarine and marine basin fill requires exceptional mechanisms of time control, such as can be achieved by high-resolution paleontology and polarity chronologies applied to continuous rock sections, as provided by the CCDP K–T Boundary Project cores.

Classically, the nonmarine sequence boundaries are equated to subaerial unconformities (types 1 and 2, Vail et al. 1984; Posamentier et al. 1988) placed at the base of amalgamated fluvial sand sheets (Olsen et al. 1995). These may be replaced laterally by correlative mature paleosol levels (Wright and Marriott 1993) or by alignments of maximum channel amalgamation caused by decreases in accommodation space (Posamentier and Allen 1993*a*, 1993*b*). The “genetic stratigraphic sequence boundaries” (Galloway 1989, p. 139), which correlate to the marine maximum flooding surfaces, are more easily recognizable within a distance of a few tens of kilometres from the coeval shoreline in which the tidal influences can be identified (Shanley et al. 1992). Farther inland from the tidally influenced area, regionally extensive coal seams may indicate the nonmarine correlatives of the marine maximum flooding surfaces, as suggested by Hamilton and Tadros (1994). As opposed to the maximum flooding surfaces and their nonmarine correlatives, which are generally conformable, the nonmarine sequence boundaries are usually associated with time gaps (considered associated with stages of base-level fall). For the purpose of this paper we use the terms subaerial unconformity for a sequence-bounding surface that displays evidences of exposure and erosion, and disconformity for a sequence boundary associated with a time gap but with no evidence of exposure and erosional processes. In general discussions of sequence-bounding surfaces the term disconformity is used.

Geological setting

The table of formations in Fig. 3 summarizes currently accepted lithostratigraphic nomenclatures applied to Maastrichtian and Paleocene of the Western Canada Basin.

Fig. 4. Paleogeographic sketch maps showing the maximum documented extent of the midcontinental seaway during specific Maastrichtian and Paleocene intervals. Shoreline positions are based in part on Lillegraven and Ostresch (1990) and Cherven and Jacob (1985), and in part on information gained during the present study (especially critical to the extension of the early late Maastrichtian seaway into southeastern Saskatchewan). See also Fig. 6. Sources: 1, Cherven and Jacob 1985; 2, Lillegraven and Ostresch 1990.



The lithostratigraphies reflect three aspects of the depositional environment. Fundamental to regional nomenclatural differences is the timing of the eastward withdrawal of the Bearpaw seaway and the transgression and regression of the final midcontinental seaway (Cannonball) along the eastern margin of the basin (Figs. 3, 4). A second factor is regional differences in climate-sensitive lithofacies, with coal commonly present in central Alberta and southern Saskatchewan and caliche-bearing clastics commonly present in southwestern Alberta (Jerzykiewicz and Sweet 1988). The final factors affecting the lithostratigraphies are regional changes in the energy levels of the transport systems and the availability of accommodation space.

Also indicated in Fig. 3 are the stratigraphic positions of two subaerial unconformities that have been considered to have regional significance, one at the base of the Paskapoo Formation (Lerbekmo et al. 1992; Dawson et al. 1994, discussed and accepted in their text but not shown in Fig. 24.3) and one at the base of the Frenchman, Scollard, Coalspur, and Willow Creek formations (Furnival 1946; Lerbekmo 1987; Russell 1983; Dawson et al. 1994). We accept the middle Paleocene – Lower Paleocene Paskapoo–Scollard subaerial unconformity as having regional significance. However, in the following discussions, it is the Battle–Whitemud contact, rather than the Frenchman–Battle and Scollard–Battle contacts, that is documented as the more profound Maastrichtian subaerial unconformity.

Data base and results

The following discussions highlight the chronostratigraphies and subaerial unconformities–disconformities recognized in the Elkwater, Wood Mountain, and Turtle Mountain core holes (Figs. 1, 5) and associated outcrops. Detailed discussions of the lithostratigraphies, biostratigraphies, and magnetostratigraphies of the core holes are provided in Braman et al. (1999), Braman and Sweet (1999), and Lerbekmo (1999), respectively.

Fig. 5. Stratigraphic correlations between the three CCDP Cretaceous–Tertiary (K–T) Boundary Project core holes, with the lithostratigraphies from Braman et al. (1999), polarity chronologies (P) from Lerbekmo (1999), and biostratigraphies mostly after Braman and Sweet (1999). The evidence for the various time contacts is based on the following: (1) a dinoflagellate-based Campanian-aged Odanah Member (McIntyre 1999) is disconformably overlain by a *W. spinata* Zone, *P. porosus – A. notabile* Subzone (eLM), assemblage; (2) the presence of *W. spinata* and *S. trapaformis*, the general absence of other taxa typical of the eLM (Braman and Sweet 1999), and a reversed polarity chron (30r) in the Bearpaw Formation of the Wood Mountain core hole argues for a unique record of latest early Maastrichtian time, which projected eastward would disconformably overlie Campanian strata; (3) the presence of *W. spinata* in the Bearpaw Formation of the Wood Mountain core hole allies this latest early Maastrichtian assemblage with the *W. spinata* Zone, *P. porosus – A. notabile* Subzone (eLM), assemblage, suggesting a conformable upper contact; (4) the relative shortness of 30n in the Wood Mountain and Elkwater core holes compared with its length in the Turtle Mountain core hole and the apparent absence of an eLM assemblage argue for a subaerial unconformity; (5) notwithstanding the lack of palynological information from the sandstones of the Eastend, Whitemud, and Frenchman formations in the Wood Mountain core hole, a hiatus between the eLM assemblage of the Turtle Mountain core hole and the lLM assemblage of the Frenchman Formation is inferred from the absence of the C subzone.(mLM) assemblage (Fig. 6) from both the nearby Claybank and Big Muddy sections (Figs. 7, 9); (6) a major subaerial unconformity is indicated by the absence of the *W. spinata* Zone, *M. scabratus – A. delicatus* var. *collaris* (lLM) and *A. reticulatus* (eeEP) subzones, from the Turtle Mountain core hole; (7) lithologically and palynologically (Sweet et al. 1999), a conformable K–T boundary is present in the Wood Mountain core hole; (8) the approximate position of the K–T contact in the Elkwater core hole is identified by the presence of a late late Maastrichtian assemblage below the base of the sandstone (123.8 m) and the lack of species characteristic of the Maastrichtian above the sandstone, but the immediate boundary interval is not preserved; (9) the *W. spinata* Zone, *A. reticulatus* (eeEP) Subzone, occurs within the upper part of polarity chron 29r in the Wood Mountain and Elkwater core holes, consistent with data from outcrops in southwestern Saskatchewan and central Alberta (Sweet 1978; Lerbekmo 1985; Lerbekmo et al. 1987); this subzone was not seen in the Turtle Mountain core hole, which is consistent with the absence of 29r; (10) a time boundary is placed between 29n and 28r to illustrate the thinning of 29n from west to east and the relative thickening of polarity chron 28 from east to west; this may imply the presence of a subaerial unconformity within 29n in the Wood Mountain and Turtle Mountain core holes; however, there were no conclusive palynological arguments to support this conclusion (Braman and Sweet 1999); and (11) lack of palynological recovery from the buff facies of the Wood Mountain and Elkwater core holes does not allow the biostratigraphic positioning of the regional Early–middle Paleocene subaerial unconformity shown in Fig. 3 nor an unquestioned determination of its lateral extent. A middle Paleocene age at the top of the Elkwater core hole is inferred from the interpreted polarity chronology.

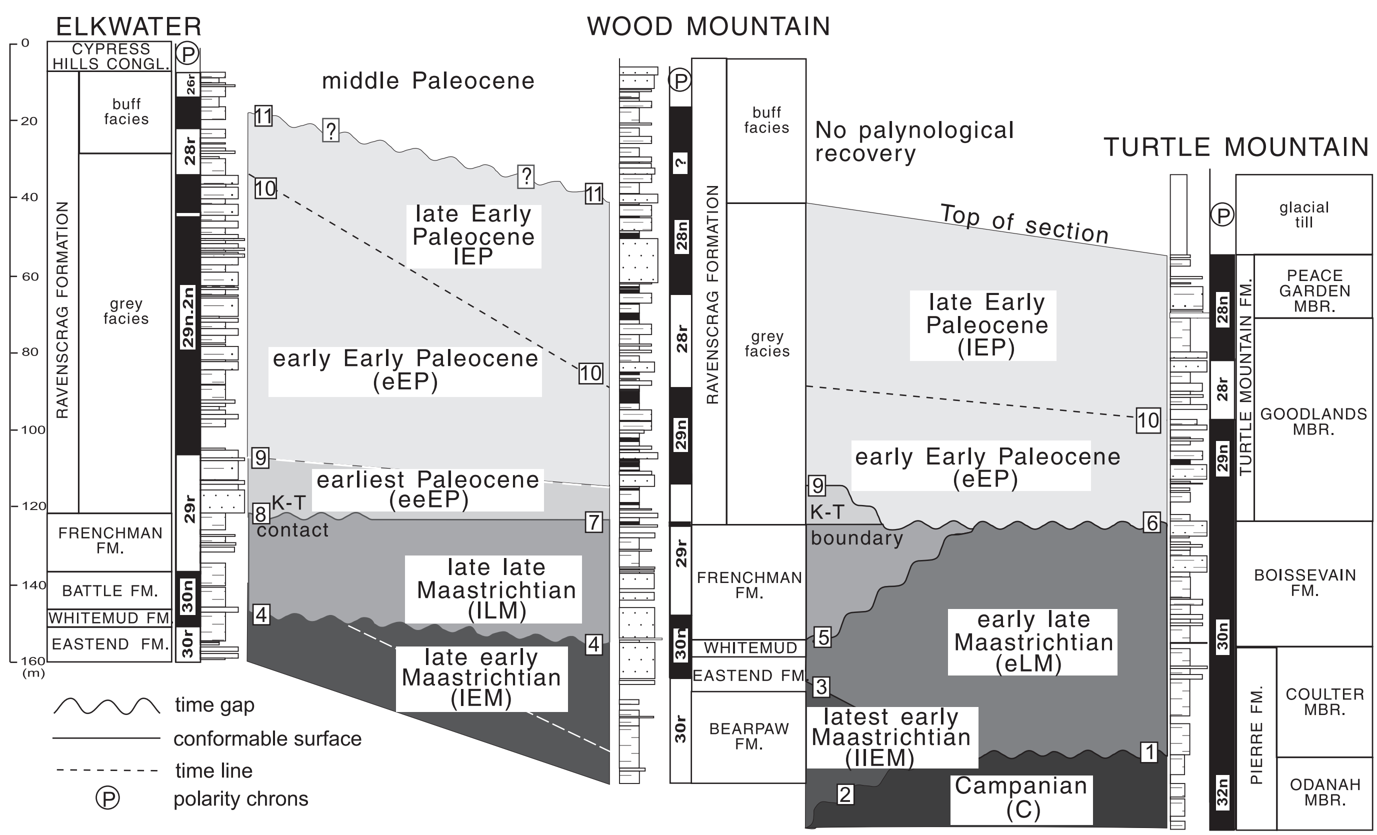
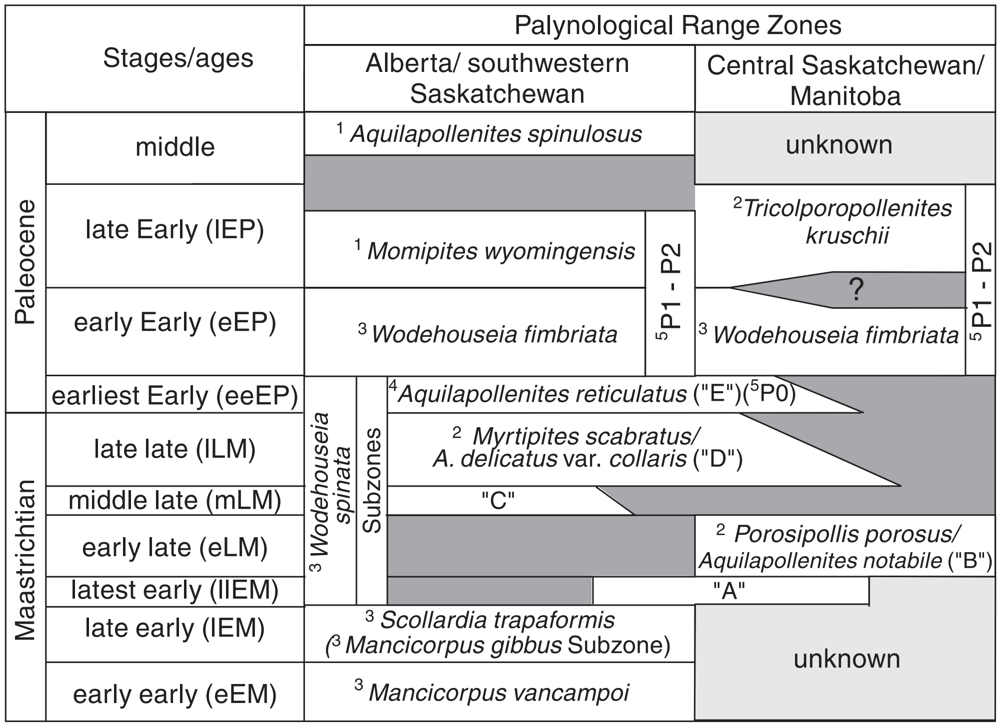


Fig. 6. Maastrichtian and Paleocene stages–ages, palynological range zones, and polarity chrons. Sources: 1, Demchuk 1990; 2, Braman and Sweet 1999; 3, Srivastava 1970; 4, Sweet 1978; 5, Lerbekmo 1999. Subzone C is based on the assemblage in the Henry's farm section (Fig. 8A), and subzone A on the assemblage in the Bearpaw Formation of the Wood Mountain core hole as discussed in the text. Not to vertical scale.



The palynological assemblages and their associated stratigraphies from the core holes are supplemented by both new and published information from outcrops in central and southern Alberta and southwestern and south-central Saskatchewan. Together these data sources have resulted in the palynozones shown in Fig. 6, which, with the exception of the informal zone C, are defined in Braman and Sweet (1999). The resolution of palynozones and associated polarity chrons allows early and late early Maastrichtian, early, middle, and late late Maastrichtian, and earliest, early, late Early, and middle Paleocene time slices to be distinguished.

Post-Bearpaw subaerial unconformities and sequences

Alberta and southwestern Saskatchewan

The lowest cored lithostratigraphic unit in the Elkwater core hole, the uppermost beds of the Eastend Formation (Fig. 5), contains a palynoflora referrable to the late early Maastrichtian *Scollardia trapaformis* Zone, *Mancicorpus gibbus* Subzone. The overlying Whitemud and Battle formations were unproductive. Elsewhere in Alberta the Thompson coal zone in the uppermost part of the Horseshoe Canyon Formation, the Whitemud Formation, and correlative strata contain a fully developed *S. trapaformis* Zone assemblage, as do the Eastend and Whitemud formations of southwestern Saskatchewan (Figs. 7C, 7D). At two localities in Alberta, Henry's farm and Scollard Canyon (Figs. 8A, 8B) and one in southwestern Saskatchewan, Dempster's Quarry (Binda et al. 1991; Nambudiri and Binda 1991), samples positioned near the contact between the Whitemud and Battle formations contain a middle or late late Maastrichtian palynological assemblage belonging to the *Wodehouseia spinata* Zone, informal subzone C and (or) *Myrtipites scabratus* – *Aquilapollenites delicatus* var. *collaris* Subzone (Fig. 7A). This substantiates an early late Maastrichtian hiatus between the Whitemud and Battle formations in Alberta and southwestern Saskatchewan.

The *W. spinata* Zone, *M. scabratus* – *A. delicatus* var. *collaris* Subzone, is also present in the Frenchman Formation of the Elkwater core hole, and this is consistent with its presence in southwestern Saskatchewan based on palynofloras reported by Sweet (1978). Additionally, it occurs in the lower Scollard and lower Coalspur formations of central Alberta (Srivastava 1970; Jerzykiewicz and Sweet 1986). The only occurrence of the older early late Maastrichtian palynological assemblage (Fig. 6) in the western portion of the foreland basin is in a volcanogenic chert and bentonitic unit of the Coalspur Formation in the Luscar–Sterco mine (Jerzykiewicz and McLean 1977; A.R. Sweet, section 77TJ6, p. 28 *in* Jerzykiewicz and McLean 1980) (Fig. 7B).

The vertical extent of polarity chron 29r and the biostratigraphy in the Elkwater core hole argue for the K–T contact to be at the base of a 5 m thick sandstone at the 123.8 m depth horizon. As the K–T boundary claystone is not present, there is at least a short K–T boundary hiatus in the Elkwater core hole. Otherwise, the K–T boundary claystone is often present in sections located in a south-southeast-directed band cutting across central Alberta and southern Saskatchewan (Sweet et al. 1999, their Fig. 3), a band more or less following the eastern limit of the Battle

Fig. 7. Locality maps showing the distribution of Whitemud–Battle formations and contiguous strata and the occurrences of either associated index species or zonal assemblages. Sections used for reference include the Elkwater (12-4-8-3 W4) and Turtle Mountain (13-17-1-23 W1) core holes and the following: BM, Big Muddy (5-27-3-24 W2); BS, Blackstone River (8-19-43-16 W5); CB, Claybank (7-28-12-24 W2); DQ, Dempster Quarry (3-6-7-21 W3); EB, Eagle Butte (10, 15-9-8-4 W4); GL, Gleichen (3,6-25-22-23 W4); GV, Grand Valley Creek (8-13-26-5 W5); HA, Hand Hills (2-24-29-18 W4); HC, Hell Creek area (northeast 1/4, sec. 22, tp. 21N, rge. 37E); HF, Henry's farm; HH, Hammer Hill (1-16-23-23 W4); HS, Horseshoe Canyon (3-27-28-21 W4); JC, Judy Creek (313A core hole, 7-27-6-23 W3); KF, Knudsen's farm (7-11-34-24 W4); LS, Luscar–Sterco mine (14-25-47-20 W5); NM, Nose Mountain (16-14-63-11 W6); OR, Oldman River (7-2-18-25 W4); RB, Ravenscrag Butte (7-27-6-23 W3); RC, Rock Creek (9-14-1-5 W3); SC, Scollard Canyon (6-20-34-21 W4); SM, Strathmore (28,30-24-22 W4); ST, Strawberry Creek; TC, Thelma Creek (5,6-18-7-2 W4); WC, Whitecourt (3-12-60-12 W5); WM, Wood Mountain Creek (2-34-5-3 W3); WMC, Wood Mountain core hole (13-21-1-2 W3); and WR, west Ravenscrag (1-27-6-24 W3). Except for the Hell Creek area, there is at present no direct palynological information on the age of the Colgate Member of the Fox Hills Formation relative to that determined for the early and late Maastrichtian Whitemud cycles, and therefore it is not distinguished on the maps. However, it is accepted as an extension of the Whitemud facies into Montana and North and South Dakota. Shoreline positions are based in part on Lillegraven and Ostresch (1990) and Cherven and Jacob (1985), and in part on information gained during the present study, which was especially critical to the extension of the early late Maastrichtian seaways into southeastern Saskatchewan.

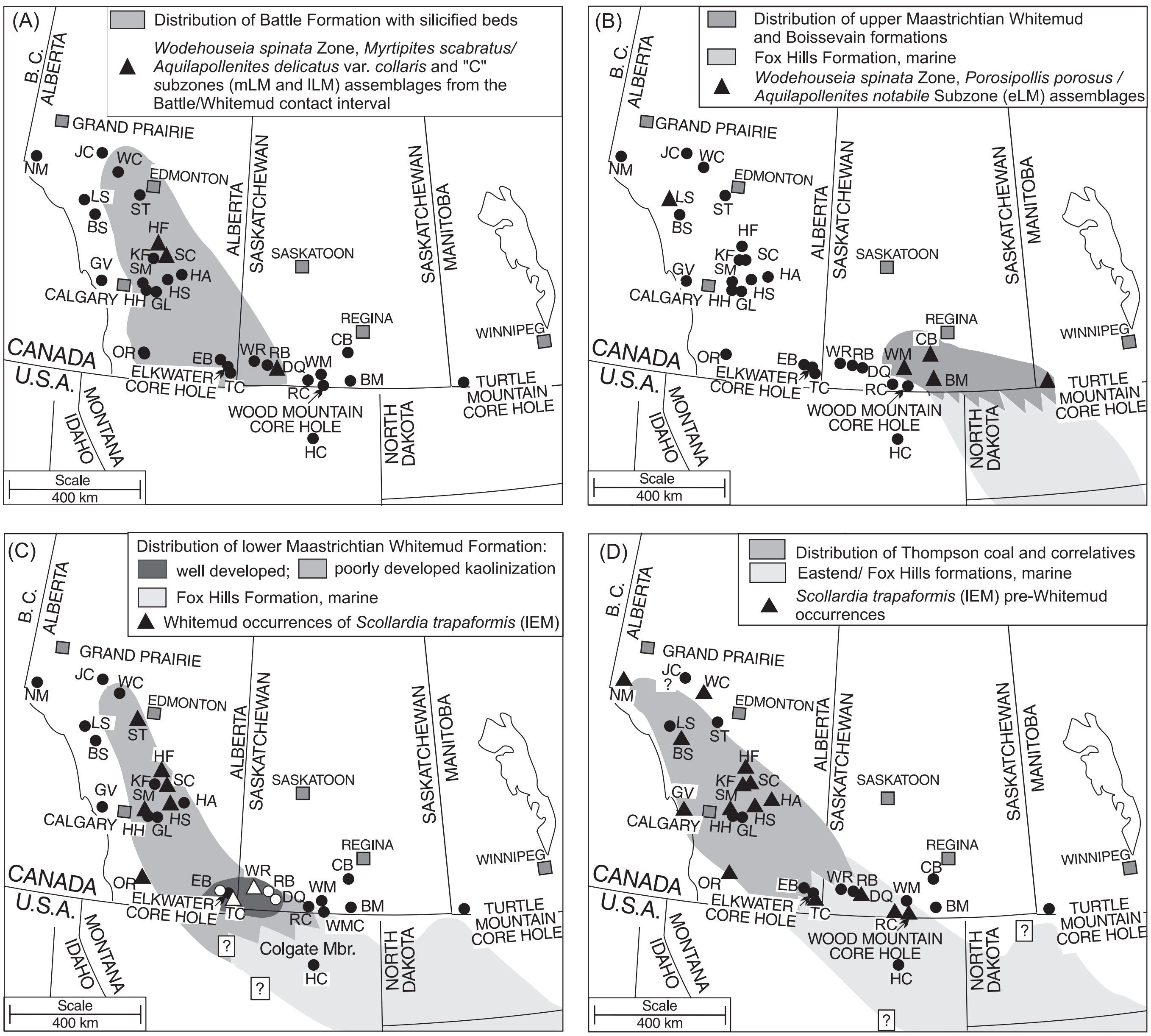
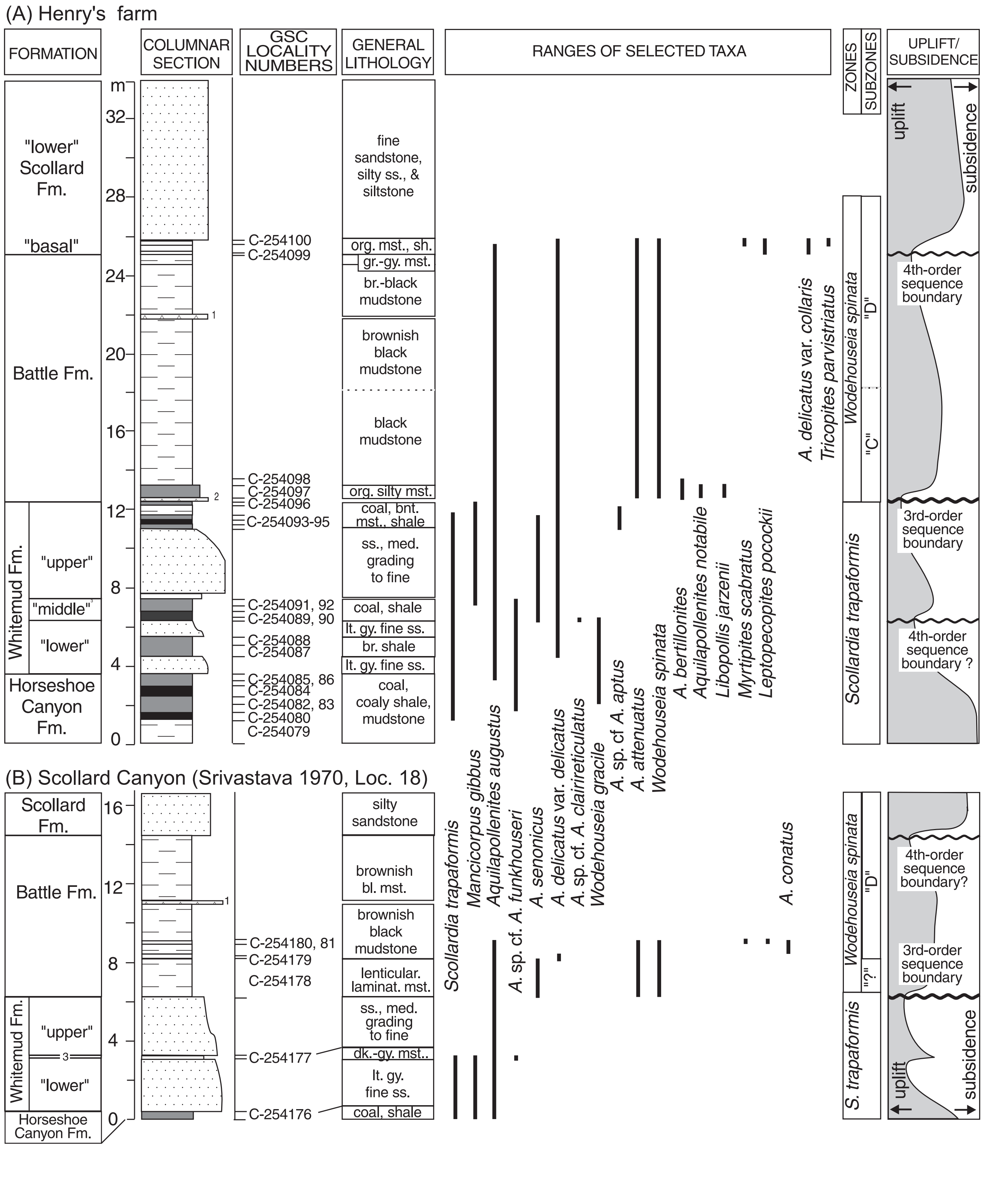


Fig. 8. The lithostratigraphy and biostratigraphy of the Henry's farm (A) and Scollard Canyon (B) sections. The column to the right is an interpretive diagram of relative uplift and subsidence. 1, Kneehills tuff; 2, unnamed silicified tuff(?); 3, "lower" Battle Formation; "C", undescribed subzone characterized by an abundance of *Aquilapollenites bertillonites*; "D", *M. scabratus – A. delicatus* var. *collaris* Subzone; GSC, Geological Survey of Canada.



Formation (Fig. 7A). Hence, its absence in the Elkwater core hole is most likely attributable to local channelling rather than a regional hiatus.

A comparatively thick Early Paleocene interval is present in the Elkwater core hole, which is illustrated as being truncated by a subaerial unconformity (Fig. 5) with the resulting hiatus possibly embracing polarity chron 27 (Lerbekmo 1999). This interpretation is consistent with the results of the Paleocene polarity chronology established for sections along the Red Deer valley in central Alberta (Lerbekmo et al. 1992) where an about 1.5 Ma hiatus, also involving polarity chron 27, was documented between the Scollard and Paskapoo formations (Fig. 2).

South-central Saskatchewan

The Wood Mountain core hole (Fig. 1) appears to be strategically located within a zone transitional between western and eastern chronostratigraphies. No major hiatuses were identified in the core hole, although, based on regional patterns, one of early late Maastrichtian age may be present. *Scollardia trapaformis* and *W. spinata* occur together in the upper beds of the Bearpaw Formation. These overlapping ranges are the bases for an interpreted latest early Maastrichtian age for the Bearpaw in the Wood Mountain core hole (Fig. 5) and an informal *W. spinata* Zone, subzone A in Fig. 6. There is no biostratigraphic information from the sandstones of the Eastend and Whitemud formations, neither were there any sediments attributable to the Battle Formation present in the core hole. This contrasts to the Claybank section to the northeast (Figs. 7, 9) where the uppermost part of the Bearpaw, Eastend, and Whitemud formations all contain mudstones that yield a prolific assemblage assignable to the early late Maastrichtian *W. spinata* Zone, *Porosipollis porosus – Aquilapollenites notabile* Subzone (Figs. 6, 7B, 9), an assemblage not generally present in the western part of the foreland basin.

The late late Maastrichtian *W. spinata* Zone, *M. scabratus – A. delicatus* var. *collaris* Subzone, occurs in the 20 m thick Frenchman Formation in the Wood Mountain core hole where its upwards truncation coincides with the K–T boundary claystone horizon (Sweet et al. 1999). Similar Frenchman thicknesses occur to the west along Rock Creek (A.R. Sweet, unpublished data) where sections also contain the K–T boundary claystone (Sweet and Braman 1992), providing evidence of continuous sedimentation across the K–T boundary. However, to the northeast and east of the Wood Mountain core hole (Claybank, Big Muddy, and associated sections; Fig. 7) the Frenchman Formation, and its contained *W. spinata* Zone, *M. scabratus – A. delicatus* var. *collaris* Subzone, thins to about 1 m. This is taken as evidence for an eastward attenuation of late late Maastrichtian aged sediments. As this relationship occurs over a relatively large area and coincides with the absence of the K–T boundary claystone, it is taken as indicating the eastward development of a latest Maastrichtian – earliest Paleocene subaerial unconformity.

The earliest Paleocene *W. spinata* Zone, *Aquilapollenites reticulatus* Subzone, and *Wodehouseia fimbriata* Zone are both well developed in the Wood Mountain core hole and along Rock Creek and in the Big Muddy section. These zones are succeeded by the *Tricolporopollenites kruschii* Zone, whose assemblage is closely comparable to that of the Early Paleocene in the Turtle Mountain core hole to the east and less so to the western *Momipites wyomingensis* Zone assemblage (Figs. 5, 6). The exact time relationship of these latter two zones is not yet fully understood. The Paleocene section in the Wood Mountain core hole is capped by beds of the buff facies of the Ravenscrag Formation that is barren of palynomorphs.

Southwestern Manitoba

As there are few, and then only discontinuous, Maastrichtian and Paleocene outcrops along the flanks of Turtle Mountain, southwestern Manitoba, this discussion focuses exclusively on the Turtle Mountain core hole (Fig. 5). The base of the core hole intersected sediments of the Odanah Member of the Pierre Formation, which is of Campanian age (McIntyre 1999). The overlying Coulter Member of the Pierre Formation, the Boissevain Formation, and basal beds of the Goodlands Member of the Turtle Mountain Formation all contain an early late Maastrichtian *W. spinata* Zone, *P. porosus – A. notabile* Subzone, palynofloral assemblage (Figs. 5, 6). This implies a hiatus embracing the early Maastrichtian exists between the Odanah and Coulter members and that the section is otherwise conformable up to the basal beds of the Goodlands Member. Within the lower part of the Goodlands Member an Early Paleocene *W. fimbriata* Zone assemblage succeeds the early late Maastrichtian assemblage, thereby defining the presence of a late late Maastrichtian and earliest Paleocene hiatus in the Turtle Mountain core hole, a hiatus compatible with the eastward attenuation of late late Maastrichtian sediments and the absence of the K–T boundary claystone in central Saskatchewan.

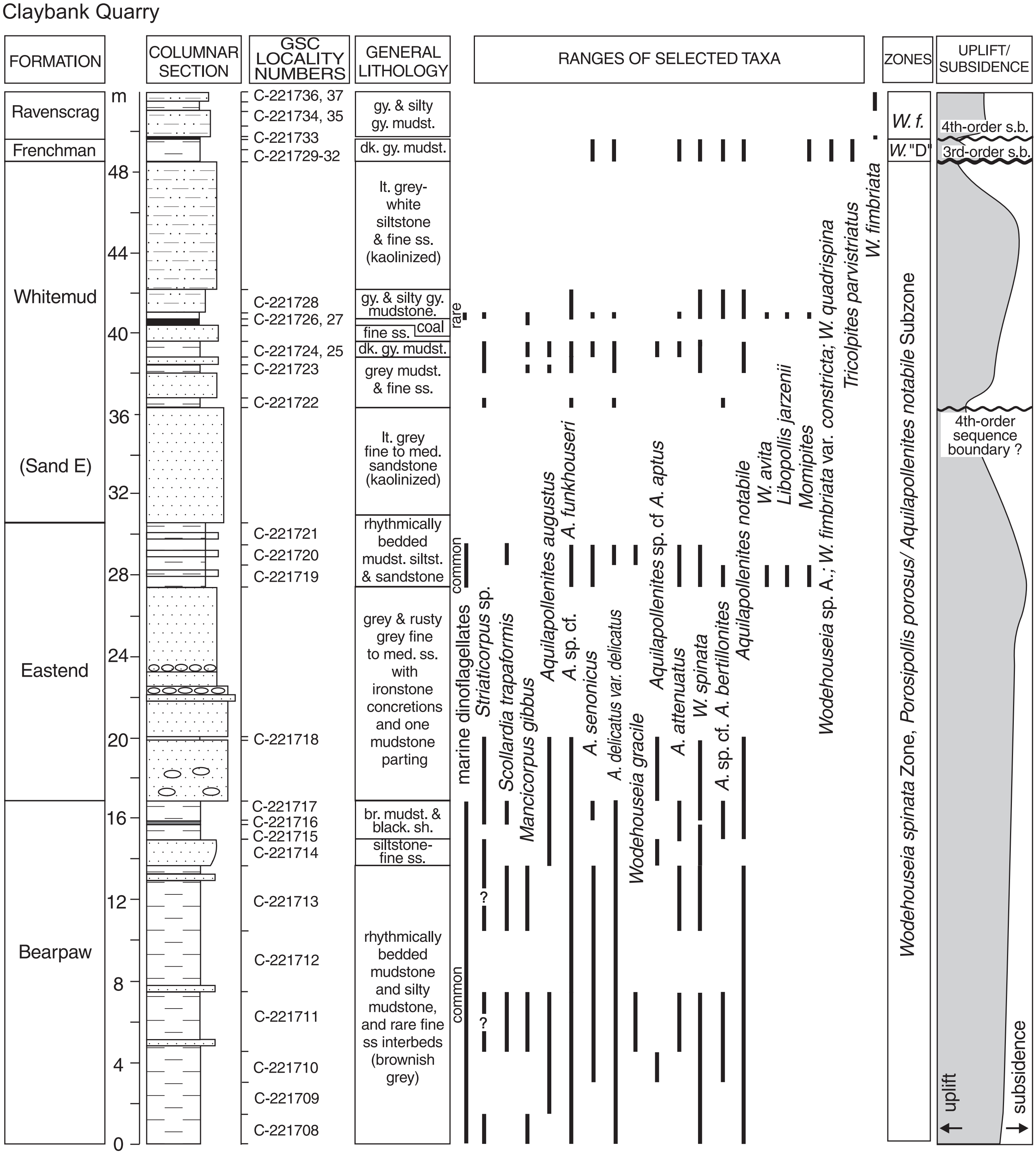
Both the upper part of the Goodlands Member and the overlying Peace Garden Member of the Turtle Mountain Formation contain the Early Paleocene *T. kruschii* Zone assemblage. As the marine Peace Garden Member is truncated by glacial sediments in the Turtle Mountain region, the question as to its upper limits in Manitoba is left unresolved. The correlative Cannonball Formation in North Dakota and South Dakota has been shown to span the polarity chrons 28n through 26r (Hartman and Kihm 1996) (Fig. 3) and hence the hiatus that occurs between the Scollard and Paskapoo formations in Alberta.

Discussion of results

Reciprocal proximal to distal stratigraphies: a key in mapping the foreland-basin hingeline

The observations made in the preceding account allow the partitioning of the Maastrichtian and Paleocene in the Western Canada Basin into two distinct sectors with reciprocal stratigraphies (Fig. 10): (1) a proximal sector with major stratigraphic gaps corresponding to the early late Maastrichtian and late Early to early middle Paleocene, but with a relatively continuous record of sediments of early Maastrichtian and late late Maastrichtian to Early Paleocene age (Figs. 5, 10); and (2) a distal sector with major stratigraphic gaps corresponding to the early Maastrichtian and late late Maastrichtian – earliest Paleocene, but with an ex-

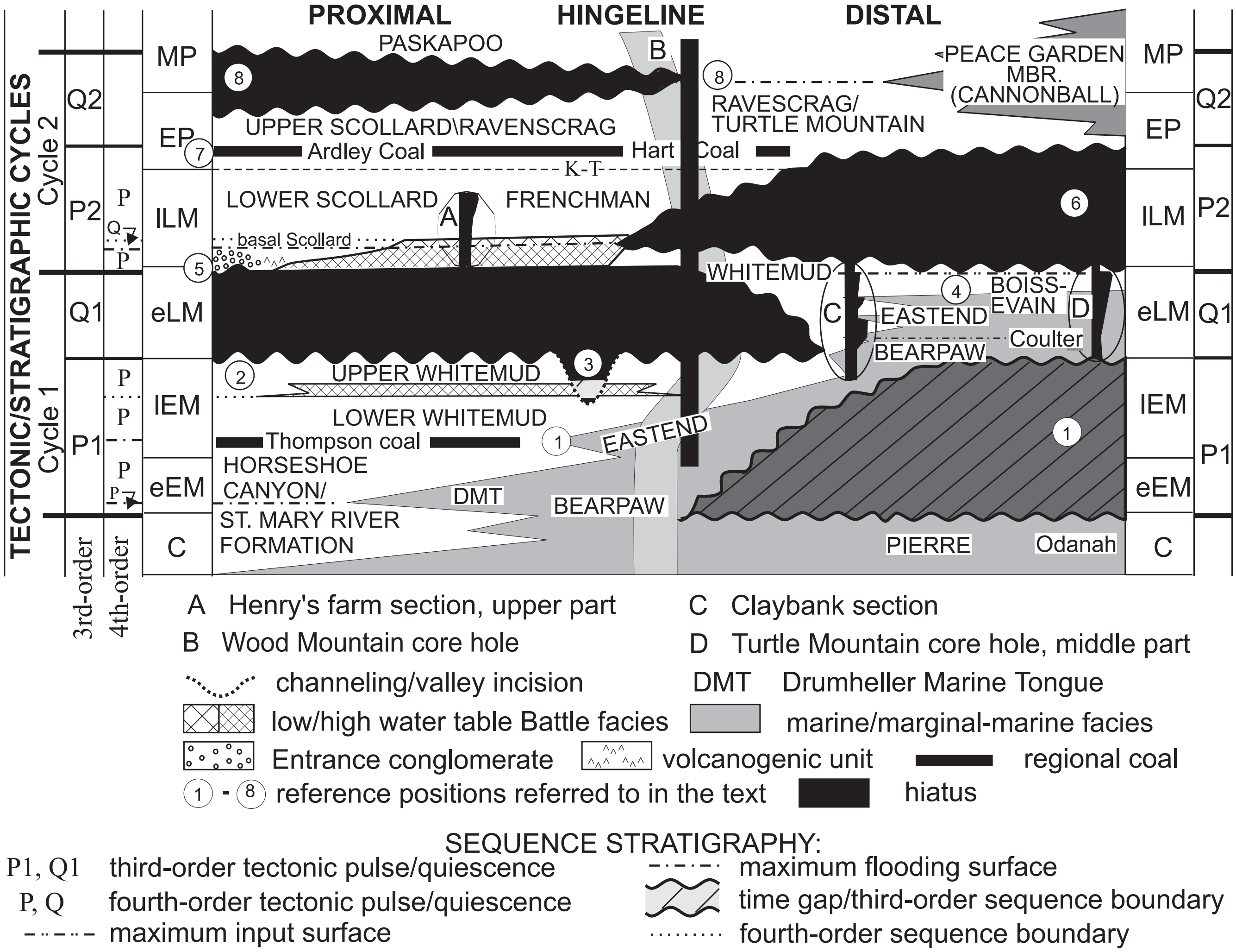
Fig. 9. The lithostratigraphy and biostratigraphy of the Claybank Quarry section. The column to the right is an interpretive diagram of relative uplift and subsidence. *W.*"D", *W. spinata* Zone, *M. scabratus – A. delicatus* var. *collaris* Subzone; *W.f.*, *W. fimbriata* Zone.



tended record of the early late Maastrichtian and late Early to middle Paleocene (Figs. 3, 5, 10).

These reciprocal stratigraphies meet at a hingeline (Fig. 10) forming the boundary between the proximal (foreland-basin "syncline") and distal ("forebulge") sectors. The correlation of proximal and distal disconformities and sedimentary wedges allows us to trace the hingeline position for the considered stratigraphic interval. The hingeline position in Fig. 10 is constrained, relative to the Wood Mountain core hole, by the location of the Campanian to early

Fig. 10. Interpretive diagram illustrating the determined regional stratigraphic relationships for the Maastrichtian and Paleocene of the Western Canada Basin.



Maastrichtian hingeline west of the Wood Mountain core hole (Catuneanu et al. 1997), the proximal-type stratigraphy in the Wood Mountains core hole for the late Maastrichtian, and the distal-type stratigraphy for the Paleocene. The eastward migration of the hingeline during the late early Maastrichtian can be explained as a result of either the strong sedimentation manifested in the current Foothills area following the deposition of the Thompson coal, or the Devonian salt dissolution (Broughton 1977, 1988) that created additional accommodation space for proximal deposition. The geographic position of the average Maastrichtian hingeline alignment is traced north-westward across the Alberta–Saskatchewan boundary (Fig. 11) in a position consistent with the results obtained from the underlying Bearpaw Formation (Catuneanu et al. 1997).

Ideally, the stratigraphic hingeline is traced as a sharp boundary in plan view that separates the reciprocally correlated proximal and distal sequences and boundaries (Fig. 11). In reality, the shift from proximal to distal stratigraphy types is not sudden, but it takes place over a distance of up to tens of kilometres. Within this transitional "hinge zone," the proximal and distal stratigraphies lose their typical signatures and gradually replace each other, which is why unconformities and sequence terminations from both sectors tend to overlap, generating a mixed and less predictable stratigraphy in central Saskatchewan (Fig. 10). The midpoint of the hinge zone at any given time defines the hingeline position for that time slice.

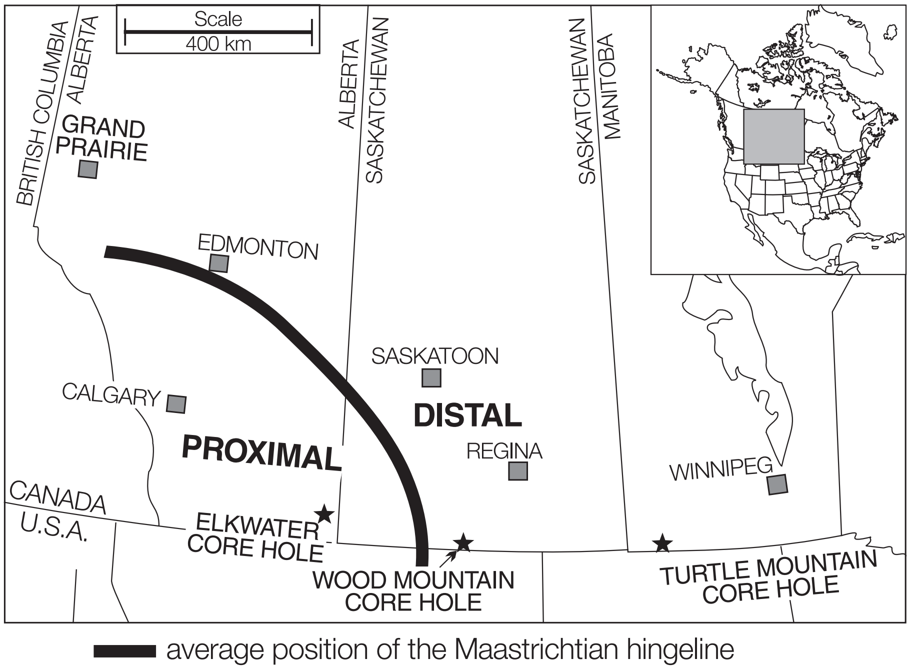
Interpretive scenario for the evolution of the study area

The tectonically controlled events illustrated in Fig. 10 are based on the following interpretive scenarios.

(1) Early Maastrichtian subsidence and sedimentation in the proximal sector culminated in a regionally developed coal zone that correlates with the development of a major disconformity in the distal sector.

(2) Deposition of Whitemud sediments occurs during the final phases of an orogenic pulse stage (P1). The early late Maastrichtian hiatus above the upper Whitemud provides the time of lowered base level required for kaolinization under favourable climatic conditions.

Fig. 11. Map showing the average position of the Maastrichtian–Paleocene hingeline.



(3) Pre-Battle channelling occurs on the eastern flank of the proximal uplifted region (southwestern Saskatchewan) and corresponds in time to the deposition of lower upper Maastrichtian Boissevain, Coulter, and correlative sediments in the distal sector.

(4) An early late Maastrichtian regression in the distal sector culminates in the nonmarine Boissevain that displays an overall coarsening-upward profile, suggesting an increase in transport energy and sediment influx to the distal area towards the upper part of the sequence. Following the deposition of the distal Boissevain Formation, initiation of backfill in the pre-Battle (formally considered pre-Frenchman) channels takes place in southwestern Saskatchewan (proximal sector).

(5) The initiation of latest early – late late Maastrichtian orogenic loading (P2) provides proximal accommodation space for the deposition of the Entrance conglomerate, the volcanogenic strata of the Luscar–Sterco mine, and lacustrine (Binda 1992) Battle Formation sediments farther to the east. The early late Maastrichtian age of the volcanogenic strata in the west, the middle late Maastrichtian age at the base of the Battle Formation at Henry's farm, and its late late Maastrichtian age in Scollard Canyon and in Dempster's Quarry in southwestern Saskatchewan indicate the base of this proximal sedimentary wedge is probably slightly diachronous, younging eastward, suggesting gradual progradation from the source area.

(6) A distal sector late late Maastrichtian – earliest Paleocene hiatus develops corresponding to the time the Battle and Scollard–Frenchman sequence was deposited. Within this overall third-order sequence were times of lower amplitude orogenic pulse–quiescence cycles in the proximal sector, one being reflected in the greenish-grey mudstones at the top of the Battle Formation and the subaerial unconformity recorded between the Frenchman and Battle formations in the Eastend area (Kupsch 1957).

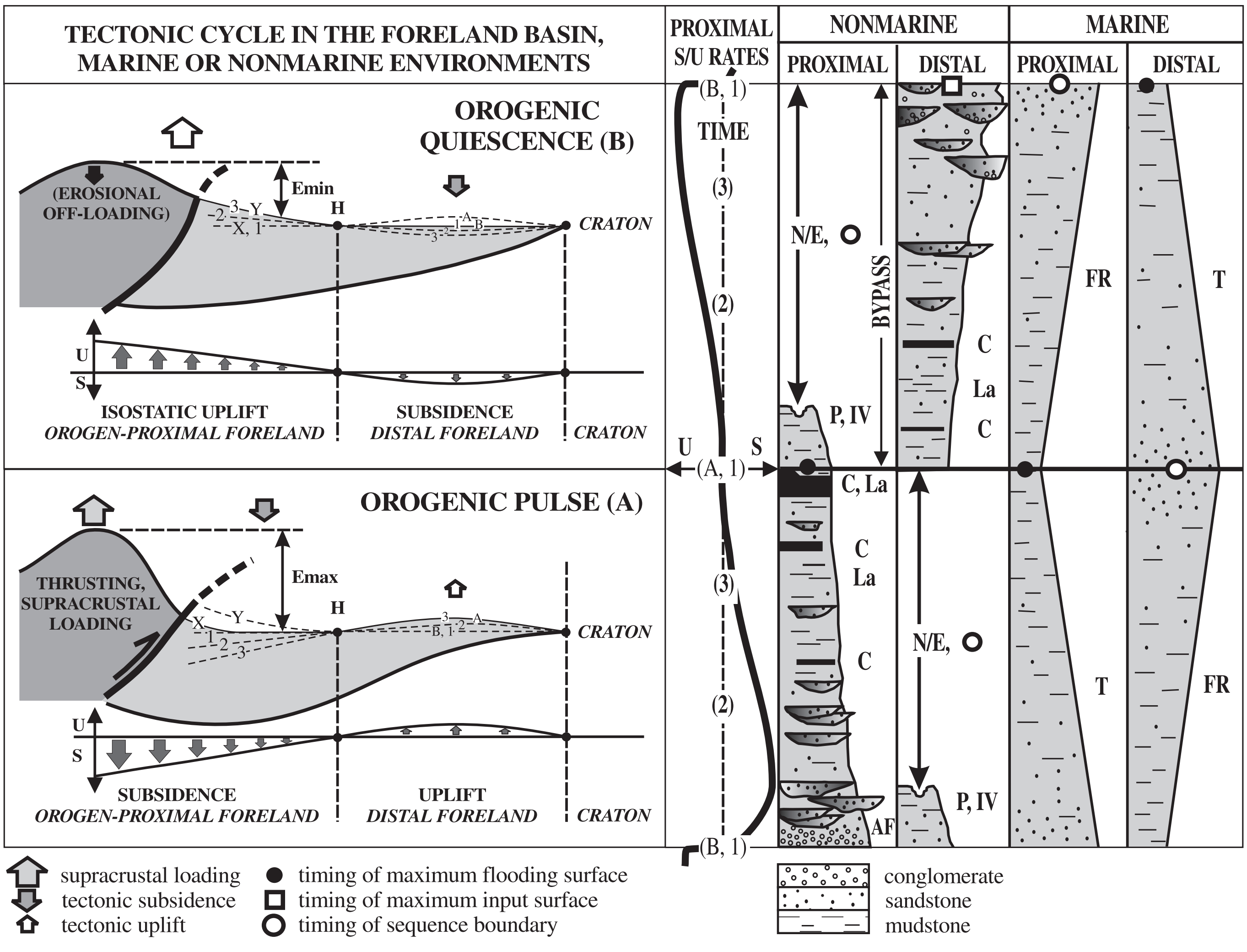
(7) The late late Maastrichtian – Early Paleocene aged proximal sequence culminates with deposition of the Ardley coal zone.

(8) Latest Early – early middle Paleocene uplift follows in the proximal sector (Q2), resulting in a latest Early – middle Paleocene subaerial unconformity that correlates, based on magnetostratigraphy (Fig. 3), with the maximum transgression of the Cannonball seaway in the distal sector.

Hierarchy of sequences and bounding surfaces

The studied Maastrichtian–Paleocene succession, partially overlapping with the Bearpaw and Cannonball second-order cyclothems (Catuneanu et al. 1997), is primarily subdivided into sequences by the major proximal and distal disconformities shown in Fig. 10. These disconformities are therefore taken as third-order boundaries, which outline the sequence architecture and stacking patterns at the level of third-order cyclicity. Because of the reciprocal nature of the basin stratigraphy, none of the third-order boundaries are developed basin-wide, being restricted to either the proximal or distal sectors of the foreland basin. This defines a pattern of reciprocal proximal and distal third-order sequences over the basin hingeline. According to this sequence stratigraphic framework, a third-order nonmarine boundary developed within the tectonic setting of a retroarc foreland basin may be described as a regional proximal or distal disconformity that has a correlative sedimentary wedge on the other side of the basin hingeline. Episodes of subsidence and increase in accommodation space adjacent to the Cordilleran orogenic belt, during thrusting, correlate to episodes of uplift and de-

Fig. 12. Flexural model explaining the relationship between orogenic cycles, base-level changes, and the deposition of third-order sequences within the foreland basin. The magnitude of uplift and subsidence in the distal foreland (bulge–sag) is exaggerated for clarity. T, transgression; FR, forced (uplift-driven) regression; NR, normal (sediment supply driven) regression; Emin and Emax, minimum and maximum difference in elevation between sediment source area and the proximal surface profile; X, surface (topographic) profile at the end of pulse stage and beginning of quiescence stage; Y, surface profile at the end of quiescence stage and beginning of pulse stage; 1, horizontal (reference) plane at the beginning of pulse or quiescence stages; 1, 2, evolution in time of the initial horizontal plane (1), suggesting proximal and distal base-level changes during pulse and quiescence stages; AF, alluvial fan, where a slope break occurs; La, lacustrine system; C, coal; P, paleosol; IV, incised valleys; N/E, nondeposition–erosion; U/S, uplift (base-level fall) – subsidence (base-level rise); H, hingeline.



crease in accommodation space over the forebulge, and the reverse during orogenic unloading.

The stratigraphic framework defined by the reciprocal stacking pattern of the third-order sequences is complicated by fourth-order boundaries developed within the proximal sedimentary wedges, associated with shorter time gaps and generated during stages of lower magnitude base-level fall such as occur in the Battle – Scollard–Frenchman sequence. Another example of a fourth-order boundary is the top of the "lower" Whitemud Formation, below the mid-Whitemud, Battle-like facies (Fig. 10). The fourth-order boundaries, within third-order proximal sequences, are interpreted as occurring as a result of lower magnitude orogenic pulse–quiescence cycles (i.e., fourth-order) but may also reflect cyclic changes in climate, fluvial discharge, or sediment supply.

The tectonic interpretation in Fig. 10 indicates the timing of sequence boundaries at the temporal limit between successive stages of tectonic uplift and subsidence. The third-order tectonic pulses responsible for the observed reciprocal sequence architecture determined the transition from uplift to subsequent subsidence stages in the proximal sector of the basin. Accordingly, when the proximal sequence boundaries are represented by subaerial unconformities (i.e., early late Maastrichtian and late Early – early middle Paleocene), the ensuing tectonic pulse terminates their generation by producing renewed subsidence and accommodation space and increased sediment input.

Tectono-depositional model for the foreland-basin evolution

Tectonic stages and phases

This study indicates that the reciprocal third-order foreland-basin stratigraphy is controlled by orogenic cycles of loading (thrusting–orogenic pulse) and unloading (orogenic quiescence), both stages inducing active vertical (flexural) tectonics within the foreland basin, but developed in opposite directions from proximal to distal and from loading to unloading (Peper et al. 1992; Beaumont et al. 1993, their Fig. 1). Field examples allowed us to refine the existent theoretical models and to relate the observed facies changes to stages and phases of tectonic evolution recorded within the foreland basin (Fig. 12).

Orogenic pulse, nonmarine environment

The orogenic pulse stage determines the amount of proximal sector subsidence and distal uplift. The configuration of the proximal sector starts as a relatively elevated, inclined surface sloping toward the hingeline (which corresponds to the topographic profile reached at the end of the previous quiescence stage; Y surface in Fig. 12) and ends as a more or less horizontal surface (X in Fig. 12) as sedimentation tends to fill the accommodation space created during the tectonic subsidence. The surface marked as 1 in Fig. 12 indicates a reference horizontal plane at the beginning of pulse stage (below the Y initial topographic profile), which reaches the flexural states 2 and 3 during the pulse stage. The evolution of the initial horizontal plane 1 into the gradually steeper surfaces 2 and 3 measures the amount of base-level rise as the result of flexural subsidence. In the distal sector, the depositional surface starts as a depressed or nearly horizontal surface (Y) and ends as an uplifted surface (X). On both sides of the hingeline the depositional processes may manifest either below or above sea level, as a function of the relative position between the sea level and the surface profile.

Within the proximal sector, the subsidence rates increase from the hingeline toward the orogenic load (Fig. 12). Nevertheless, when a seaway is present in the foredeep area, the marine facies develops on the distal side of the proximal sector, indicating that sedimentation rates usually exceed the subsidence rates in the most proximal areas of the basin. Farther cratonward from the shoreline, within the marine environment, subsidence may exceed or may be exceeded by the sedimentation rates, the balance between the two generating transgressions or normal regressions, respectively (Catuneanu et al. 1997). Consideration of the lower part of the Battle Formation provides a parallel nonmarine example. The distribution of this primarily “lacustrine” unit is restricted to the distal part of the proximal sector (Fig. 7A), away from the main sediment source, the Battle Formation being the equivalent of a marine “transgressive event.”

The orogenic pulse stage generates a proximal overall fining-upward rhythm, and may be divided into two phases dominated by distinct depositional processes.

Phase of proximal foreland topographic slope reduction

This phase starts with the orogenic pulse stage and lasts until the topographic profile reaches a horizontal position. The attenuation in the dip angle of the initial surface Y (Fig. 12) indicates that during this phase, subsidence outpaces the sedimentation rate. As base level rises with the flexural subsidence, sedimentation is active throughout the foredeep area, starting with coarse-grained alluvial-fan facies in the most proximal parts of the basin, adjacent to the orogenic front where a strong slope break occurs between the source and the depositional areas (e.g., the Entrance conglomerate of the Alberta foothills, Fig. 12). The coarse-grained facies is laterally replaced by finer grained deposits with increasing distance from the source area (e.g., the dark grey mudstone of the lower part of the “upper” Battle Formation, considered to be more or less the lateral correlative of the Entrance conglomerate; Fig. 10).

During the phase of proximal foreland topographic slope reduction, the distal depositional surface starts as a horizontal or slightly depressed surface that may still receive sediment from the more elevated proximal area (e.g., basal beds of Turtle Mountain Formation, Goodlands Member). As proximal sector subsidence progresses, sediment supply to the distal sector is cut off. This corresponds with the loss of distal accommodation space through flexural uplift, resulting in little or no sediment accumulating.

Phase of proximal balance between subsidence and sedimentation (quasi-horizontal topographic profile)

As the result of tectonic subsidence, sedimentation continues in the proximal sector but is discontinued in the elevated distal sector because it is deprived of sediment input and accommodation space. This is a phase of distal pedogenic processes and valley incision, coeval with proximal fining-upward deposition following the peak of highest subsidence rates. The fining-upward trend is explained by gradual lowering of the energy regime of the fluvial systems with the decrease in the rate of accommodation space generation proportional to the decrease in subsidence rates. Sedimentation tends to maintain a horizontal topographic profile, by matching the subsidence rates. With the decrease in the subsidence rates, the sedimentation rates decrease as well, explaining why the orogenic pulse stage usually ends with the lowest energy depositional systems, which may generate extensive coal seams or lacustrine strata. In this case, the coals or lacustrine strata, both indicating high levels of the water table relative to the depositional surface, may be correlative to marine maximum flooding surfaces, as suggested by Hamilton and Tadros (1994). Any contemporaneous distal mature soil has the significance of a sequence boundary.

Orogenic quiescence, nonmarine environment

The duration of the orogenic quiescence stage determines the amount of distal sector subsidence (base-level rise and sediment accumulation) and proximal uplift (base-level fall, insignificant sediment accumulation). The initial proximal surface (Fig. 12, X surface, coinciding with the reference horizontal plane 1) undergoes progressive uplift, with higher rates adjacent to the orogen. As a result, a gradual steepening of the topographical profile occurs together with the fall in base level and destruction of accommodation space.

The orogenic quiescence stage is modelled as generating an overall coarsening-upward distal rhythm as the gradual steepening of the proximal slope increases the energy regime and transport competency of the incoming fluvial systems. The orogenic belt acts as main sediment source area for the distal deposition, whereas the proximal sector acts as a ramp of sediment bypass. Within the proximal sector, as the transport energy of the fluvial systems increases with the sloping of the topographic profile, fluvial style changes from high to low sinuosity, and the process of valley incision gradually progresses. With increasing channel incision, the frequency of overbank flooding, the depositional energy in the off-channel regions, and the coarseness of sediments deposited in the floodplain environment are expected to decrease. Therefore, the increase in slope angle gradually increases the degree of floodplain sediment starvation, reduces the available accommodation space and preservation potential, and generates a thin, overall fining-upward rhythm. The dominant processes of the foredeep area are therefore valley incision coeval with pedogenic processes in the former floodplain areas deprived of sediment supply and with a lowered water table (e.g., as occurred during the proximal sector early late Maastrichtian hiatus above the Whitemud Formation), accompanied by minor floodplain deposition during exceptional flooding events, especially in the early stages of orogenic quiescence when the proximal profile is still dominated by a low-relief, gently sloping surface. The proximal mature paleosols and the correlative incised valley unconformities assume the significance of sequence boundaries.

During both pulse and quiescence stages, the nonmarine base level is placed at the hingeline elevation, as sedimentation (in subsiding areas) and erosion (in uplifted areas) tend to keep up with and balance the effect of flexural tectonics, in an attempt to level the topographic profile to this position (Fig. 12). Within the subsiding foredeep (Fig. 12A), the topographic profile reaches the hingeline elevation over most of the proximal sector, except for the very adjacent part near the orogenic front where the surface profile will always dip towards the basin; this is because a transition is needed between the orogenic slope and the basin profile.

Marine environment

If the foreland basin is transgressed by a seaway, establishing a marine environment during both pulse and quiescence stages, transgressive and regressive sequences are expected to correlate reciprocally across the hingeline (Fig. 12), providing that the subsidence–uplift rates outpace the sedimentation rates. When sedimentation outpaces tectonics, a "normal" (sediment supply driven) regression develops across the entire foreland basin, which may explain, as an alternative to eustatic fall, the eastward retreat of the Western Interior seaways. In any case, sedimentation in the marine environment takes place during both stages of rise and fall in the base level, as long as accommodation space is available below sea level, which is in contrast with the situation described for the nonmarine environment.

Stratigraphic consequences

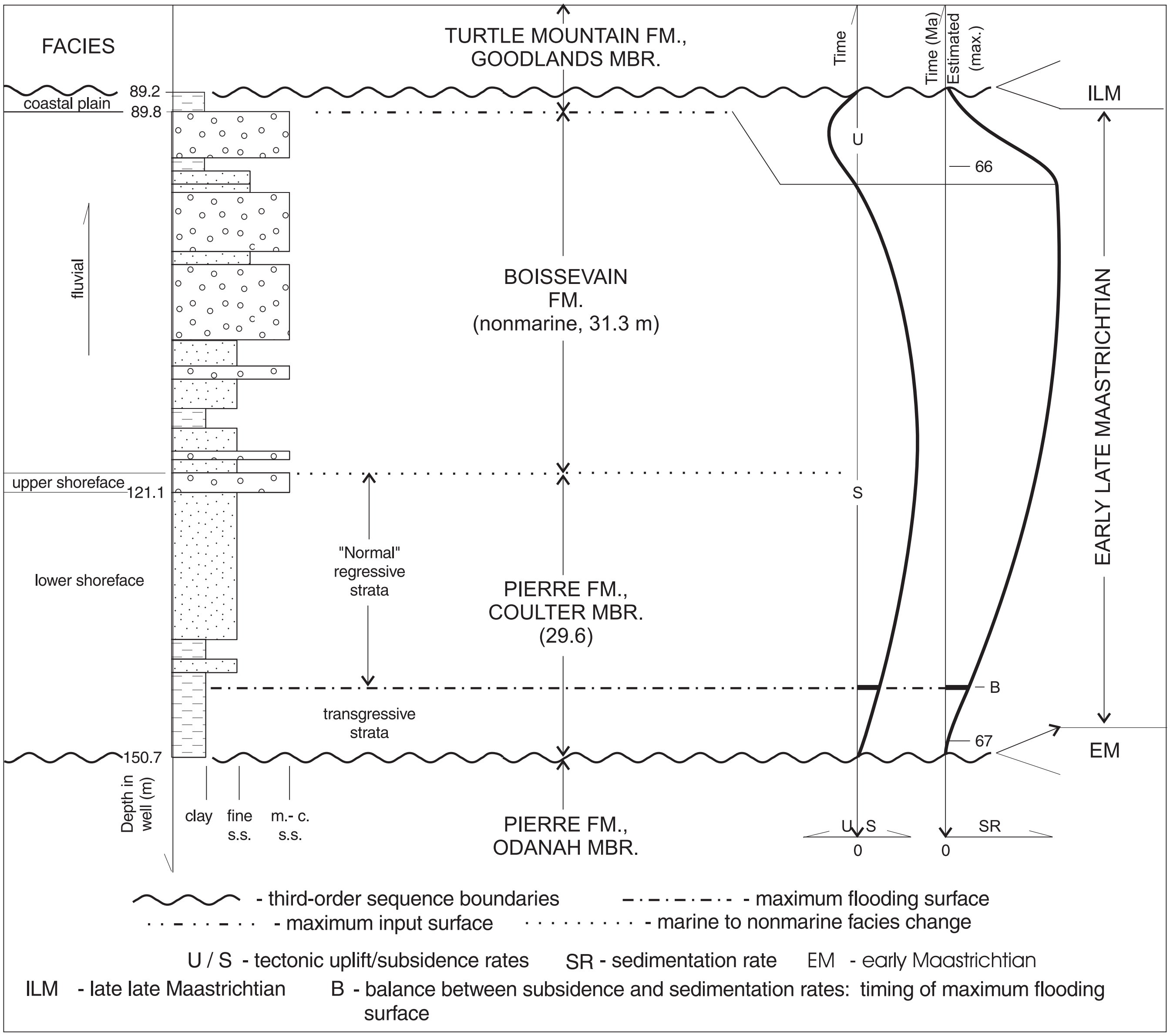
Unlike the marine stratigraphic record, which may be continuous over extended periods of time encompassing successive stages of pulse and quiescence, the preserved nonmarine stratigraphic record in both foredeep and forebulge areas is discontinuous, being mostly generated during stages of base-level rise (Fig. 12). As the rise in the base level occurs in the foredeep area only during the orogenic pulse (loading) stage, and in the forebulge area only during the orogenic quiescence (unloading) stage, it follows that no complete record of nonmarine stratigraphy can be found in any given point within the basin. Assuming that no record was erosionally removed during the subsequent stages of base-level rise, the proximal and distal stratigraphies should perfectly compensate each other, with very little overlap. According to our model (Fig. 12), the significant increase in the thickness of the foredeep succession toward the orogenic load does not immediately imply a more complete stratigraphic record, the thickness mainly depending on the rates of base-level rise and the sedimentation. Considering that the quiescence stages tend to be longer than the pulse stages (Catuneanu et al. 1997), it appears that more geological time should be represented in the stratigraphy of the distal sector than in the foredeep, although the preservation potential may be lower distally.

The concept of "maximum input surface"

Maximum flooding surfaces and sequence boundaries are easily recognizable within the foreland basin, as they have distinctive field expression in both marine (e.g., Catuneanu et al. 1997) and nonmarine (Fig. 12) strata. The interpretation of these boundaries in relationship to the orogenic cycles is relatively simple, the bounding surfaces being usually generated at the end of either pulse or quiescence stages (Fig. 12).

An interesting problem is raised by the study of distal-type stratigraphic sequences. The uniqueness of the situation in the distal setting derives from the progressive increase in sediment transport energies during the orogenic quiescence as the proximal slope steepens (Fig. 12B) and their relatively abrupt reduction with the onset of an orogenic pulse. An example is provided by the early late Maastrichtian 61.5 m thick third-order sequence encountered in the Turtle Mountain core hole (Fig. 13). This sequence consists of the marine Coulter Member (29.6 m) followed by the nonmarine Boissevain Formation (31.3 m). A coarsening-upward trend follows the start of the Coulter regression. This requires both the continuous creation of accommodation space and an increase in transport energies corresponding to progressive uplift in the proximal sector during a stage of orogenic quiescence (Figs. 10, 12, 13). The shift from the fine sandstones of the Boissevain Formation to organic-rich mudstones at the base of the Turtle Mountain Formation, Goodlands Member (Fig. 13) is taken to mark an abrupt decrease in accommodation space and sediment supply, corresponding to renewed distal uplift in response to a new orogenic pulse and subsidence in the proximal sector (Fig. 12A). The surface separating the fine sandstones of the Boissevain Formation and the organic-rich mudstones at the base of the Goodlands Member is conformable, but has a strong field expression and tectonic significance (end of orogenic quiescence stage). We call this surface a maximum input surface, as it corresponds to an apogee in sediment supply.

Fig. 13. Sediment-dominated (normal) regression within the Coulter–Boissevain third-order sequence.



In this context, the Coulter transgressive phase (Fig. 13) is explained by a subsidence-dominated stage in which the subsidence rate exceeded the sedimentation rate and the Coulter regression is explained by a sediment-dominated stage in which the sedimentation rate exceeded the subsidence rate, manifesting a normal regression. The point of balance between subsidence and sedimentation rates then determines the timing of the Coulter maximum flooding surface (Fig. 13). In this case, the distal maximum flooding surface does not have a tectonic significance, being generated before the end of the orogenic cycle (Figs. 10, 12, 13).

Conclusions

(1) The study of Maastrichtian and Paleocene nonmarine sections reveals that the foreland-basin stratigraphy is characterized by a reciprocal sequence architecture at the level of third-order cyclicity: proximal–distal stages lacking accommodation space (timing of sequence boundaries) correlate to distal–proximal stages of accommodation space creation and sediment accumulation. Further, this indicates that tectonics is the main controlling factor on foreland-basin deposition and stratigraphy.

(2) Four regional third-order hiatuses are recorded: those in the proximal sector are of early late Maastrichtian and late Early to early middle Paleocene age, and those in the distal sector are of early Maastrichtian and late late Maastrichtian to earliest Paleocene age (Figs. 5, 10).

(3) The tectonic control on foreland-basin stratigraphy is explained through a tectonic model (Figs. 10, 12), which takes into account the alternative manifestation of stages of tectonic subsidence and uplift in both proximal and distal sectors of the basin, which are separated by a hingeline. The existence of major proximal disconformities correlative to

distal sedimentary wedges denies the hypothesis that the forebulge would maintain a more elevated position relative to the foredeep at all times.

(4) Nonmarine sequence boundaries may be expressed as subaerial unconformities, incised valleys, top of mature paleosol levels, or base of amalgamated fluvial channels, whereas the position of maximum flooding surfaces may be indicated by extensive coal seams, lacustrine sediments, or other conformable fine deposits (Fig. 12).

(5) Nonmarine sequence boundaries are associated with time gaps corresponding to flexural uplift stages, whereas sequences reflect stages of flexural subsidence. During proximal subsidence (orogenic pulse), the gradual decrease in the dip angle of the topographic slope explains the upwards lowering of transport and depositional energies, which is translated into fining-upward foredeep sequences. During distal subsidence (orogenic quiescence), the steepening of the proximal slope explains the increase in the proximal transport and distal depositional energies, translated into coarsening-upward forebulge sequences topped by a maximum input surface.

(6) As only one sector of the basin has the accommodation space to preserve sediment at any one time, one can never expect to be able to develop a complete biostratigraphic zonation without merging records from both the proximal and distal sectors. Reciprocal vertical tectonics can also be used to explain the distribution of K–T boundary sections in western Canada and northwestern United States. Complete records of the K–T boundary claystone occur in the proximal sector but have yet to be found in the eastern distal sector (Sweet et al. 1999, their Fig. 3).

Acknowledgments

This manuscript represents the culmination of stratigraphic ideas generated during and subsequent to the Canadian Continental Drilling Program Cretaceous–Tertiary (K–T) Boundary Project; hence, the authors especially wish to acknowledge the support of the CCDP Drilling Program Committee, and colleagues and companions on the K–T Boundary Project, Dennis Braman and Jack Lerbekmo. Len Hill's review and advise on the manuscript contributed substantially to its final form. We acknowledge the effort involved in the very detailed review of the manuscript by Ashton Embry and the constructive comments offered by Pierre Binda. We alone take responsibility for the final content of the paper. The congenial and uncomplaining computer drafting support of Odeta Abacioaei and Brenda Davies is gratefully acknowledged. The Geological Survey of Canada, the Royal Tyrrell Museum of Palaeontology, and the University of Alberta supported this project during all phases of its development through to completion. Contributions by Amoco Canada Petroleum Company Ltd. and the Natural Sciences and Engineering Research Council of Canada funded the drilling portions of this project. Additionally, O. Catuneanu wishes to acknowledge financial support from a Natural Sciences and Engineering Research Council of Canada operating grant to Andrew D. Miall, University of Toronto, and Rhodes University of South Africa, his current employer.

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